

https://dspace.org/

The evolution of the Molopo drainage

Bootsman, Cornelis Siebe

2016-08-16
http://hdl.handle.net/10539/20876

Downloaded from a DSpace repository. https://dspace.org/

THE EVOLUTION OF THE MOLOPO DRAINAGE

Cornelis Siebe Bootsman

A thesis submitted to the Faculty of Science, University of the Witwatersrand, in fulfillment of the requirements for the degree of Doctor of Philosophy.

Johannesburg, 1998

ABSTRACT

The appearance of the dry Molopo River with its generally wide and shallow valley cut into the Kalahari Group sediments, but more especially its few impressive rock-cut gorges, has intrigued many people over the ages, and led to many theories concerning its history. The rock-cut gorges, in particular, have traditionally been attributed to ancient, previously much more extended drainage lines, more or less in the same geographical position as the present Molopo Valley. An a. 'ysis of the sediment body of the Kalahari Group, and both, alluvial gravels and geomorphic features on the rim of the Cainozoic Kalahari Basin, have indicated that the Molopo drainage has gradually shifted westward over time, in response to a tilting of the drainage area, which lies across the south-eastern rim of the Kalahari Basin. The present geographical position of the Molopo River is thus a relatively recent one in its evolution. The earliest traces of drainage lines in the area predate the Permo-Carboniferous glaciation of Gondwana. A pre-glaciation valley system with only some similarities to the present-one flowed in a north-westerly direction. There is a long hiatus in the evidence from the end of the glaciation to the beginning of the formation o.' the Cainozoic Kalahari Basin. The most significant feature of that intermediate period, is a large meteorite impact which occurred near Morokweng at the J-K boundary.

The Cainozoic evolution of the Molopo drainage has been strongly influenced by both tectonics and climatic change. Tectonics, which include both the initial formation of the Kalahari Basin and Neogene warpings of the intra-continental axes of uplift, caused the interruption of a pre-Kalahari southward flowing drainage system, an extended upper Molopo, the existence of the Molopo as an endoreic system for an extended period of time, and a westward shift of the entire Molopo drainage system. Progressively more arid conditions interrupted by humid climatic pulses of decreasing intensity have occurred since the Cretaceous. The aridifying conditions caused the existence of playa-like conditions over long periods of time in the back-tilted section of the proto-Molopo. This was followed by a rapid sediment infilling of the sub-basin and a major rejuvenation phase, which caused the incision of the Molopo River into the duricrusted sediments of the Kalahari Group, and the re-establishment of the Molopo River as an exoreic drainage system in its present position. Rock-cut terrace remnants in the upper Molopo give evidence of much smaller climatic changes during the Quaternary. There has been no integrated flow in living memory.

DECLARATION

This thesis is my own, unaided work. It is being submitted for the Degree of Doctor of Philosophy in the University of the Witwatersrand, Johannesburg. It has not been submitted before for any degree or examination in any other University.

.....

11TH day of AUGUST 1998

DEDICATION

Soli Gloria Deo

"...For waters shall burst forth in the wilderness, and streams in the desert. The parched ground shall become a pool, and the thirsty land springs of water..."

Isaiah 35: 6-7 (NKJV)

PREFACE

The Molopo drainage of the southern Kalahari margin has long been one of the great mysteries of southern African geomorphology. The absence of an integrated approach towards the studying of the evolution of this drainage has led to many hypotheses about the duration of its present inactive nature, and its role in the in the drainage of the interior of the subcontinent over time. It is the aim of this thesis to explain the long-term, large-scale evolution of the Molopo drainage, with specific emphasis on the Cainozoic time-frame. In particular, the objectives also include the determination of the influences of tectonics and climatic change on the drainage evolution.

The study is subdivided into three major parts. Part 1, 'Aims and Background', includes the introduction and the physical background to the study. In the Introduction (chapter 1) the general character of the Molopo drainage and its relation to surrounding drainage lines is discussed, and the aims and objectives, approach and techniques used in this study are outlined. This is followed by a discussion of the physical background of the region and the Molopo drainage basin in particular (chapter 2). In part 2, 'Drainage Indicators', the various indicators to the drainage development of the Molopo drainage area are discussed. An outline of indications of very ancient Permo-Carboniferous drainage lines on Gondwana (chapter 3) is followed by a discussion of the sediments within the Cainozoic Molopo (Kalahari) sub-basin (chapter 4), which was formed through the initiation and lowering of the Kalahari Basin in relation with the Griqualand-Transvaal and the Kalahari-Zimbabwe axes of uplift. An understanding of the distribution of the Kalahari Group sediments is deemed crucial for the understanding of the drainage developments of the Molopo drainage. Geomorphological evidence of two areas on the very rim of the Kalahari Basin, successively the upper Molopo drainage area (chapter 5) and the Griqualand West area (chapter 6), is then presented. In part 3, 'Drainage Reconstruction and Conclusions', the major phases that can be identified in the evolution of the Molopo drainage are discussed (chapter 7) and the summary, conclusions, and implications are included in chapter 8.

Some of the material from this thesis has already been published or submitted for publication as:

* Bootsman, C. S., 1997: On the evolution of the upper Molopo drainage, *South African Geographical Journal*, (special issue), 79(2), 83-92.
* Bootsman, C. S. and Reimold, W. U., 1998: The significance of the Morokweng meteorite impact for the long-term Molopo drainage evolution (extended abstract), in: Rowntree, K. M. (ed.), *Proceedings of the Southern African Association of Geomorphologists Biennial Conference, 28th June - 1st July*, Grahamstown, 217-219.
* Bootsman, C. S., (in press): Large-scale, long-term palaeodrainage evolution of the Molopo River (southern Africa), *Proceedings of the Institute of Australian Geographers and New Zealand Geographical Society Second Joint Conference. Hobart Australia, 28th - 31st January 1997*, Department of Geography, University of Waikato.
* Reimold, W. U., Koeberl, C., Brandstätter, F., Kruger, F. J., Armstrong, R. A., Bootsman, C. S., (in press): The Morokweng impact structure, South Africa: geological, petrological, and isotopic results, and implications for the size of the structure, *Proceedings of the Conference on Large Meteorite Impacts and Planetary Evolution (Sudbury 1997)*, GSA Special Paper XXX.

In a study of this nature there are many people who have contributed to its completion. I sincerely acknowledge the support received from the following persons without who this study would not have been possible. My loving wife Phumelele always supported me and showed unlimited understanding for the long hours during weekends and late evenings invested in this study. The love and affection of my precious daughter Maaike have been a continuous stimulant to finish this study. My father, Feitse Bootsman, and mothers, Rie Bootsman-Loots and Maartje Bootsman-Van der Kommer, have given me the foundations to achieve this. My brother Siebe Cornelis Bootsman is greatly acknowledged for having given me moral support during difficult times. My uncle, Evert

Bootsman, gave me fieldwork support during June 1997 in the Griqualand West area. The Dlamini-family in Swaziland is thanked for their support and for understanding my long absence from Swaziland during the writing of this thesis. My *sibari* (Swazi relative), Siphiwe Simelane helped as a family friend and gave unlimited assistance through her work in the Interlibrary Loan section of the library of the University of North West. Pastor Kenneth Makopo of Grace Christian Centre, Mmabatho is thanked for his encouraging prayers and words towards the progress of this study and my new life.

I am indebted to my supervisors Professor T. C. Partridge and Dr. B. P. Moon for their solid guidance and constructive support towards the completion of this study. Professor Dave Thomas and Dr. Mark Bateman of the Sheffield Centre for International Drylands Research of the University of Sheffield, England, Professor Mike Thomas of the University of Stirling, Scotland and Dr. Duncan McGregor, of the Royal Holloway and Bedford New College of the University of London, England contributed by spending precious time with me in the field and discussing the evolution of the Molopo drainage. I wish to thank Professor Nic. Beukes, from the Rand Afrikaans University, who helped to analyse some gravels from the Griqualand West area (chapter 6). I am indebted to Dr. Bob Bourman, Professor of Geomorphology at the University of South Australia in Adelaide, for showing me examples of long-term landscape evolution in the field in southern Australia and for critically perusing an earlier version of chapter 7. James Drummond, a colleague in the Department of Geography at the University of North West, is thanked for stimulating my research endeavours, endlessly discussing the Molopo drainage, and critically reading earlier versions of chapters 5 and 7, and a final version of the whole thesis. Joe Alfers from the Academic Development Centre of the University of North West (now employed at Rhodes University) is thanked for his excellent photography and introducing me to 4x4 driving and Kalahari 'survival'. The Senate Research Committee of the University of North West provided crucial financial support towards this research. Professor Andy Manson of the Senate Research Committee assisted in getting to use the only 4x4 vehicle available on campus for this type of research, in often hopeless situations. Dr. G. L. Humphrey, Dean of the Faculty of

Science and Technology at the University of North West, helped with financial support towards conference funding.

I am also indebted to all the chiefs of the Tswana land as well as the farmers over whose fields I drove and walked, often without being able to secure approval first, during the various stages of this project. Nico Bleeker, mine geologist of the Mamatwan manganese mine near Hotazel is thanked for getting access to and help in understanding the Cainozoic deposits at the mine. I wish to thank Nimal Wijesiri from the Department of Geography for logistical support in the form of the supply of office equipment, and Connie Senna for assistance with some word processing problems and printing, who both are working at the University of North West.

CONTENTS

Page
Abstract ii
Declaration iii
Dedication iv
Preface v
Contents ix
List of figures xiv
List of plates xix
List of tables xx

PART 1 AIMS AND BACKGROUND

CHAPTER 1 INTRODUCTION 1

1.1 The Molopo Drainage 1
1.1.1 The ephemeral character of the Molopo drainage 3
Flood records 6
1.1.2 Implications of this study 10
Pre-Kalahari topography and palaeodrainage 10
Kalahari Group sediments 18
Neotectonics and climatic change 18
1.1.3 The Molopo in context with regional drainage systems 20
The endoreic middle Kalahari 20
The Zambezi drainage system 22
The Limpopo drainage system 23
The Harts/Vaal drainage system 23
The Orange drainage system 24
The Auob/Nossob tributary system 25
The Fish River drainage system 26
1.2 Research Objectives 26
1.2.1 Approach and techniques 27
1.2.2 Data collection and analyses 28

CHAPTER 2 THE PHYSICAL BACKGROUND 30

2.1 The Region 30
2.1.1 Macro-scale southern African geomorphology 32
Larger-scale structural framework 32
Development of the Kalahari Basin 34
The southern African erosion surfaces 40
Gondwana influences 42

Major geomorphological aspects of the African erosion surface in the interior 43
The post-African erosion cycle in the interior 44
Geomorphic aspects of the Pliocene uplift in the interior 45
2.1.2 Regional geology 47
Discrepancies between the Botswana and South African geological maps 47
Basement Complex 49
Late Archaeozoic and Proterozoic rocks 51
Karoo Sequence 53
Kalahari Group 56
2.1.3 Regional structural influences 58
Basement structures 58
Karoo structures 60
Morokweng meteorite impact structure 61
Neotectonics 64
2.1.4 Climate, soils and vegetation of the Molopo Basin 65
Climatic characteristics 65
Soils 67
Vegetation 68
2.2 Topography of the Molopo Drainage Basin 70
2.2.1 Areal extent and description of divides 70
2.2.2 Description of the Molopo valley system, drainage patterns, and thalweg 71
Molopo Valley descriptions 71
Drainage patterns 81
Thalweg profiles 82

PART 2 DRAINAGE INDICATORS

CHAPTER 3 PERMO-CARBONIFEROUS DRAINAGE INDICATORS 86

3.1 Introduction 86
3.2 Dwyka Tillite Formation 87
3.2.1 Dwyka Tillite outcrops 90
3.2.2 Dwyka Tillite suboutcrops 92
3.2.3 Kalahari Basin glacial stratigraphy 93
3.2.4 Depositional model for the Dwyka Formation 94
Glacial stages 96
3.3 Dwyka Palaeogeography 97
3.3.1 Preglacial topography 101
3.3.2 Palaeogeography during maximum glaciation 101
3.3.3 The Hardap Interglacial and Tses Glaciation 105
3.3.4 Postglacial flooding 108

CHAPTER 4 CAINOZOIC MOLOPO KALAHARI SUB-BASIN.....111

4.1 Introduction 111
4.2 Kalahari Group Stratigraphy 111
4.2.1 Mega-Kalahari 112
4.2.2 Molopo drainage stratigraphy 112
Central section 113
Southern section 114
Northern section.....115
4.3 Palaeoenvironmental Reconstructions for the Molopo Drainage Area 117
4.4 A discrete Molopo Sedimentary Sub-basin 119
4.4.1 Kalahari Group isopach pattern 120
4.4.2 Pre-Kalahari surface 120
Weathering characteristics 121
Palaeotopography 121
Pre-Kalahari structural drainage control 124
4.5 Back-tilted Drainage Basin 126
4.5.1 Lacustrine phase 129
4.5.2 Causes of back-tilting 137
4.6 Implications for the Auob/Nossob Drainage Evolution 139

CHAPTER 5 UPPER MOLOPO DRAINAGE AREA 142

5.1 Introduction 142
5.2 Geology and Geomorphology of the Area 142
5.2.1 Review of the diamondiferous gravels as palaeodrainage indicators 147
5.2.2 Review of the role of the intra-continental axes of uplift 150
5.3 Pre-Kalahari Drainage Indicators 152
5.3.1 Proto-drainage lines and the Molopo sub-basin 152
5.4 Extended Upper Molopo Drainage 153
5.4.1 Development of the tectonic 'trough' 153
5.4.2 Reworking and redirection of the diamondiferous gravels 154
5.5 Reduced Upper Molopo Drainage 155
5.5.1 Development of the tectonic 'saddle' 156
5.5.2 Malmani Dolomites as a hydrological barrier 156
5.3 Calcrete tongues of the Slurry Calcrete Field 157
Nature of the calcrete crust 157
Sediment-filled palaeovalley at Ramatlhabama 158
Implications of the nature of the calcrete tongue 161
5.6 Rejuvenation Evidence 161
5.6.1 Flushed-out Molopo River section 163
5.6.2 Hanging valleys 164
5.7 Quaternary Climatic Change 164

5.7.1 Lower terraces 164
5.7.2 Underfit river 165
5.8 Further Palaeoenvironmental Indicators 166
5.8.1 Lobatse Caves 167
5.8.2 Pseudo-karst grikes 167

CHAPTER 6 THE GRIQUALAND WEST AREA 170

6.1 Introduction 170
6.2 Geology and Geomorphology of the Area 171
6.3 Kalahari Group Sediments in the Mamatwan Mine 173
6.3.1 Kalahari Group Stratigraphy in the Mamatwan mine 174
Red-brown Mamatwan clay body 180
6.3.2 Interpretation of the Kalahari Group stratigraphy in the Mamatwan mine 183
6.4 Alluvial Sediments near and across the Griqualand-Transvaal Axis of Uplift 185
6.4.1 Mahura Muthla diamondiferous gravels 185
6.4.2 River-terrace/alluvial gravels between Sishen and Griquatown 190
6.4.3 Wind gaps and gap-crossing gravels in the Asbestos Hills near Danielskuil 194
6.5 Sand River/Orange River Confluence Terrace Gravels 197
6.5.1 Nature of the terrace gravels 197
6.5.2 The Trans-Tswana River 202
6.5.3 Uplift along the Griqualand-Transvaal axis and drainage diversion 204
6.6 Kathu Pan Doline and Wonderwerk Cave 206
6.7 Griqualand West Palaeodrainage Indicators 209

PART 3 DRAINAGE RECONSTRUCTION AND CONCLUSIONS

CHAPTER 7 MOLOPO DRAINAGE PHASES 211

7.1 Introduction 211
7.2 Permo-Carboniferous Drainage Indicators 212
7.2.1 Preglacial landscape 212
7.2.2 Modifications during maximum glaciation and deglaciation 214
7.2.3 A geological hiatus 214
7.3 Mid to Late Cretaceous Drainage Phase 218
7.3.1 Implications of the Kalahari Group isopach pattern 218
7.3.2 The proto-Morokweng River 219
7.3.3 The proto-Molopo River 221
7.4 Mid-Tertiary Drainage Phase 221
7.4.1 Wessels Gravel Formation 223

7.4.2 Playa environments 223
7.4.3 Rapid infilling of the lacustrine environment 225
Mamatwan rejuvenation phase 225
7.5 Plio-Pleistocene Drainage Phase 226
7.5.1 Humid pulse and a re-established exoreic Molopo drainage 226
A joint Auob/Nossob/Kuruman/Molopo system 227
7.6 Quaternary Drainage Conditions 231
7.6.1 Terraces 231
7.6.2 Dens se of the Molopo drainage 232

CHAPTER 8 SUMMARY, CONCLUSIONS, AND IMPLICATIONS 233

8.1 Summary of the Evolution of the Molopo Drainage 233
8.2 Major Conclusions 236
8.2.1 Structural influences and the Molopo drainage evolution 237
8.2.2 The diamondiferous gravels of the Mafikeng-Bakerville-Lichtenburg area 238
8.2.3 The Molopo drainage and the Griqualand-Transvaal and Kalahari-Zimbabwe axes of uplift 238
8.2.4 Climatic fluctuations and terrace formation 239
8.2.5 Duricrust formation and the Molopo drainage evolution 240
8.2.6 The Kalahari Group stratigraphy in the Molopo drainage area 240
8.3 Implications 241
8.3.1 Implications for regional-scale geomorphology 241
8.3.2 Implications for subcontinental-scale geomorphology 241
8.3.3 Further studies in the Molopo drainage basin 241

REFERENCES 243

LIST OF FIGURES

Page

Figure 1.1: Location of the Molopo drainage basin. 2

Figure 1.2: The mega-Kalahari, Kalahari Desert, and Bakalahari *Schwelle*. 4

Figure 1.3: Most frequently reported flooded sections of the Molopo drainage in recent historical records. 7

Figure 1.4: Thickness of Kalahari Group sediments (after Thomas and Shaw, 1991). ..12

Figure 1.5: Pre-Kalahari surface contours (a) and pre-Kalahari drainage lines (b). After Smit (1977). 13

Figure 1.6: Pre-Kalahari drainage in the Molopo drainage according to Levin *et al.* (1985). 15

Figure 1.7: The course of the proposed Trans-Tswana palaeodrainage system plus the present courses of major mekgahca after Dardis *et al.* (1988) and Nash (1992). 16

Figure 1.8: *Mekgacha* networks in the Kalahari (after Thomas and Shaw, 1991). 21

Figure 2.1: Morphological and structural elements of Africa (after Summerfield, 1985a). 31

Figure 2.2: The Southern African Superswell (after Summerfield, 1996). 33

Figure 2.3: Sedimentary basins and deposits in southern Africa at two periods of the Phanerozoic (modified after Dingle *et al.*, 1983). 35

Figure 2.4: Synopsis of basement features in Botswana (after Meixner *et al.*, 1985). ..37

Figure 2.5: Schematic cross section of southern Africa following the splitting of Gondwana (after Thomas and Shaw, 1991). 38

Figure 2.6: Continuous elastic plate model of southern Africa (after ten Brink and Stern, 1992). 39

Figure 2.7: Evolution of the continuous plate model through time (after ten Brink and Stern, 1992). 39

Figure 2.8: Axes of maximum uplift during the Miocene and Pliocene (after Moon and Dardis, 1988). 41

Figure 2.9: Simplified map of the principal cyclic landsurfaces of southern Africa (after Partridge, 1997b). 41

Figure 2.10: Pre-Kalahari geology, emphasising Basement Complex rocks in the Molopo drainage area. 50

Figure 2.11: Pre-Kalahari geology, emphasising Late-Archaeozoic and Proterozoic rocks in the Molopo drainage area. 52

Figure 2.12: Geological sketch map showing extent of basic and ultrabasic rocks correlated with the Bushveld Complex (after Gould *et al.*, 1987). 54

Figure 2.13: Pre-Kalahari geology, emphasising Karoo Sequence rocks in the Molopo drainage area. 55

Figure 2.14: Distribution of Kalahari Beds (>15 m) in the Molopo drainage area. 57

Figure 2.15: Tectonic provinces of southern Africa (after Pretorius, 1985). 59

Figure 2.16: Aeromagnetic field across the Morokweng structure (after Hart *et al.*, 1997). 62

Figure 2.17: Kaapvaal province showing the Morokweng impact structure (after Corner *et al.*, 1997). 62

Figure 2.18: Schematic cross section through a complex impact structure (after Reimold and Gibson, 1996). 63

Figure 2.19: Schematic geological map of the Vredefort dome (after Reimold and Gibson, 1996). 63

Figure 2.20: Earthquake centres in Botswana, September 1965 to August 1971 (after Reeves, 1973). 66

Figure 2.21: Schematic soil map of the Molopo drainage area (based on von M. Harmse, 1978). 69

Figure 2.22: Thalweg profiles of the Molopo, Moselebe, and Kuruman Rivers. 83

Figure 3.1: Isopach map of the glacial sequences in the Karoo and Kalahari Basins and on the Falkland Islands (after Visser, 1987). 88

Figure 3.2: Dwyka Tillite Formation outcrops and suboutcrops in the Molopo drainage area. 89

Figure 3.3: Stages in the growth and disintegration of the Permo-Carboniferous marine ice sheet in the Dwyka Basin (after Visser, 1989). 95

Figure 3.4: Palaeotopography (a) and interpretative palaeoice flow map (b) of the Dwyka Formation around Loeriesfontein (after Visser, 1981). 98

Figure 3.5: Outline of the Kalahari Basin during the maximum glaciation (after Visser, 1987). 99

Figure 3.6: Palaeoice flow directions in the Molopo drainage area. Based on Visser (1983a). 100

Figure 3.7: Schematic pre-glaciation palaeogeography of southern Africa (after Visser, 1987). 102

Figure 3.8: Geomorphic elements of part of south-western Gondwana during maximum glaciation (after Visser, 1987). 102

Figure 3.9: Pre-Kalahari topography and Dwyka Formation suboutcrops in relation with the Morokweng impact structure. 104

Figure 3.10: Major palaeovalleys in southern Africa (after Visser, 1987). 106

Figure 3.11: The ice extent during the Tses Glaciation (Early Permian) at its maximum and at an advanced stage of disintegration (after Visser, 1983a). ... 107

Figure 3.12: Distribution of post-glaciation seas over southern Africa (after Visser, 1987). 109

Figure 4.1: Graphic borehole logs across a pre-Kalahari palaeochannel. Based on data from Gould *et al.* (1989). 116

Figure 4.2: Isopach map of the Kalahari Group in Gordonia (after Botha *et al.*, 1986) 118

Figure 4.3: Isopachs of Kalahari Beds and proto-drainage lines in the Molopo Farms Project area (after Gould *et al.*, 1987). 122

Figure 4.4: East-west geological profile in the area between the Orange and Kuruman Rivers (after Levin, 1980). 125

Figure 4.5: Geological structures of the Molopo Farm Complex area (after Gould *et al.*, 1987). 127

Figure 4.6: Thickness of the Kalahari Beds in the Molopo drainage area. 128

Figure 4.7: The back-tilting principle applied to the proto-Molopo. 130

Figure 4.8: Longitudinal profile of the proto-Molopo after back-tilting and isostatic adjustments. 131

Figure 4. 9: Cross sections of the sediment-filled proto-Molopo (modified after Smit, 1977). 132

Figure 4.10: Location of alluvial cross sections (Fig. 4. 9). Based on Smit (1977). 133

Figure 4.11: Diagrammatic representation of the principle of modelling denudational isostatic rebound for the Murray Basin (a, b, and c) and distribution and (d) magnitude of loads (after Bishop and Brown, 1992). 138

Figure 4.12: Southern Kalahari drainage and thickness of Kalahari Beds. 140

Figure 5.1: Proto-Molopo sedimentary sub-basin and tectonic structures (after Bootsman, 1997). 143

Figure 5.2: The upper Molopo drainage (after Bootsman, 1997). 144

Figure 5.3: Major occurrences of diamondiferous gravels in the upper Molopo drainage area (after Bootsman, 1997). 146

Figure 5.4: Schematic cross section of the Cainozoic deposits of a palaeovalley near Ramathlabama (after Brinn, 1991). 159

Figure 5.5: Calcrete tongues and palaeodrainage lines around Mafikeng. 162

Figure 5.6: Map of central southern Africa showing the Lobatse Caves and the Lebatse Pan, and other sites of earlier palaeoclimatic studies (after Holmgren, 1995). 168

Figure 6.1: Griqualand West study area. 172

Figure 6.2: Simplified face plan of the Mamatwan manganese mine (after Samancor, undated). 177

Figure 6.3: Overburden at the north wall of the Mamatwan mine (after Williams, 1993). 178

Figure 6.4: Mahura Muthla diamondiferous gravels on the Ghaap Plateau. 186

Figure 6.5: River-terrace gravels south-east of the Asbestos Hills. 192

Figure 6.6: River-terrace gravels north-west of the Asbestos Hills. 193

Figure 6.7: Regional geology and location of the Sand River/Orange River confluence (after McCarthy, 1983). 198

Figure 6.8: Composition of gravels on (a) the younger and (b) the older terraces in the Sand River/Orange River confluence area (after McCarthy, 1983). ..201

Figure 6.9: The proto-Orange as suggested by Schwarz (1919). 203

Figure 6.10: Palaeodrainage indicators in the Griqualand West study area. 207

Figure 7.1: Pre-Dwyka palaeodrainage in the Molopo drainage area. 213

Figure 7.2: Schematic geological cross section through the Morokweng impact structure (after Reimold *et al.*, in press). 216

Figure 7.3: Alluvial cross section across a palaeovalley around the Morokweng impact structure near Heuningvlei. Based on Smit (1977). 217

Figure 7.4: Middle to late Cretaceous drainage. 220

Figure 7.5: Late Cretaceous drainage. 222

Figure 7.6: Mid-Tertiary drainage. 224

Figure 7.7: Late Tertiary drainage. 228

Figure 7.8: Synopsis of Molopo drainage shifts. 229

Figure 7.9: Palaeodrainage reconstructions for (a) late Cretaceous and (b) Quaternary. 230

LIST OF PLATES

Page

Plate 2.1: Rock-cut terrace remnant with calcretised sediment along the upper Molopo west of Mafikeng. 72

Plate 2.2: Molopo Silcretes near Logagane. 73

Plate 2.3: Wide, open valley of the Molopo River near Watersend Farm. 74

Plate 2.4: Molopo gorge cut into quartzites near Lover's Leap Farm. 75

Plate 2.5: Entrance of the lower Molopo into the Abiquas Puts Pan. 76

Plate 2.6: Deep rock-cut gorge in the lower Molopo near Riemvasmaak. 77

Plate 2.7: Molopo/Orange confluence area. 78

Plate 3.1: Dwyka Tillite Formation exposed in the floor of a pan in the lower Molopo area. 91

Plate 5.1: Exposed sediment-filled palaeovalley in a quarry near Ramathlabama. ... 160

Plate 6.1: Overview of the Mamatwan manganese mine. 175

Plate 6.2: Red Gordonia Sand Formation exposed in the Mamatwan mine. 176

Plate 6.3: Gullies in the Eden Sandstone Formation at the north wall in the Mamatwan mine. 181

Plate 6.4: A tongue of the Budin Clay formation at the south-west wall in the Mamatwan mine. 182

Plate 6.5: Mahura Muthla diamondiferous gravels. Raisin loaf layer with many well-rounded orange agates (carnelians). 187

Plate 6.6: Mahura Muthla diamondiferous gravels. Subangular gravel of the Kanguru layer present. 188

Plate 6.7: Danielskuil wind gap in the Asbestos Hills. 195

Plate 6.8: Gap-crossing gravels in the Danielskuil wind gap. 196

Plate 6.9: Calcretised gravels of the Sand River terraces. 199

LIST OF TABLES

Page

Table 2.1: Comparative legend for the combined South African and Botswana sections of the geological map of the Molopo drainage. 48

Table 2.2: General thalweg data of the Molopo, Moselebe, and Kuruman Rivers. 82

Table 4.1: Stratigraphy of the Kalahari Group in Gordonia. 114

Table 4.2: Grain size and clay mineral analysis of the red-brov -3udin Clay Formation. 134

Table 4.3: Soil analyses of the Budin Clay Formation samples. 135

Table 6.1: Comparisons of analyses of the Mamatwan clay and the Ongeluk Formation. 183

Table 8.1: Indicators to the evolution of the Molopo drainage. 234

PART 1 AIMS AND BACKGROUND

CHAPTER 1 INTRODUCTION

1.1 The Molopo Drainage

The geomorphic evolution of the Molopo drainage system is poorly understood. It has not been studied as an integrated system from its source to its confluence with the Orange River. The Molopo drainage can be characterised as an extensive, essentially dry, right-bank tributary network of the Orange River with episodic flows mainly in its upstream ends. Only the two major branches of the network, the Molopo and Kuruman Rivers, receive limited quantities of perennial spring flow. Its catchment is situated in the semi-arid to arid interior of southern Africa between latitudes $23^{0}20^{'}$ S and $28^{0}30^{'}$ S, and between longitudes $20^{0}15^{'}$ E and $26^{0}10^{'}$ E. The potentially-exoreic Molopo drainage network is made up of the combined Molopo, Moselebe, and Kuruman Rivers and their tributaries It forms the eastern section of the larger Auob/Nossob/Molopo/Kuruman drainage network (Fig. 1.1), which drains the region of the southern Kalahari south of the Bakalahari *Schwelle* (German translation of a rise). The Bakalahari *Schwelle* is a term that was introduced by Passarge (1904), to indicate the low watershed between the rivers that contribute to the Orange River system in the south, and the northward-flowing endoreic drainage lines of the central Kalahari Desert. The use of the term Kalahari in the literature has been marked by little consistency and precision, as was outlined by Thomas and Shaw (1991). The term Kalahari Basin as used in this study follows their definition, which refers to the Kalahari physiographic basin or 'mega-Kalahari', as a down-warped

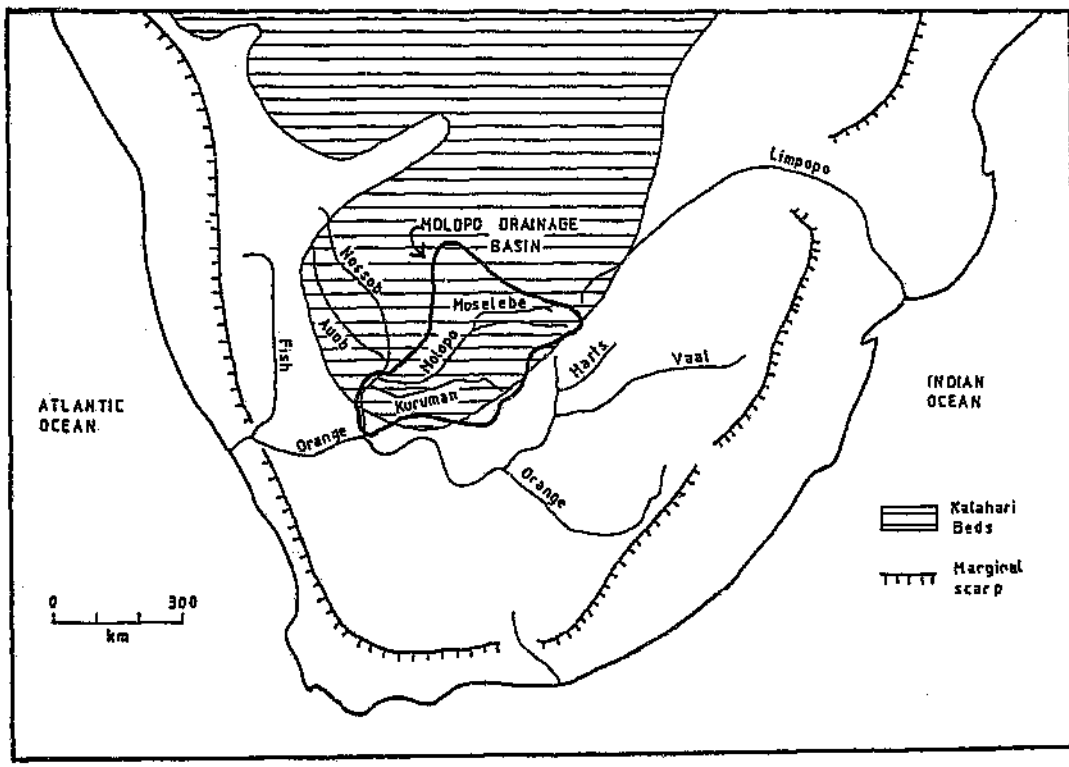


Figure 1.1 Location of the Molopo drainage basin.

series of contiguous sub-basins, into which continental sediments have been deposited since the Jurassic. The use of the name Kalahari Desert in this study is also consistent with their definition, and refers to the area of the Kalahari Basin, south of the Etosha-Okavango-Zambezi swamp zone (Fig. 1.2). It should be noted, however, that the present Kalahari Desert is not a true desert but, in fact, a semi-arid Acacia savanna (Partridge, 1997a).

The Molopo, Moselebe, and the Kuruman Rivers all originate on the western Highveld of South Africa to the east of the sediment-filled Kalahari Basin. Both the Nossob and Molopo Rivers, form over a considerable part of their length the border between the Republics of Botswana and South Africa, and join further south to exit the Kalahari Basin, and enter the Orange River some 17 km downstream from the Augrabies Falls. A part of the lower section of the Molopo River near Abiquas Puts is presently blocked by sand dunes, and it appears that there has been no continuous flow past the sand dunes into the Orange River, over a substantial period of time. Wellington (1955) cited Lewis (1936: 30) who mentioned that there was a stump of a Camelthorn tree (*Acacia eriobola* or *giraffae)* of about a thousand years old, on the sand dune which had diverted the water during the 1894 and 1918 floods into the nearby pan. Wellington concluded that, until the blocking dune is removed by suitable wind and flood conditions, the Molopo/Nossob system is actually endoreic, with its drainage sump at the pan near Abiquas Puts. Wellington (1955,1958) further commented that the lower Molopo, south of Abiquas Puts, was also formerly known as the Hygap River.

1.1.1 The ephemeral character of the Molopo drainage

The episodic character of the Molopo drainage is captured by the meaning of the word Molopo which according to Stigand (1912, as cited in Rogers, 1922) means 'intermittent creek or backwater'. According to Nash (1996) the valleys in the Kalahari region generally do not contain permanent flow, except in their headwater regions, but may contain seasonal pools of water along their courses. Nash (1996) presented historical evidence that describe the condition of the Kuruman and Molopo Rivers over time. The

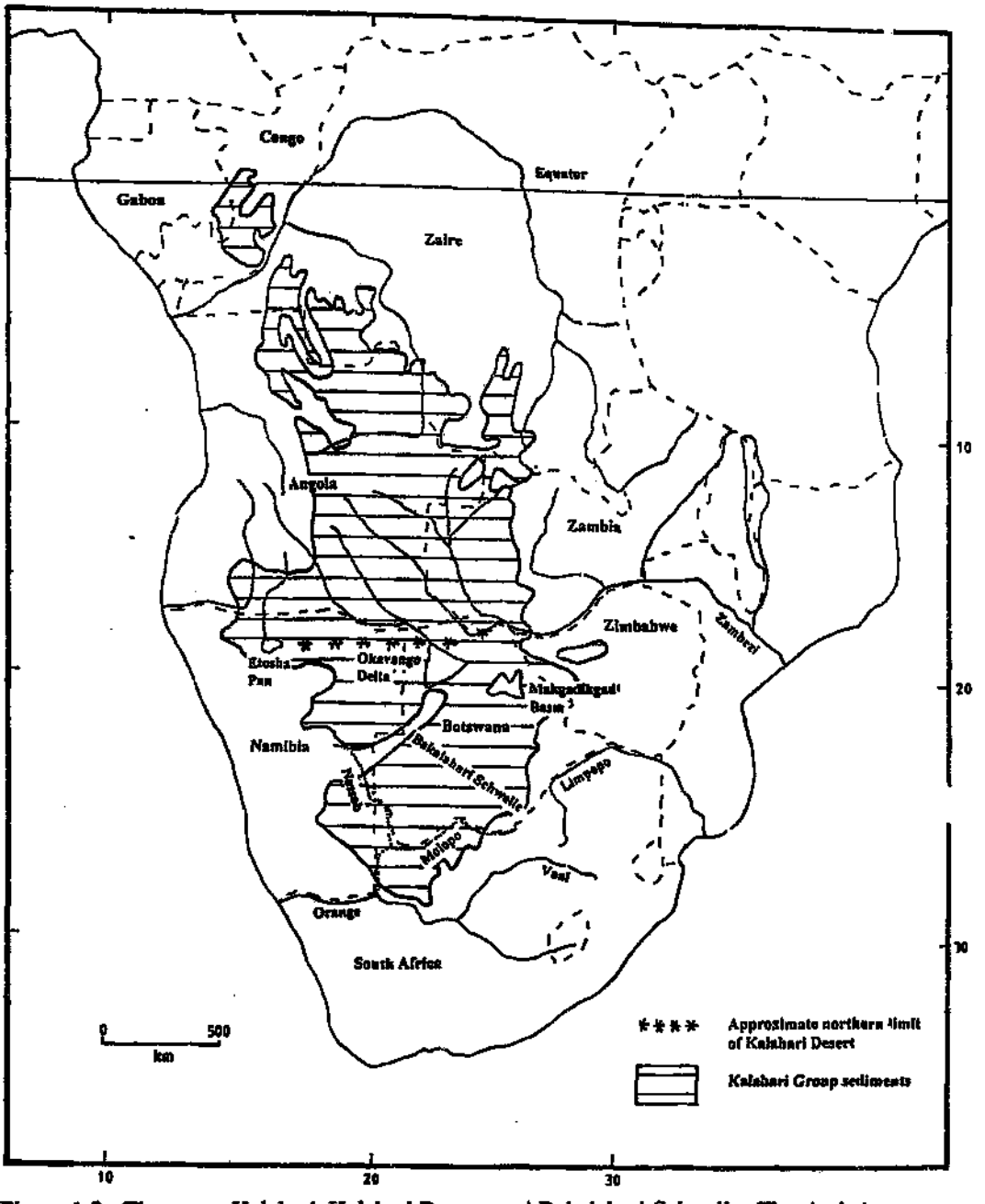


Figure 1.2 The mega-Kalahari, Kalahari Desert, and Bakalahari *Schwelle*. The shaded area is the mega-Kalahari; the Kalahari Desert corresponds to the part of the mega-Kalahari between the Orange River in the south and the Etosha-Okavango-Zambezi line in the north; the term Southern Kalahari refers to the area of the Kalahari Desert south of the Bakalahari *Schwelle*. Modified after Thomas and Shaw (1991).

first description of these rivers dates from 1812 by William Burchell (Burchell, 1824 as cited by Nash, 1996). Andrew Brain (Lister, 1949 as cited by Nash, 1996) described (in 1826) the upper section of the Molopo River, at some distance upstream from Bray (formerly known as Kristonia), as a 'fine running stream' with herds of hippo. This is consistent with the use of the name *Makhubung* ('place of Hippos' in SeTswana, the language of the major population group living in the drainage area), for an area along the Molopo River just outside Mafikeng. Thomas and Shaw (1991) suggested, on the basis of reports of European travellers (e.g. Bain in 1826, Cornwallis Harris in 1836, Gordon Cumming in 1843, and Cotton Oswell in 1845), that the upper Molopo was, during that time, a permanent stream of fair dimensions with thickets of reeds and long grass along it. The reports also suggested that elephant, buffalo, hippo, rhino, waterbuck and reedbuck were still present in the 1850s.

du Toit (1907) initially mentioned that the Molopo flowed only rarely for more than a week at a time during the wet season, and later (1933) described the Molopo as an insignificant river, that normally did not flow much beyond the Setlagole/Molopo River confluence and was incompetent to fill its channel further downstream below Khuis. Wellington (1955) suggested that the current misfit appearance of the Molopo/Nossob/Auob network, could be attributed to the loss of some of its headwaters through capture by the Fish River in Namibia and by the Limpopo system to the east. Wellington later (1958) also mentioned that the much reduced flow of the Molopo, might be attributed to the obvious possibility of climatic change, as well as to capture by the Limpopo. The nature of the Molopo system has been described by Thomas and Shaw (1991), as representing a transitional stage between permanently dry valleys (*mekgacha* in Botswana and *laagtes* in South Africa) and seasonal rivers.

Both the Molopo and Kuruman Rivers, are fed by a series of dolomite springs, called eyes (*oog* in Afrikaans), located outside the Kalahari Basin. The Molopo Eye yields $7.61 \times 10^6 m^3$ per year or 870 m^3 per hour and the Kuruman Eye $6.57 \times 10^6 m^3$ per year (Smit, 1977) or about 750 m^3 per hour. The total yield of the Molopo Eye is readily

consumed by the Mafikeng municipality, and during the past decade no permanent river flow has occurred beyond the eye. Some of the water is, however, re-routed back to the Molopo River from the sewerage facility west of Mafikeng, and is stored in the Modimola and Disaneng dams further downstream. Water from the Modimola dam is, after treatment, pumped back to Mafikeng for renewed usage as drinking water. A substantial part of the yield of the Kuruman Eye is also used as drinking water for the Kuruman municipality. Only the Kuruman River is now a perennial river in its upstream-end, but the water rapidly disappears in the sandy stream bed, only a few kilometres downstream of the dolomite springs (Thomas and Shaw, 1991; Nash, 1996).

Flood records

Large sections of the Kuruman River may contain water for one or two years in response to spring flow, after heavy summer rainfall in the recharge area (Nash, 1996). During above-average rainfall years, floods may also occur in response to short-term high-intensity rainfall events in the upstream ends of the Kuruman and Molopo Rivers and their tributaries, where basement rocks are incompletely covered by Kalahari Sand. A rainfall occurrence of 100 mm (Smit, 1977), within a relatively short (but not specified) period of time, may lead to local floods. The courses of tributaries which are located in thick sand-covered areas (e.g. Khudunkwe, Ganyesa, Phepane, and Kgokgole tributaries), however, have no recorded historical flows (Smit, 1977). Nash (1996) and Smit (1977) each provided a record of dates and extents of floods in the Molopo drainage, that have occurred in living memory. The Molopo River is known to have flooded in 1891-92, 1896, 1915, 1917-18, 1933-34 and 1988, during the period of more recent historical records (Nash, 1996). There are various records of the Molopo flowing as far as Phitshane and near Bray, but there are no records of the Molopo flowing much beyond Watersend Farm. This is also confirmed by Wellington (1958). There are, however, no recorded incidents of flooding of the middle section of the Molopo River from around Watersend Farm downstream, past Khuis, to the Molopo/Nossob confluence (Fig. 1.3). Nash (1996) mentioned several records of flooding of the lower section of the Molopo River, further downstream from the Kuruman/Molopo confluence, which had originated

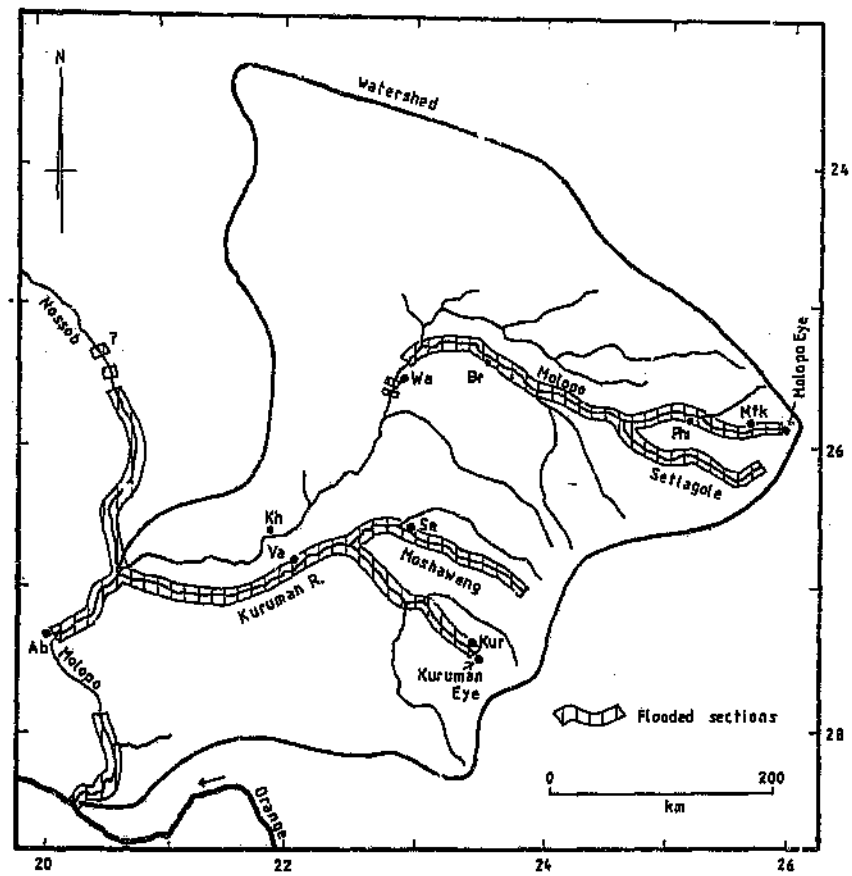


Figure 1.3 Most frequently reported flooded sections of the Molopo drainage in recent historical records. (Ab = Abiquas Put, Br = Bray, Kh = Khuis, Mfk = Mafikeng, Phi = Phitshane, Se = Severn, Va = Vanzijlsrus, Wa = Watersend Farm).

either from the Kuruman River alone or the combined Nossob and Kuruman Rivers. In all instances the floodwater reached not much further than Abiquas Puts. Snyman (1992) and Jacobs (1996) gave unconfirmed reports, based upon a historical letter of the Kuruman missionary Roger Price, of an unbroken river flow in 1894, from the Kuruman Eye down the Kuruman, Molopo and Orange Rivers to the Atlantic Ocean. Wellington (1955: 57) made reference to the same 1894 flood, and stated that the floodwater had come down the Kuruman River, passed Witdraai at the Kuruman/Molopo confluence, and had kept to the Molopo course as far as Witsputs Farm, where it was then diverted by a dune into the pan near Abiquas Puts. Nash (1996: 160) citing Rogers (1907) and Range (1912), who both referred to the same 1894 flood, also stated that the floodwater was unable to follow the partially blocked Molopo River, and formed a large lake at Abiquas Puts.

The fact that Abiquas Puts was the absolute end of the flow of the Molopo River during the floods has been contested by Main (1987). He reported that in 1934 it reached the settlement of Noenieput about 27 km downstream of Abiquas Puts, and that in 1976 it reached Springbok Vlei about 15 km from Abiquas Puts. During both floods the pan at Abiquas Puts developed into a fair size lake. The dimensions of the lake in 1934 were calculated on the basis of aerial photographs, to cover an area of 50 km^2, with depths varying between 4.5 and 6 m (Green, no date). The 1976 floods created a large flooded area, with at least 5 km^2 in which they could safely use a powerboat, and which had depths of 3 to 4 m in the centre (Main, 1987). It took from April to May 1976 to fill the lake, and the water stood for more than two years.

Nash (1996) in an attempt to reconstruct environmental changes on the basis of historical documentary evidence, concluded that fragmentary documentary evidence for hydrological change and fluctuations existed for the Molopo drainage, but cautioned the use of it in terms of climatic or human causative factors. Lowering groundwater tables in the area over the past 150 years have been linked with human interference (Shaw *et al.*,

1992). The negative influence of some farming activities on the physical environment in the Kuruman area, during historical times has also been discussed by Snyman (1992).

A detailed account of recharges and floods exists of the Molopo River and the Moshaweng River, over the period 1960 to 1974 (Smit, 1977). The floodwater of the Molopo River originated from the catchment of the Molopo River in the surroundings of Mafikeng, as well as from the catchment of the Setlagole River in the surroundings of the Setlagole village. In the specified 14 year period the floods along the Molopo River reached the Molopo/Setlagole River confluence (flood distance 160 km from Molopo Eye) about every second year, Bray (flood distance 305 km) five times, the Watersend Farm (flood distance 455 km) three times, and the Arizona Farm (flood distance 515 km) only once (Smit, 1977). In living memory floodwaters have reached the Arizona Farm only twice (1870 and 1967) or about once every 100 years. All floods cited coincided with above-average annual precipitation in the catchment area. Smit (1977) also reported on the flooding of the Moshaweng tributary river during the same period. Flood water reached Severn (flood distance 110 km) five times, and Vanzylsrus (flood distance 210 km) only twice. The Kuruman River is known to have flooded in 1891-92, 1894, 1896, 1915, 1917-18, 1920, 1933-34, 1974-77 and 1988-89, during the period of more recent historical records (Nash, 1996).

It may be concluded that, although there is historic evidence of fluctuating upstream flows and episodic floods within certain sections of the Molopo drainage, there is no confirmed historical documentation of integrated river flow, from either the Molopo or the Kuruman River, reaching the Orange/Molopo confluence. An attempt to determine periods of former perennial flow in the Molopo River on the basis of radiocarbon dating of shell material was made by Heine (1978a, b; 1979a, b; 1981; 1982). He suggested perennial flow in the Molopo River during the period 17 000 to 15 000 BP, and possibly 12 000 BP. The imprecise context of some of Heine's radiocarbon dates, however, have received considerable criticism (Thomas and Shaw, 1991). None the less, the Molopo network contains much circumstantial evidence through its inactive appearance, meander

and sediment sizes, impressive gorges and potholes in areas where the drainage line cuts through solid bedrock, that massive volumes of water have previously flowed through the system.

1.1.2 Implications of this study

This study of the long-term evolution of the Molopo drainage, is part of a recent re-appreciation of long-term landscape studies in geomorphology. It has been stressed that geomorphology, while striving for a physics-based, analytical science methodology, had disregarded long-term studies of landscape evolution, which required the methodology of interpretative, historical science (Sugden, 1996). The need for long-term landscape evolution studies, as well as placing long-term drainage evolution studies in their tectonic framework was also expressed by Nash (1992). The importance of understanding the relationship between large-scale regional tectonics, and the evolution of African drainage has also been indicated by Summerfield (1996). A similar trend is apparent in Australia, where a large volume of publications on long-term landscape evolution of another former section of Gondwana has developed (e.g. Twidale, 1976 and 1994; Gale, 1992; Nott, 1992; Fried and Smith, 1992; Haworth and Ollier, 1992; Bishop, 1995).

The implications of this large-scale, long-term drainage study are that its results will lead to a better understanding of the pre-Kalahari topography and palaeodrainage pattern of the southern Kalahari, the stratigraphy and distribution of the sediments of the south-western Kalahari Basin, and more particularly the role of both neotectonics and climatic change in the evolution of the Molopo drainage.

Pre-Kalahari topography and palaeodrainage

The dry and inactive appearance of the Molopo River has intrigued many people, and led to a variety of interpretations by earlier workers. A brief review of evidence of former flow in the Molopo, as well as postulated ancient links with other valleys and river systems, to the north of the Molopo drainage (e.g. Schwarz, 1920; du Toit, 1927; Rogers, 1936; Wellington, 1955) has been provided by Nash (1992). Some sections of the

Molopo Valley which are more than 30 metres deep, have been suggested to be remnants of a former great river draining the (former) Transvaal. It has been suggested that a 30 m deep gap through banded ironstone rocks of the Kraaipan Group near Phitshane, and the Khuis gorge, where the Molopo has cut a 10 km long, about 30 m deep gorge in quartzites, were due to a Cretaceous drainage line that possibly was connected with the Limpopo drainage (Wellington, 1955). Some other authors (e.g. Bruno, 1985; Rathbone and Gould, 1982) also described sections of a deeply incised pre-Kalahari valley system buried beneath Kalahari Group sediments. Wellington (1929, 1955) suggested that capture of its headwaters by the Vaal River, had to a great extent been responsible for the demise of the Molopo River.

The topography of the pre-Kalahari subsurface has to date, either been described in rather general terms for the entire Kalahari Basin (Fig. 1.4, after Thomas, 1988), or has been discussed with more detail over limited areas, such as within the Molopo drainage area on either side of the border between Botswana and South Africa (e.g. Smit, 1977; Thomas and Thomas, 1989; Gould and Rathbone, 1985; Levin *et al.*, 1985; Gould *et al.*, 1987, 1989).

Smit (1977) interpreted the pre-Kalahari palaeodrainage together with a stratigraphy of the Kalahari Group sediments, on the basis of detailed analyses of several thousand boreholes, in an area which forms now the western part of the North West Province. Smit (1977: 47) unequivocally stated, that the pre-Kalahari palaeodrainage of that area, was in a northerly direction. Although such a direction is indeed suggested by his map of the pre-Kalahari topography, this was not confirmed on his map of the pre-Kalahari drainage (Fig. 1.5 a, b), where he indicated a major southerly drainage line between Bray and Vanzylsrus. He suggested further, that during the Tertiary the palaeodrainage lines became completely filled with clayey and gravelly material, which was transported during periodic floods from nearby higher-lying erosion areas. The old drainage lines thus became blocked to such an extent that the floods changed their courses. A pan line between Bray and Vorstershoop would still indicate the position of the central course of

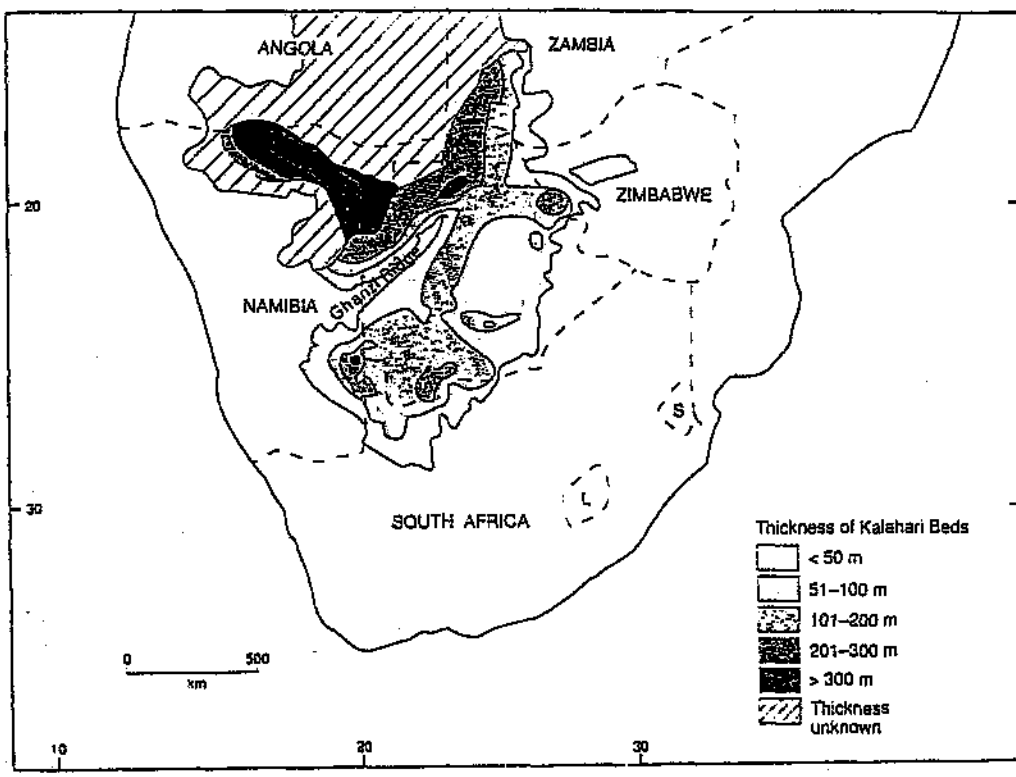


Figure 1.4 The thickness of the Kalahari Group sediments, after Thomas and Shaw (1991).

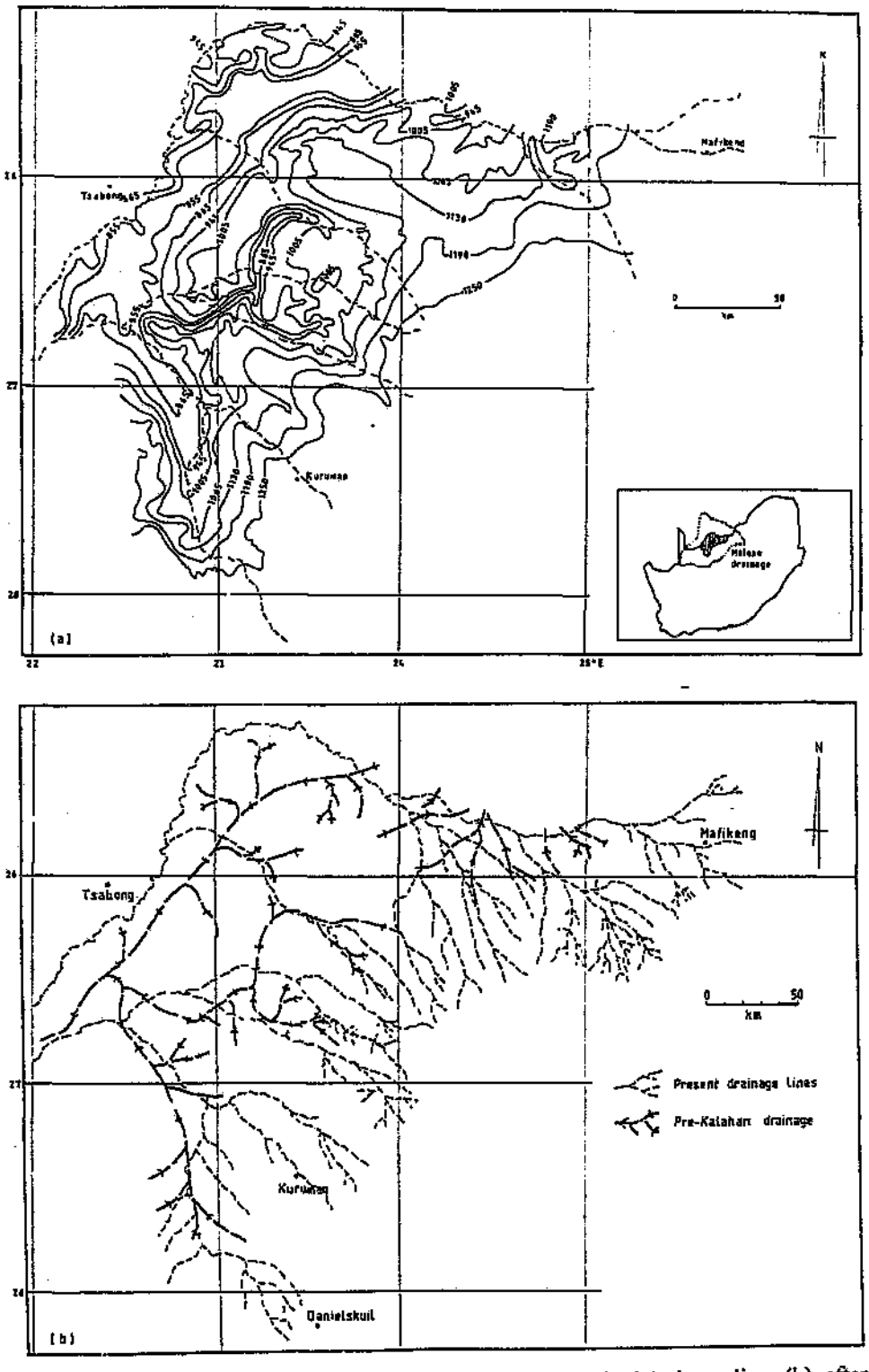


Figure 1.5 Pre-Kalahari surface contours (a) and pre-Kalahari drainage lines (b), after Smit (1977).

the ancient (or proto-) Molopo, as it lies directly above the main sediment-buried palaeodrainage line. He further stated that groundwater flow is still mainly along the palaeodrainage lines.

Smit's palaeodrainage direction interpretation has been cited by a number of authors since then (e. g. Tankard *et al.*, 1982; Dingle *et al.*, 1983; P[illegible]idge and Maud, 1987; Thomas, 1988; Thomas and Shaw, 1991), with the notion that there existed in the southern Kalahari in general, a northward-oriented drainage system prior to, and/or at an early stage of, the initiation of the Kalahari Basin. It has been suggested, that with the deepening of the Kalahari Basin, a centripetal system of northward-flowing channels developed, from an initially southward-flowing system (Partridge and Maud, 1987). Thomas and Shaw (1991) suggested the original post-Gondwana, northward-directed palaeodrainage to have been of an endoreic nature. A northward palaeodrainage direction, on the basis of isopachs of the Kalahari Group sediments, has further been indicated for the northern [illegible] Gordonia (Botha *et al.*, 1986) and for the Noenieput area (Thomas and Thomas, 1989). Southerly-draining buried channels (Fig. 1.6) have, however, also been identified (e.g. Levin *et al.*, 1985; Gould and Rathbone, 1985). Levin *et al.* (1985) suggested a southerly-directed drainage for the Molopo drainage area situated south of the present course of the Molopo in South Africa, which included the area studied by Smit (1977). Gould and Rathbone (1985) and Gould *et al.* (1987) also indicated southerly draining palaeochannels for the Molopo Farms area, north of the present Molopo River in Botswana.

It has been concluded on the basis of analyses of alluvial deposits found near the Sand River/Orange River confluence just downstream of the Vaal/Orange confluence, that the Kalahari was previously drained southwards, by a 'Trans-Tswana' River (Fig. 1.7), across the Griqualand-Transvaal axis, before it became interrupted due to uplift at this axis (McCarthy, 1983). Although McCarthy did not refer directly to the role of the Molopo River in this Trans-Tswana drainage, it can be inferred that a proto-Molopo would have been part of it. Shaw and de Vries (1988) noted that the relationship between the Molopo

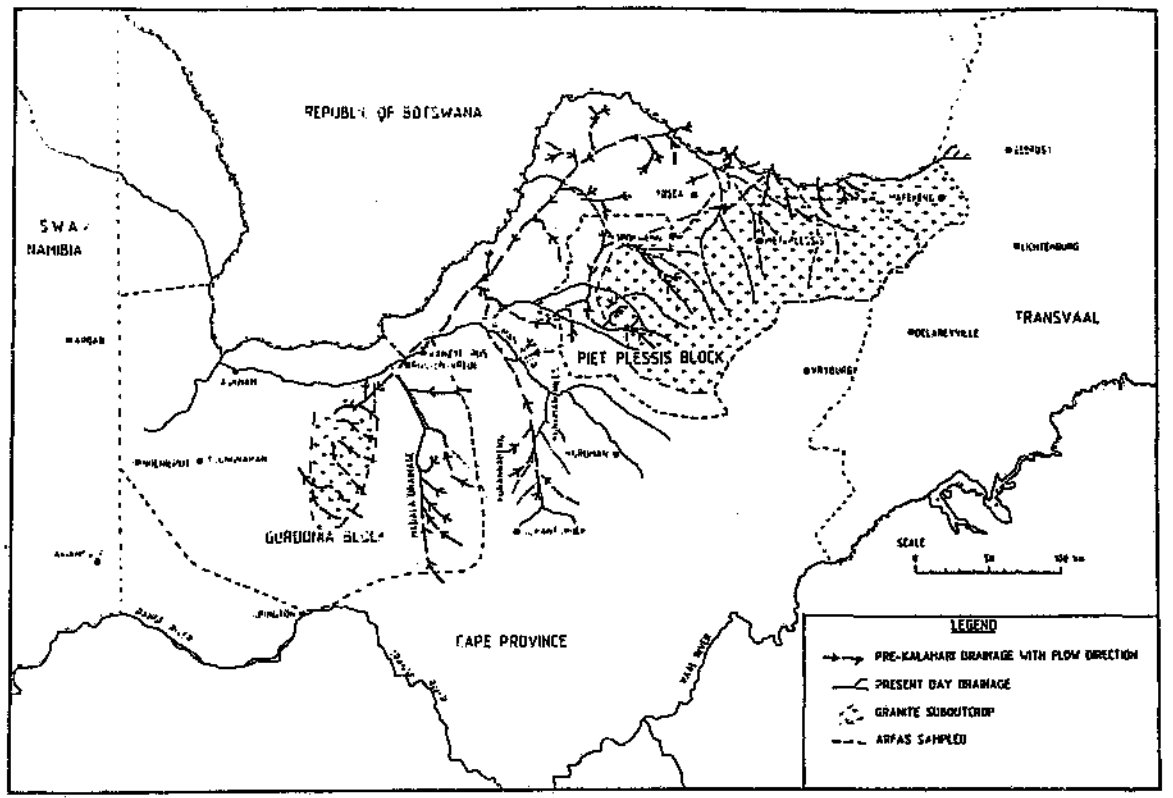


Figure 1.6 Pre-Kalahari drainage in the Molopo drainage according to Levin *et al.* (1985).

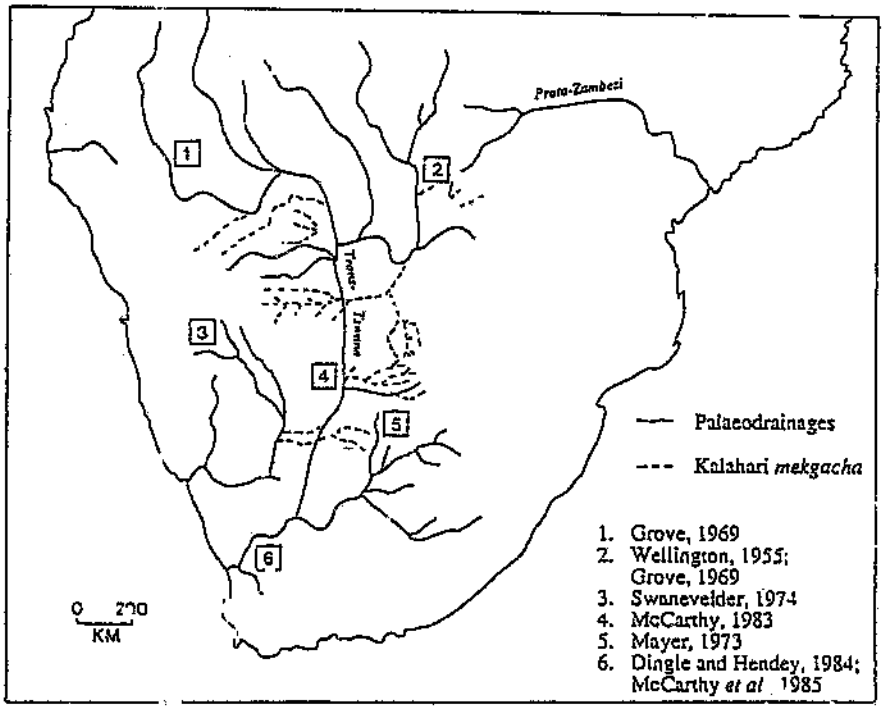


Figure 1.7 The course of the proposed Trans-Tswana palaeodrainage system plus the present courses of major mekgacha, after Dardis *et al.* (1988) and Nash (1992).

drainage network and the pre-Kalahari geology was until recently not fully understood. Nash (1992) cited Smit (1977), Rathbone and Gould (1982) and Bruno (1985), to suggest that the Molopo incised its course along an alignment, that was predetermined by a buried palaeochannel of the Molopo. Nash further suggested that the present Molopo Valley could have developed by headward erosion, since its course cuts across buried north-trending palaeochannels. It has been suggested that a southerly-flowing Cretaceous drainage fed a proto-Molopo situated to the west of the Griqualand-Transvaal axis, to enter the Atlantic via the lower Orange River (de Wit, 1993 citing Behr, 1989). Dingle and Hendey (1984) indicated that since the late Cretaceous, through the Palaeogene and Neogene the Molopo River, exited more or less at the present Molopo/Orange confluence point, into the lower Orange River.

It appears that the study of the Molopo drainage has long been hampered by the presence of the extensive Kalahari sediment cover, the flat terrain, and a lack of natural outcrops (Thomas and Shaw, 1991). Thomas and Shaw also cautioned about the generally uncertain nature and chronology of drainage evolution studies within the Kalahari Basin to date, due to the scarcity of evidence and the fact that former traces of ancient rivers are likely to have been covered by the reworking of the Kalahari Sand. It is evident from the above variety of interpretations of the direction of flow, as well as the actual position of the proto-Molopo, that the dry Molopo drainage system has so far been poorly understood, in terms of its longer term evolution as a river system, especially in relation to the Griqualand-Transvaal axis of uplift.

A number of questions in relation to the development of the Molopo drainage need to be addressed. Can a drainage direction of the proto-Molopo be determined on the basis of the pre-Kalahari topography within the entire Molopo drainage basin? What was the relationship between the proto-Molopo and the proposed Permo-Carboniferous Cargonian Highlands axis by Visser* (1987) and the proposed Cainozoic Griqualand-Transvaal axis by du Toit (1933)? Did the proto-Molopo flow to the west of the Griqualand-Transvaal

* Reference without initials refers to Visser, J. N. J.

axis and was it part of the lower proto-Orange or did it flow across the axis (before it was activated), to be a part of the upper proto-Orange? This thesis provides a model of the drainage evolution of the Molopo, which attempts to answer such questions.

Kalahari Group sediments

The poorly understood stratigraphy of the sediments of the Kalahari Group is a consequence of the lack of natural outcrops, and limitations of available borehole data (Thomas, 1988; Thomas and Shaw, 1990 and 1991). Some authors (e.g. Cahen and Lepersonne, 1952 and Wright, 1978, as cited by Thomas, 1988 and Thomas and Shaw, 1991) suggested, that the Kalahari Basin is actually a number of sub-basins. The role of the Ghanzi Ridge in subdividing the Kalahari Desert, as well as to the existence of the tectonically controlled Okavango Rift and the Makgadikgadi and Etosha depressions was pointed out by Thomas and Shaw (1991). Three sub-basin structures south of the Ghanzi Ridge have been identified (Thomas, 1988). The largest is situated in the south-east, across the Namibian/South African/Botswana borders, another in the south-west, within the Molopo drainage area, and a smaller, shallower basin centred on about 24^0 S 25^0 E. Regional differences in the stratigraphy of the sediments of the Kalahari Group could be related to specific regional factors, which influenced more or less discrete sub-basins. It is suggested here that a further consideration of the Molopo drainage sub-basin and its particular regional influences, will lead to a better understanding of the stratigraphy as well as the distribution pattern of the Kalahari Group sediments.

Neotectonics and climatic change

A further analysis of the role of neotectonics in the down-warped interior of southern Africa, can lead to a better understanding of drainage developments at the southern rim of the Kalahari Basin, as has been demonstrated by Thomas and Shaw (1988, 1992, and Shaw and Thomas (1988), in the case of the drainage evolution of the Zambezi River. The influence of both the Griqualand-Transvaal and the Kalahari-Zimbabwe axes of uplift, as first identified by du Toit (1933) in determining the drainage evolution in the interior of southern Africa, seem to be fundamental factors in determining the drainage

developments of the Molopo drainage. Besides neotectonics, the possible rejuvenation of older geological structures will also be considered.

Within the Molopo drainage area there is further ample evidence of climatic change during the Cainozoic. Analyses of sediments and landforms can provide indicators of climatic fluctuations as well as long-term climatic trends. Partridge and Maud (1987) for example mentioned the incision of the Molopo River into the Kalahari Group sediments, together with its re-establishment as an exoreic drainage system, as an indication of the magnitude of the Pliocene humid pulse. Terraces have also been mentioned as indicators of fluctuating climatic conditions during the Pleistocene (Heine, 1982). The possible larger-scale climatic implications, which can be drawn from the development of the Molopo drainage will be examined.

Finally it should be noted, that research on the Molopo drainage system until now has not centred on the evolution of the integrated network, but rather has dealt with particular aspects along limited stretches only (e.g. du Toit, 1906 - geology of Mafikeng region, 1933 - interior axes of uplift, 1951 - diamondiferous gravels Lichtenburg area; Wellington, 1929 - Vaal/Limpopo watershed; Goudie, 1973 - calcretes in the Molopo drainage; Mayer, 1973 - morphotectonic development Molopo/Harts; Smit 1977- geohydrology upper and middle Molopo; Strauen, 1979 - diamondiferous gravels south-western Transvaal; Levin, 1980 - uranium potential Kuruman/Orange River area; de Wit, 1981 - diamondiferous gravels near Lichtenburg, 1993 - Northern Cape drainage evolution; Heine, 1978a, 1978b, 1979a, 1979b, 1980, 1982, 1990 - late Quaternary environmental conditions in the southern Kalahari; Botha *et al.*, 1986 - geology of Gordonia area; Thomas and Thomas, 1989 - geology of Noenieput area; Marshall, 1990 - diamondiferous gravels south-western Transvaal; Brinn, 1991 - Cainozoic deposits upper Molopo; Nash, 1992 - dry valleys of the Kalahari, 1996 - documemary evidence of climatic conditions; Holmgren, 1995 - late Pleistocene environmental change from Lobatse Caves). It is suggested therefore, that a more holistic approach to the study of the entire drainage system, with the help of a new interpretation of data from an ever

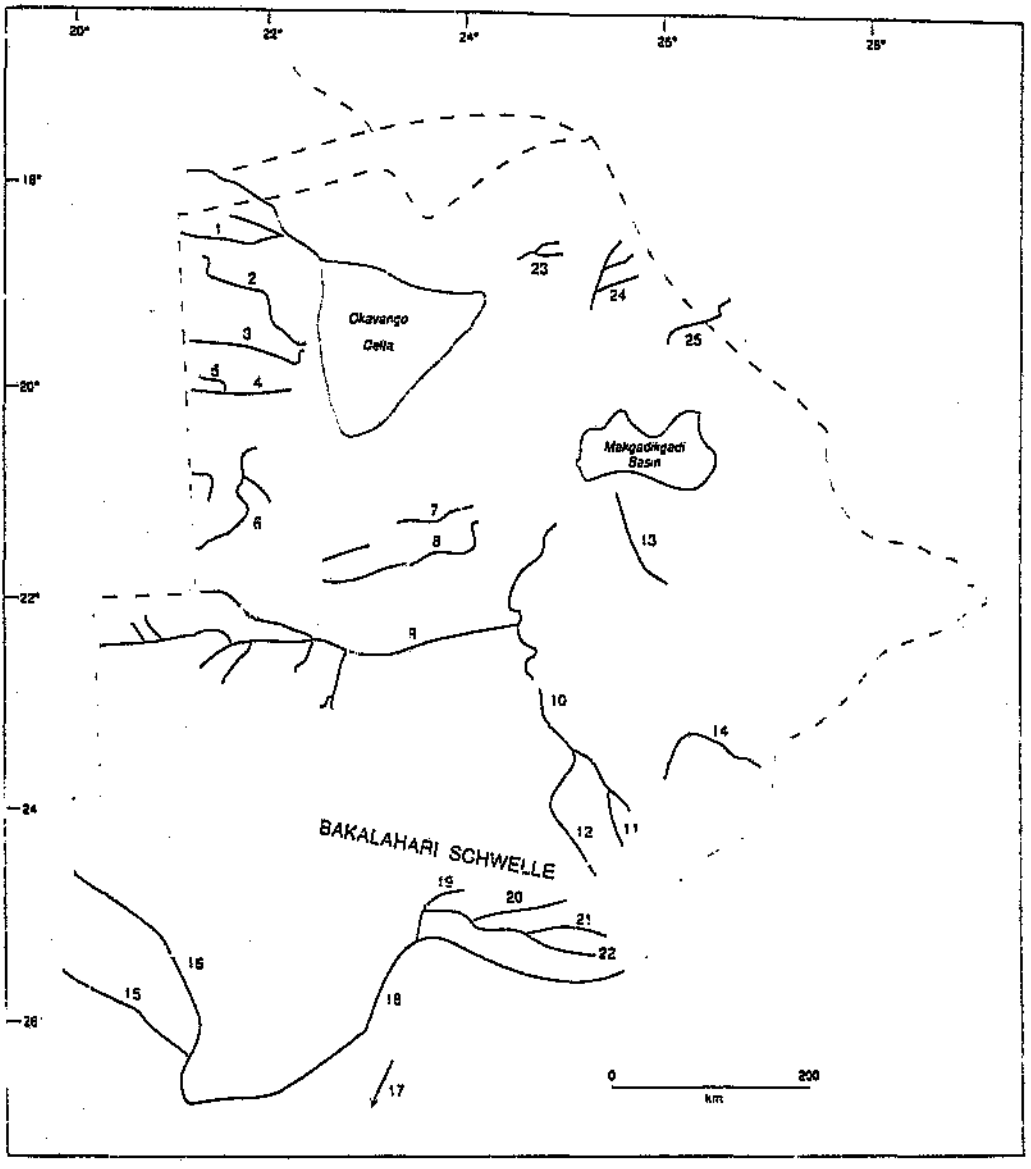
increasing number of boreholes, as well as from large exposed sections of Kalahari Group sediments in some large open-cast mines at the southern edge of the Kalahari, can now provide a new insight into the Molopo drainage development.

1.1.3 The Molopo in context with regional drainage systems

The Molopo drainage is surrounded by a number of drainage networks, that vary considerably in nature. Immediately to the north of the Molopo drainage network, north of the Bakalahari *Schwelle*, are the dry valleys or *mekgacha* of the endoreic middle Kalahari, the Makgadikgadi Basin and the Okavango Delta. This area is bounded further to the north by the perennial Zambezi River. Towards the north-east of the Molopo drainage lies the Limpopo, to the east the Harts/Vaal system, to the south the Orange River, and to the west the Auob/Nossob tributary system and the Fish River system. Except for the Fish River and the Auob/Nossob tributaries, which are ephemeral systems, these are mainly perennial systems with distinct seasonal fluctuations in flow. The influence of neotectonics on the development of all the surrounding drainage networks, has been a common point of discussion by various authors. These networks and some of their major developments will be discussed briefly, in context with the Molopo drainage.

The endoreic middle Kalahari

The *mekgacha* of the endoreic middle Kalahari (Fig. 1.8) are all directed towards the Makgadikgadi Basin and the Okavango Delta (Thomas and Shaw, 1991). By far the largest *mekgacha* network is the Okwa/Mmone system and Shaw *et al.* (1992) indicated that the Okwa/Mmone system flowed between 15 000 and 12 000 BP into palaeolake Makgadikgadi. Recent studies of these dry valley systems have provided new insights into their development. Evidence of deep weathering and groundwater erosion along pre-Kalahari lineaments in the Lethlakeng valley system, which forms part of the larger Okwa/Mmone system was reported by Shaw and de Vries (1988). They further reported on the complex nature of the duricrusts in the valley system, the role of uplift along the Zimbabwe-Kalahari axis on the valley development, and similarities with the Molopo network. The role of groundwater sapping, deep-weathering and structural control, as



MEKGACHA NETWORKS

1	NGAMASERE	10	MMONE	18	MOLOPO
2	XAUDUM	11	LETLHAKENG	19	MABUASEHUBE
3	QANGWADUM	12	NALEDI	20	UKWI
4	EISEB	13	LETLHAKANE	21	MOSELESE
5	GOMHABEDUM	14	SERORome	22	SEKHUTANE
6	GROOT LAAGTE	15	AUOB	23	GHAUTUMBI
7	PASSARGE	16	NOSSOP	24	NUNGA
8	DECEPTION	17	KURUMAN	25	LEMEMBA
9	OKWA				

Figure 1.8 Mekgacha networks in the Kalahari (after Thomas and Shaw, 1991).

important factors in dry valley development, plus the recognition that tectonic changes and drainage capture need to be considered, have further been emphasised by Nash, Thomas and Shaw (1994), Nash, Shaw and Thomas (1994), and Nash (1995).

The Zambezi drainage system

The headwaters of the Zambezi River originate on the Angolan Highlands. Further downstream the river crosses the Kalahari Basin, and then continues its easterly course to exit into the Indian Ocean. Since, both, the Zambezi and Molopo drainage systems can be traced over considerable distances through the Kalahari Basin, and both exit the Kalahari Basin across its rim, it seems relevant to consider the current views on the drainage evolution of the Zambezi River. It is generally believed that the present course of the Zambezi River is relatively recent, and that the upper Zambezi and middle Zambezi evolved as separate systems until the Pliocene to mid-Pleistocene.

Thomas and Shaw (1991) cited various authors who believed that the upper Zambezi once continued its southerly course to join either the Orange River (e.g. Lister, 1979), or what Thomas and Shaw considered more probable, the Limpopo. They also postulated that the upper proto-Zambezi existed some time as an endoreic river, which deposited its sediment in the Kalahari Basin, before connecting with the Limpopo. They further suggested that this southward course of the upper proto-Zambezi was disrupted by gentle movements of the Kalahari-Zimbabwe axis near the Makgadikgadi Basin, and down warping near the Okavango region. Renewed endoreism of the upper Zambezi, which may have provided a source for a pre-Pleistocene Lake Palaeo-Makgadikgadi, may thus have existed again, prior to a final connection between the upper and middle Zambezi. In this process a former left-bank tributary of the upper proto-Zambezi might have undergone drainage reversal. It was believed that this connection between the upper Zambezi and the middle Zambezi occurred through overtopping of the Lake Palaeo-Makgadikgadi in a cataclysmic flood (Nugent, 1990). The evidence for a catastrophic flood has subsequently been doubted by Thomas and Shaw (1992).

The Limpopo drainage system

The Limpopo drainage system originates on the eastern edge of the Kalahari Basin, north-east of the Bakalahari *Schwelle*. Its catchment covers eastern Botswana, most of the Northern Province of South Africa, southern Zimbabwe and enters the Mozambique Channel via the coastal plain of Mozambique. Neotectonics also seem to have played a major role in the development of the Limpopo drainage. The rise along the Kalahari-Zimbabwe axis, which Cooke (1980) suggested to have taken place since the late Tertiary, led to the isolation of the Makgadikgadi Basin, and the beheading of the upper Molopo (Wellington, 1929), and caused the strong development of the left bank tributaries of the Limpopo network (Shaw, 1989). An anticipated 75-100 km southward shift of the watershed between the Limpopo and the Orange/Vaal system since late Cretaceous was mentioned by Partridge and Maud (1987). Presently the watershed between the Molopo and the Limpopo lies mainly across the Malmani Dolomite, which lacks integrated surface drainage. It has been speculated that there had been a Cretaceous drainage line joining the Limpopo with the Molopo at its great south-westward bend (Wellington, 1955).

The Harts/Vaal drainage system

The Harts River system is a right-bank tributary network of the Vaal River and lies immediately to the east of the Molopo drainage. The Harts/Vaal drainage system is a major right-bank tributary network of the Orange River, which drains most of the eastern part of the Western Province, the Free State, and the southern part of Mpumalanga Province. Interactions between the Molopo and Harts Rivers have been discussed by various authors (e.g. du Toit, 1933, 1951; Wellington, 1929; Mayer, 1973; Stratten, 1979; de Wit 1981; Marshall, 1990). These interactions invariably dealt with diamondiferous gravel ridges of the Bakerville-Lichtenburg area and river piracy. The Dry Harts River follows a pre-Dwyka drainage line, as was suggested by du Toit (1910). du Toit (1933) was the first to discuss the influence of the interior Griqualand-Transvaal axis on the divide between the Molopo and Harts networks. A slightly different position of the axis on the stretch between the Harts and Vaal Rivers was suggested by Mayer (1973).

Partridge and Maud (1987) indicated values of vertical warping and a time frame for warping along the axis, which were further refined by Marshall (1990) for the Western Transvaal section. Actual warping of the Griqualand-Transvaal axis has been doubted by Wellington (1958) and de Wit (1993), and they described the 'so-called Griqualand-Transvaal axis' as a geomorphological barrier, formed by the exhumation of the Cargonian Highlands from its Karoo cover, rather than as being an axis of uplift.

The Orange drainage system

The Orange River catchment covers most of the interior of southern Africa, has a mean annual discharge of 1061 x 10^7 m^3 , and is with 953 200 km^2 in area one of the largest drainage basins in the world (Dingle and Hendey, 1984). Dardis *et al.* (1988) and de Wit (1993) summarised the major course changes that have occurred during the river's history. There is evidence of a separate lower proto-Orange River and upper proto-Orange/Vaal system for a considerable period of time. Dingle and Hendey (1984) indicated that the upper proto-Orange system entered the south Atlantic through the Cape Canyon, 300 km south of its present mouth. The lower proto-Orange River drained a much smaller area, and entered the south Atlantic at the mouth of the present Orange River. de Wit (1993) named the lower proto-Orange River the 'Kalahari River', and the upper proto-Orange River the 'Karoo River', which were separated by the Griqualand-Transvaal axis. de Wit further cited Behr (1989), as reporting that a Cretaceous drainage from the Kalahari fed the proto-Molopo, to enter the Atlantic via the lower proto-Orange River. A progressive decrease in annual sediment yield for the proto-Orange River from 10 x 10^6 m^3 in the late Cretaceous, to 0.3 x 10^6 m^3 in Neogene times has been reported by Dingle and Hendey (1984). Such decreasing sediment yields had been interpreted as reflecting increasing desert conditions (Milliman and Mead, 1983 as cited by Dardis *et al.*, 1988), but could also be explained by the abandonment of the Trans-Tswana River (McCarthy, 1983), as suggested by Dardis *et al.* (1988). McCarthy (1983) estimated that the Trans-Tswana could have had a discharge of as much as four times the discharge of the Orange River, and would have drained much of south-central Africa

Because of its anticipated geographical position, the course of this Trans-Tswana would have coincided with a proto-Molopo system. Crustal warping of the Griqualand-Transvaal axis, which extended across the proposed course of the Trans-Tswana River, would have caused internal drainage and associated filling of the Kalahari Basin. The Okavango Delta, Lake Ngami, and the Makgadikgadi Pans were suggested to be the final remnants of this system (McCarthy, 1983). The major changes in the drainage evolution of the Orange River, which ultimately led to the connection between the upper and lower proto-Orange River systems, were suggested by de Wit (1993) to be due to humid pulses in a gradual aridifying climate, rather than through crustal warping. It appears that the early Molopo drainage played an important role in the drainage evolution of the Orange River, either as a tributary of the 'Kalahari' or lower proto-Orange River (de Wit, 1993), or as a much-extended upper proto-Orange drainage (McCarthy, 1983).

The Auob/Nossob tributary system

The Auob and Nossob Rivers originate in the highlands to the east of Windhoek in Namibia, and form the western section of the larger Auob/Nossob/Molopo/Kuruman drainage network. The Khomas axis of uplift (du Toit, 1933) separates the southward draining Auob/Nossob system, from the Okavango-directed flow to the north of it. The Auob and Nossob can be considered perennial in their upper sections, with a flash flood regime due to seasonal rainfall in their mountainous catchments (Nash, 1996). Wellington (1955) mentioned that floodwaters from the highlands during the rainy seasons stop abruptly at the foot of the highlands, as the water reaches the plain, and the river courses become a series of pans. He further mentioned that floodwaters rarely reached the lower courses, and that between the Auob/Nossob and Nossob/Molopo confluences a sand dune with tree growth on it extends across the bed of the Nossob. The middle sections of the Auob, Nossob and their major tributary, the Olifants River, have a characteristic parallel to sub-parallel drainage pattern, which also occurs along certain sections of the Molopo drainage. The lower course of the Auob and Nossob has been described as having partially dune-obscured, broad flat beds, typical of rivers influenced by flash floods in arid environments (Thomas and Shaw, 1991).

The Fish River drainage system

The Fish River network, downstream of the Molopo drainage, is the next major right-bank tributary system of the lower Orange River. It is situated near the western rim of the Southern Kalahari and originates along the Great Escarpment south of Windhoek, to join the Orange River near the Richtersveld, where the Orange River breaks through the Great Escarpment. Wellington (1955) mentioned that practically no water from the highlands finds its way into the lower Fish River, as most of it gets already absorbed by the Kalahari Sands in the Rehoboth Depression. Where the lower Fish River has cut itself deeply through the Huns Plateau it, forms the well-known Fish River Canyon.

Wellington (1958) discussed some outstanding problems with the Fish River Canyon. He stated that in the canyon section the Fish River has eroded its course in a structural trough. Wellington suggested that subsequent to trough-faulting, the Fish River must have flowed northward, and that the present pattern is due to headward erosion by a steeply graded tributary of the Orange River. This tributary would have cut back in the trough, and reversed the northward flow of the trough drainage. Some support for this was found by Wellington (1958) further north of the canyon, near Tses, where the Fish River changes sharply from an easterly course to a southerly direction. He further mentioned directly to the east of this bend, the presence of a sandy shallow valley stretching south-eastwards towards the pans of 'Mier Country' and further to the Molopo/Nossob confluence. He concluded that the section upstream of the Fish River bend at Tses was the headstream of a former south-eastwards flowing river, and that headward erosion of a south-flowing tributary of the Orange River had reversed the canyon drainage downstream of the bend near Tses.

1.2 Research Objectives

The general aim of this study is to determine how the Molopo drainage has evolved since the Mesozoic, and how this fits in the context of the general drainage development in southern Africa.

More specifically the objectives are:

(i) to further the understanding of the pre-Kalahari surface and the influence of tectonics and geology on the developing drainage network,

(ii) to clarify the relationship between the diamondiferous gravels of the Mafikeng-Bakerville-Lichtenburg area and the evolution of the Molopo drainage,

(iii) to acquire a better understanding of the relationship between the Molopo drainage and the Griqualand-Transvaal and Kalahari-Zimbabwe axes of uplift,

(iv) to explain the relationship between climatic fluctuations and terrace formation,

(v) to determine the conditions of duricrust formation and the interaction between duricrusts and the Molopo drainage evolution, and,

(vi) to acquire a better comprehension of the stratigraphy of the sediments of the Kalahari Group in the drainage area.

The research proposal is therefore based on the following hypotheses:

(i) The evolution of the Molopo drainage is the result of an interplay between tectonics and climate.

(ii) The proto-Molopo River shifted west-north-west and incised into cemented sediments of the Kalahari Group due to the warping of the Griqualand-Transvaal axis of uplift.

(iii) The Mafikeng diamondiferous gravels belong to an episode in the evolution of the Molopo drainage when its headwaters were extended in a north-east direction prior to both warping along the Griqualand-Transvaal axis of uplift and subsidence of the Bushveld Basin.

(iv) Cementation of some of the deposits of the Kalahari Group is due to climatic fluctuations under variable conditions of cementation.

1.2.1 Approach and techniques

The general approach towards this study is to make use of interpretative historical science alongside fieldwork, to produce a model that provides an explanation for the long-term, large-scale evolution of the Molopo drainage. Interpretative reasoning and narrative logic

(Frodeman, 1995) together with a holistic approach will be used to analyse data from various sources concerning the integrated Molopo drainage. As the research progressed, the emphasis of the research gradually shifted from a regionally more detailed study of surface features in the Molopo region (e.g. diamondiferous gravels, terraces) alone, towards the development of a larger scale model of the Molopo drainage evolution, strongly influenced by the analysis and interpretation of the spatial aspects of the Kalahari Group sediment body identified within the Molopo sub-basin.

1.2.2 Data collection and analyses

Data were compiled from documentary sources, such as a re-evaluation of already published material by other authors, from available satellite imagery of the upper Molopo, aerial photographs of the Mafikeng area, topographical maps, geological maps, hydrogeological maps, and landtype maps of the drainage. In addition, information has been gathered through field research by the way of identification of surface features, sediments, duricrusts and artefacts, as well as through sampling of sediments and duricrusts of the Kalahari Group. A global positioning system (GPS: Trimble, Pathfinder-basic) was used to determine accurately locations in the fieldwork area. Although the entire area was visited and photographed during various field excursions, more extensive field research occurred in two areas on the easterly rim of the Kalahari Basin. Firstly in the upper Molopo drainage area between Makgobistad/Phitshane in the west to Swartruggens in the east. This area was chosen as a key area where many surface features and sediments can readily be observed, and the effects of the warpings of, both, the Griqualand-Transvaal and the Kalahari Zimbabwe axes, as well as the subsidence of the Bushveld Basin have interacted. In the area downstream from Makgobistad/Phitshane many older river traces are covered by Kalahari Sands. A second key area that was more extensively visited, was the Griqualand West area, to gather surface evidence of an anticipated crossing of the proto-Molopo palaeodrainage and the axis of the Griqualand-Transvaal upwarp. Data analysis techniques used included interpretation of satellite

imagery and aerial photographs, and comparisons of chemical, sedimentological and chronological analyses.

* * * * *

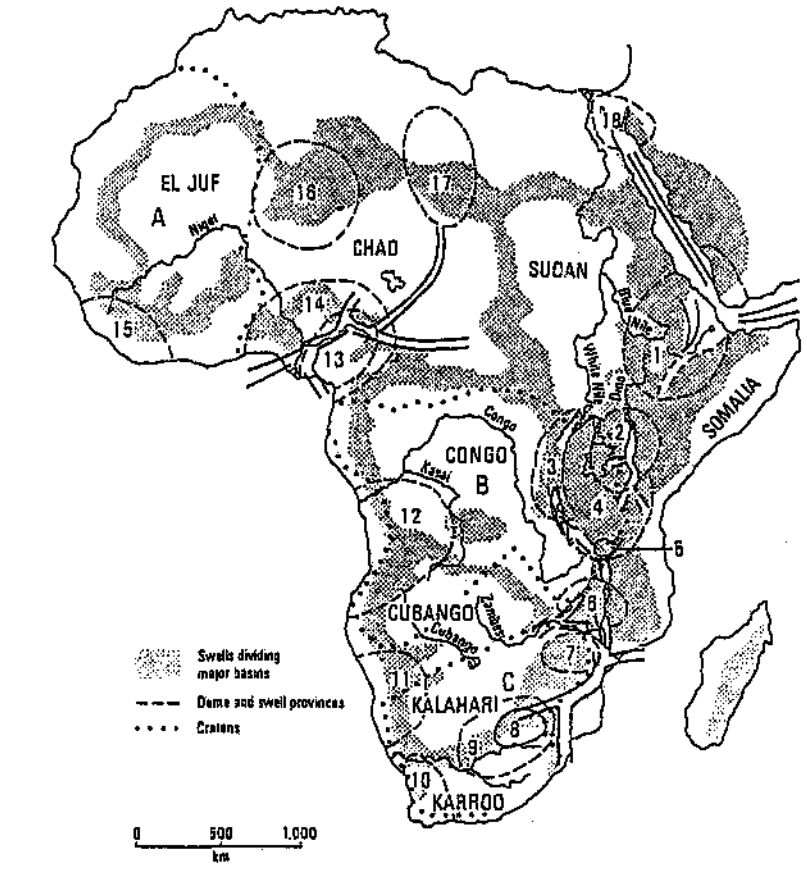
This introductory chapter has described the ephemeral character of the Molopo drainage and its geomorphological relations with neighbouring drainage systems. All surrounding drainage systems display a strong common influence of neotectonics on their development. Furthermore, interactions between the Molopo and Limpopo, Harts, upper and lower Orange, and Fish River systems have been suggested by various earlier authors during diverse stages in the evolution of the Molopo drainage. Finally the nature, methods and techniques of the study were outlined. This study of the Molopo drainage development has predominantly made use of the interpretative historically scientific method to determine its long-term, large-scale evolution. In the next chapter the physical background for the investigation of the Molopo drainage is discussed.

"Comparison can more properly be made with a warped gramophone record unaffected at its centre but crinkled along its margins. These crinkles are, however, not radially disposed, but oriented roughly in a north-east south-west direction" (du Toit, 1933: 16).

CHAPTER 2 THE PHYSICAL BACKGROUND

2.1 The Region

Southern Africa forms part of the largest plateau of the Earth's surface, the great African plateau, which stretches from South Africa to the southern limits of the Sahara, and formed the largest segment of Gondwana (Wellington, 1955). Wellington further stated, that with the exception of the middle Zambezi, the middle Limpopo, and the lower Orange tectonic and erosional troughs, the lowest parts of the plateau surface are the lower Molopo and Makgadikgadi Basins, with altitudes of just under 900 m. The high mean elevation of the African continent, its distinct basin-and-swell morphology (Fig. 2.1), the depositional and erosional surfaces typical of low local relief, and the rift valley systems were emphasised by Summerfield (1985a) as some of the major physical characteristics typical of the African continent. The African depositional and erosional surfaces are extensively covered by duricrusts and in some places separated by spectacular escarpments (Summerfield, 1985a). The mega-Kalahari is an example of an area dominated by a depositional surface which has extensive duricrusts. The African Superswell as the first-order morphological feature in southern Africa, with large areas of this superswell lying above 1000 m was introduced by Nyblade and Robinson (1994). They referred to the observation, that the combined eastern and southern African plateaus have had broadly similar uplift histories in the Mesozoic and Cainozoic. Summerfield (1996) later termed this feature the Southern African Superswell (Fig. 2.2). Superimposed



Names and approximate age

1 Ethiopia-Aden (Cenozoic)
2 Kenya (Cenozoic–Holocene)
3 Western Rift (Uganda–Zaire) (Late Cenozoic–Holocene)
4 East African (Late Cretaceous–Cenozoic)
5 Rungwe (Late Cenozoic–Holocene)
6 Chilwa–Zambesi (Cretaceous)
7 Zimbabwe (Early Mesozoic)
8 Transvaal (Precambrian)
9 Kimberley–Pretoria (Mesozoic)
10 Bushmanland–Sutherland (Late Cretaceous–Early Cenozoic)
11 Namibia–Uruguay (Mesozoic)
12 Angola–Southern Brazil (Cretaceous)
13 Cameroon (Cenozoic–Holocene)
14 Nigeria–north-east Brazil (Cretaceous)
15 Guinea–Guiana (Cretaceous)
16 Hoggar (Cenozoic–Holocene)
17 Tibesti (Cenozoic–Holocene)
18 Suez–Sinai (Cenozoic)

Figure 2 1 Morphological and structural elements of Africa (after Summerfield, 1985a).

on this large feature shorter wavelength-higher amplitude positive anomalies can be found (Summerfield, 1996).

2.1.1 Macro-scale southern African geomorphology

With the absence of the effects of the Pleistocene glaciations and Cainozoic mountain building from its landscapes, the geomorphology of southern Africa differs fundamentally from continents in the Northern Hemisphere (Moon and Partridge, 1993). Ancient landforms (e.g. the mid-Tertiary African erosion surface) are extensively preserved in southern Africa

Larger-scale structural framework

The present landscape of Africa is to a large extent the product of the break-up of Gondwana and the tectonic processes associated with this event (Summerfield, 1996). This statement underscores however the evidence of large-scale denudation that has occurred since the break-up. Summerfield's view has further been countered by the findings of Nyblade and Robinson (1994) and Partridge (1997a, b, c). Partridge found evidence that the present altitudinal distribution of southern Africa was caused by uplift, which occurred chiefly during the Neogene. It has further been suggested that the origin of the elevated topography of the African Superswell must be sought in a buoyancy, that resides deep within the earth's mantle (Partridge, 1997c). This supposition is confirmed by the studies on tomography of Su *et al.* (1994).

Moon and Dardis (1988) in a brief review of the regional geomorphology, agreed that the general outline of the southern African landmass dates from the break-up of Gondwana. Gondwana may be subdivided into east Gondwana (Antarctica, Madagascar, India, Australia) and west Gondwana (Africa and South America). The beginning of the break-up of Gondwana took place around 200 Ma which led to the separation of Antarctica from Gondwana around 180 Ma (Thomas and Shaw, 1993). The further division of east and west Gondwana took place around 142-133 Ma, and the eastern African coastline and the Indian Ocean formed thus by rifting. The South Atlantic Ocean came into being

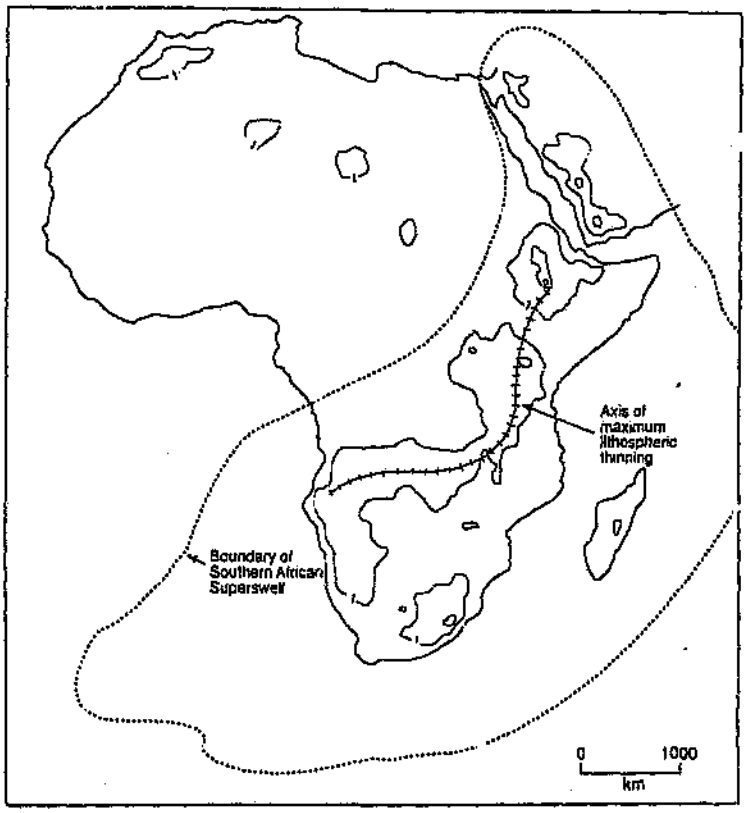


Figure 2.2 The Southern African Superswell (after Summerfield, 1996). Contours are drawn at 1 km intervals and are generalised from one-minute digital elevation data.

somewhat later, when west Gondwana broke up, and South America separated from Africa at 127 Ma. Full marine conditions were achieved in the south Atlantic by late Cretaceous times, around 80 Ma (Thomas and Shaw, 1993). The altitude of the prevailing landsurface, prior to the rifting of Gondwana, was estimated to range in southern Africa between 2350 m at Letseng Le Terai in the Lesotho Highlands to around 1800 m at Kimberley (Partridge and Maud, 1987). After initial significant downwarping along the continental margins, subsequently an elevated zone parallel to the continental margin developed in response to the continental fragmentation (Dardis and Moon, 1988). The elevated zone developed into the present Great Escarpment, which has retreated from the coast due to periods of uplift and erosion. Similar escarpments can be found in eastern Australia and India (ten Brink and Stern, 1992). The Great Escarpment determines to a great extent the macro-scale configuration of southern Africa, with a central depression bounded by a higher outward-facing escarpment, separated from the coastal regions. The influence of the presence of the rocks of the Karoo Sequence has been emphasised by Partridge and Maud (1987) and they suggested that, with the exception of the Cape Fold Mountains, the rocks initially covered almost all of southern Africa. As the Karoo rocks are susceptible to erosion and they were removed from large areas, the transfer of drainage patterns from the Karoo surface to more resistant underlying lithologies, took place through the process of superim tion. The implication of this has been discussed by various authors (e.g. Wellington 1929, 1941).

Development of the Kalahari Basin

The Kalahari Basin has been interpreted as a syneclise (Burollet, 1984; Thomas, 1988). The syneclise varies in age according to the sediments it contains on either side of the Ghanzi Ridge. While the oldest deposits in the southern syneclise are Dwyka glacial tillites of Permo-Carboniferous age (Thomas, 1988), the basin north of the tectonically stable Ghanzi Ridge does not contain any sediments from before the Mesozoic (Fig. 2.3). These two basins may have continued to evolve independently after the break up of Gondwana (Thomas and Shaw, 1990). South of the Ghanzi Ridge, three deep sedimentary ancient basement basin structures have been identified through an

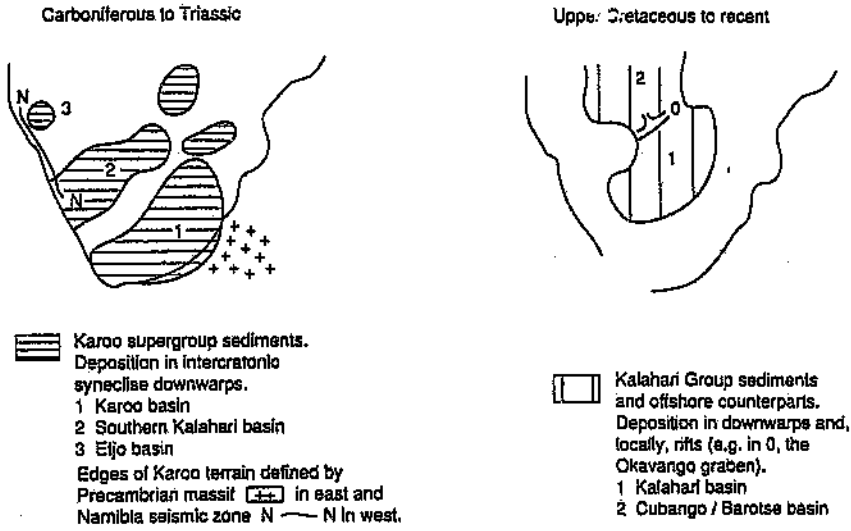


Figure 2.3 Sedimentary basins and deposits in southern Africa at two periods of the Phanerozoic (modified after Dingle *et al.*, 1983).

aeromagnetic survey in the mid-1970s (Meixner *et al.*, 1985). These are the Passarge, Ncojane and Nossob Basins (Fig. 2.4). The sedimentary infill of the Ncojane and Nossob Basins is presumed to be of Nama and Karoo age, while in addition to the Nama and Karoo sediments, equivalents of Precambrian sediments may even occur in the Passarge Basin. This would make the Kalahari Basin structures much older than the break-up of Gondwana.

The development of the present Kalahari Basin has been interpreted both as an area of subsidence directly related to continental margin rift flank uplifts (e.g. de Swardt and Bennet, 1974; Summerfield, 1985b; Thomas, 1988; Thomas and Shaw, 1991; ten Brink and Stern, 1992) and as an area that lagged behind during Neogene regional uplifts (e. g. du Toit, 1933; Partridge and Maud, 1987). The presence of an elevated hingeline along the margins of southern Africa, from which the surface of the continent sloped steeply outwards towards graben scarps, and more gently inwards towards internal basins (e.g. Kalahari Basin) was envisaged by de Swardt and Bennet (1974). The influences of sediment loading on the margin and isostatic uplift at the margin due to erosion on the hinge zone were further discussed by Summerfield (1985b). Thomas (1988) and Thomas and Shaw (1991) supported the idea that marginal uplift generated a flexural bulge, which developed into the Great Escarpment, with subsidence occurring in the pericratonic offshore basins and the intercratonic Kalahari Basin (Fig. 2.5). A theoretical model, where the uplift of the Great Escarpment is seen as an upward deflection of a continuous elastic plate (Fig. 2.6), which is modified by the downward load of sediments on the continental margin was developed by ten Brink and Stern (1992). They saw the Kalahari Basin as a hinterland flexural basin or 'outer low' which subsided simultaneously with uplift along the west coast of southern Africa, following the opening of the South Atlantic Ocean. Their model suggested that the centre of uplift and the centre of subsidence shifted inland with time (Fig. 2.7). Although certainly the above mentioned mechanisms will be at play, they have not conclusively explained the existence of the high topography and Kalahari Basin in southern Africa (Partridge, pers. comm., 1998). Nyblade and

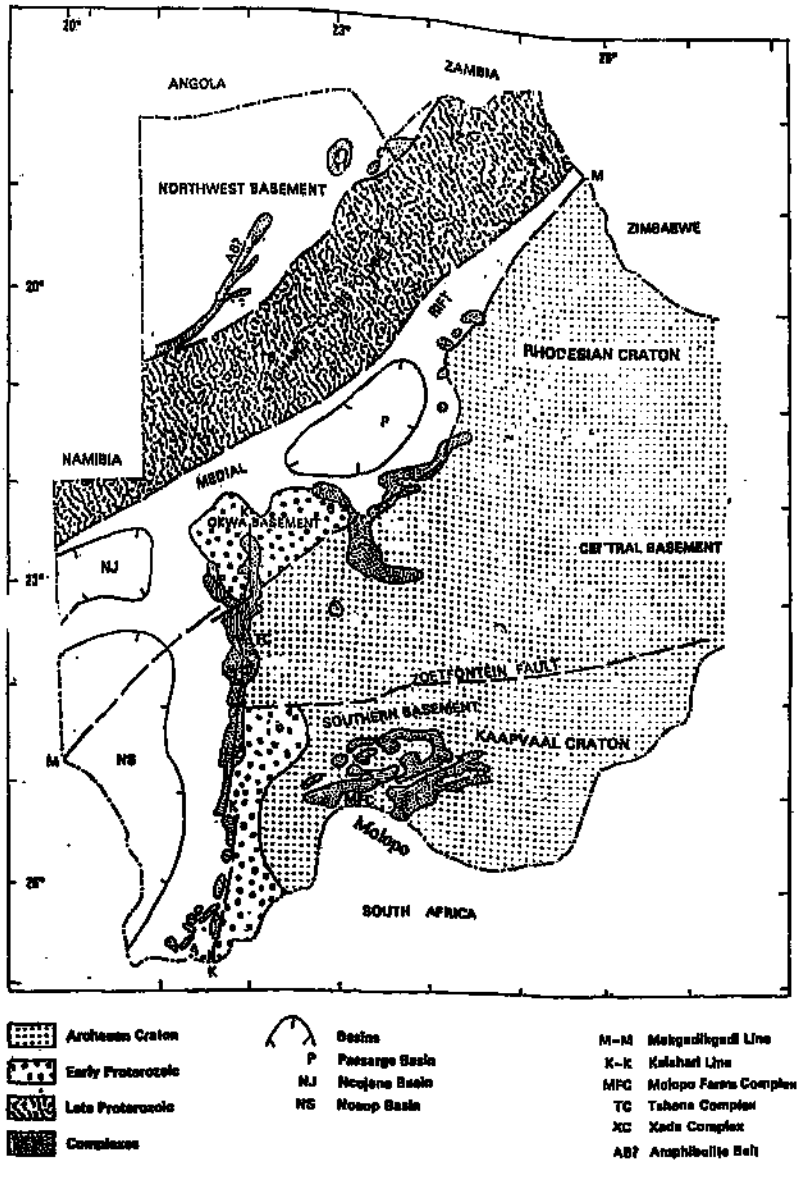


Figure 2.4 Synopsis of basement features in Botswana (after Meixner *et al.*, 1985).

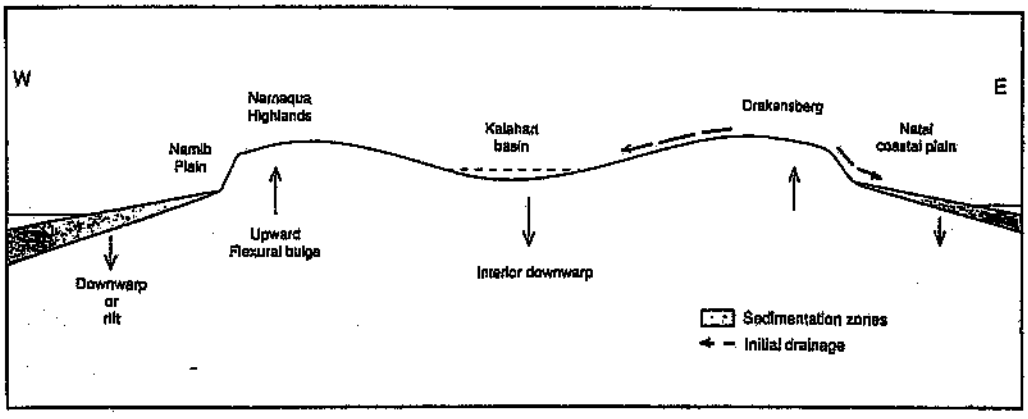


Figure 2.5 Schematic cross section of southern Africa following the splitting of Gondwana (after Thomas and Shaw, 1991).

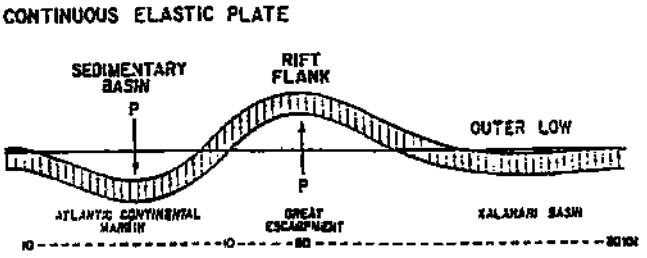


Figure 2.6 Continuous elastic plate model of southern Africa (after ten Brink and Stern, 1992)

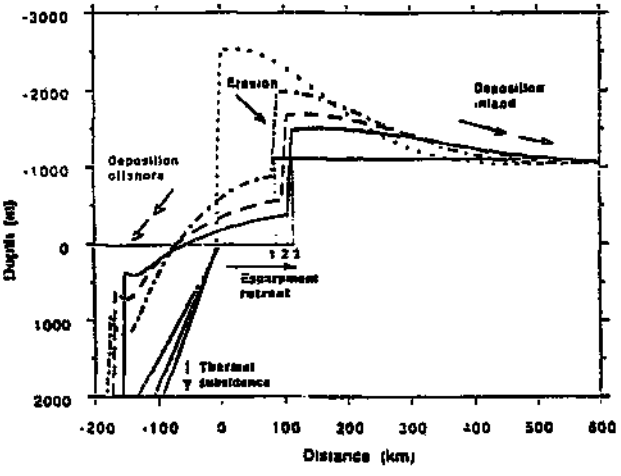


Figure 2.7 Evolution of the continuous plate model through time (after ten Brink and Stern, 1992). The initial uplift profile (heavy dotted line) is eroded by three erosional cycles, and the topography is allowed to rebound by regional isostasy after each cycle (other heavy lines).

Robinson (1994) also indicated that the African superswell may have resulted from a combination of uplift mechanisms.

The similarity in the orientation of the axes of uplift and the longer diameters of the areas of depression, which tend to be north-east to east-north-east, and the possibility that they owed their origin to one and the same controlling set of tectonic forces was noted by du Toit (1933). He suggested that the uplift of the continent must have varied and that this caused the warpings including the Kalahari Basin. Although Partridge and Maud (1988) did not provide a larger-scale mechanism, that caused the uplifts and monoclinal tilting that affected the margins of the sub-continent during the Cainozoic, they did suggest that the first basining of the Kalahari may have occurred in response to uplift of surrounding areas between 80 and 100 Ma ago (Partridge and Maud, 1987). They hinted that the formation of the Kalahari Basin could have been related to a rejuvenation of much older basin and swell structures related to the 'Pan-African Event'. Partridge and Maud further stated that the uplifts during the Miocene and Pliocene were concentrated along specific axes (e g. Ciskei-Swaziland and Griqualand-Transvaal axes of uplift, see Fig. 2.8), and that the area of the Kalahari Basin subsequently lagged behind in these uplifts, and thus further deepened the Kalahari Basin. Recently the Neogene uplifts in southern Africa Partridge (1997a, b, c) have been linked with the formation of the African Superswell (Nyblade and Robinson, 1994).

The southern African erosion surfaces

The development of the landscape at the macro-scale has been a controversial issue (Dardis and Moon, 1988). The macro-scale geomorphology in southern Africa has focused long on the identification and correlation of erosion surfaces. Diverse numbers of surfaces have been recognised at various times. The different interpretations were related to the correlation of erosion surfaces across the Great Escarpment. A review of the literature related to this issue can be found in Partridge and Maud (1987, 1988), Dardis and Moon (1988), and Moon and Partridge (1993). Partridge and Maud (1987) and Partridge (1997b) provided a reinterpretation of the earlier material based on the

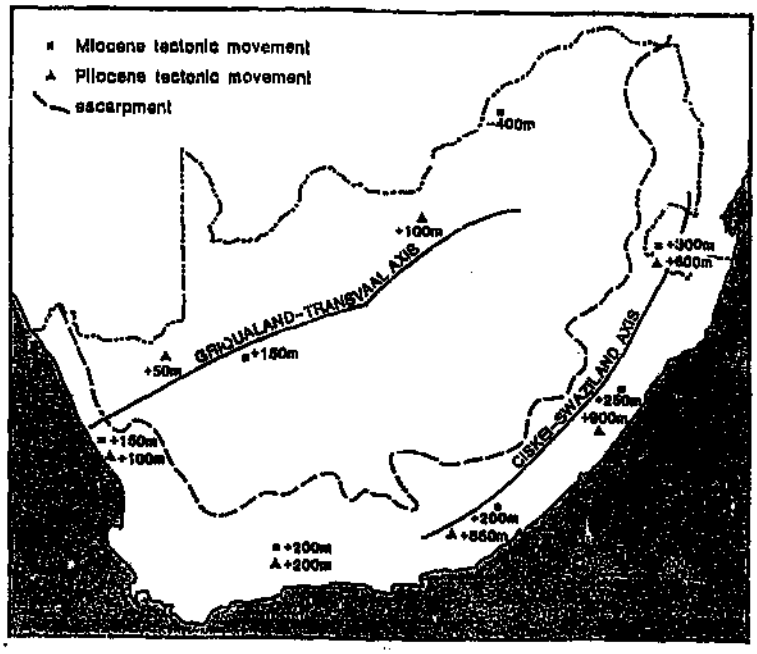


Figure 2.8 Axes of maximum uplift during the Miocene and Pliocene (after Moon and Dardis, 1988).

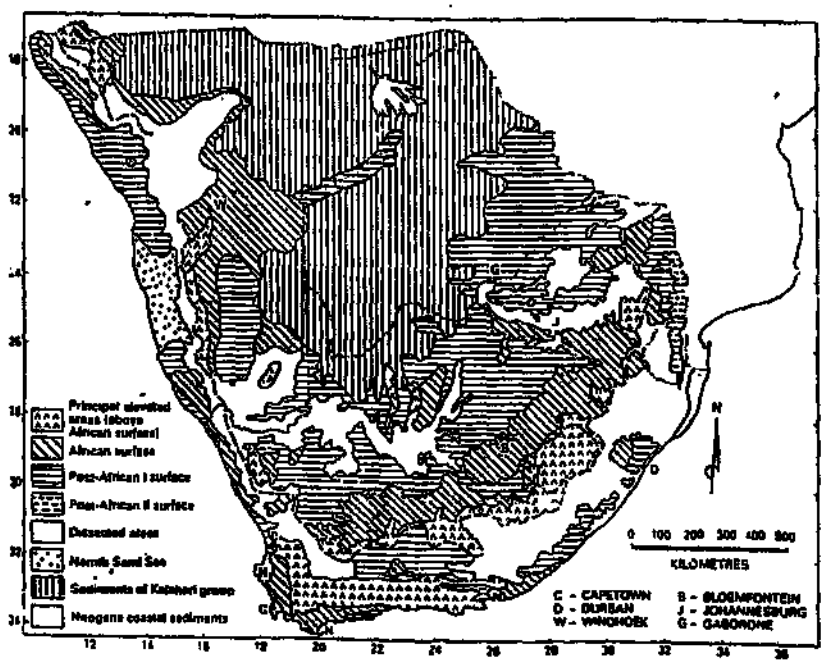


Figure 2.9 Simplified map of the principal cyclic landsurfaces of southern Africa (after Partridge, 1997b).

work of previous authors (e.g. Dixey, 1938, 1942, 1955; King, 1949, 1951, 1955), and recognised the following erosion surfaces. The oldest and most wide-spread surface was termed the African surface (Fig. 2.9). The surface below the African surface in the interior has been named the Post-African, while in the coastal areas two Post-African surfaces have been identified; Post-African I and Post-African II. In addition to that, mountainous areas rising above the African surface and the Great Escarpment were recognised, while depositional landscapes in the interior were identified as areas dominated by Kalahari sediments. The review of the geomorphological evolution of the interior of southern Africa, below, is largely based on the work of Partridge and Maud (1987) and Partridge (1997b).

Gondwana influences

The pre-rifting surface, which was termed the 'Gondwanaland' surface by King and King (1959) is according to Partridge and Maud (1987) and Partridge (1997b) nowhere preserved as a sub-aerial planation feature. They further suggested that it also has not been preserved beneath the Kalahari Group sediments, as the earliest dated sediments post-date the break-up of Gondwana by about 40 Ma. Major features of the Gondwana surface, of importance to the later geomorphic evolution of the interior of southern Africa, include: the high initial elevation of the continent, the overall westward trend of the drainage, and the presence of tabular Karoo rocks beneath most of the landsurface (Partridge and Maud, 1987). The rifting of Gondwana initiated the landscape cycle which produced the African surface and created the Great Escarpment. The latter was of great geomorphological importance, as it separated the elevated interior from the coastal margins. This meant that from then onwards a dual drainage system evolved (Thomas and Shaw, 1990), and erosion proceeded simultaneously at different levels on either side of the Great Escarpment. In the coastal margins relatively short, steep streams extended from the hingeline to the coast and erosion was controlled by the oceanic base level. The base level for the erosion in the interior plateau was the point of outlet at the Great Escarpment of major low gradient river systems such as the Orange/Vaal.

Major geomorphological aspects of the African surface in the interior

Although the African landscape cycle was of a poly-cyclic nature, the surfaces must be regarded as a single unit within the southern African landscape (Partridge and Maud, 1987). The first sediments generated in the interior during the African cycle were deposited in the present Kalahari Basin and probably constitute the basal deposits of the Kalahari Group, the Wessels Gravel Formation. Those deposits, where preserved in diatremes of the southern Kalahari, have been assigned a late Cretaceous age on palynological grounds (Partridge and Maud, 1987). They suggested that the basal Kalahari Group sediments overlie the African surface, and indicate the switch from erosion to deposition in the Kalahari Basin at the end of the Mesozoic. It appears that the major part of the erosion within the landscape cycle occurred in the period immediately following rifting, and that by late Cretaceous to early Tertiary times the continent had been reduced to a vast undulating plain, with elevations in the interior of 500-700 m above sea-level, and some remaining areas of higher ground.

The drainage network established on the African surface, exerted an important influence on the later geomorphic evolution of the subcontinent (Partridge and Maud, 1987). Two major westward draining river systems existed at that stage: the upper proto-Orange system, which entered the Atlantic via the lower course of the Olifants River, and the lower proto-Orange system, which entered the Atlantic near the point of entrance of the present Orange River. A pre-Kalahari northward-flowing drainage line for the Molopo drainage (section 1.2.2) has unequivocally been put forward by Smit (1977). McCarthy (1983), however, indicated the presence of a major southward-flowing drainage system, the 'Trans-Tswana', across what is now the Kalahari Basin, prior to the accumulation of the Kalahari group sediments. This Trans-Tswana would have connected with the upper proto-Orange River. The existence of a proto-Nossob/Molopo, situated near its present position, as an extensive southward flowing system, across the present southern rim of the Kalahari Basin was further suggested by Partridge (1997b). As the southern rim of the Kalahari became better defined, a significant northward component was added to the drainage, and with progressive deepening of the basin the southward-draining rivers

became dismembered and were replaced by a centripetal system of northward flowing channels (Partridge and Maud, 1987). Towards the east of the upper proto-Molopo diamondiferous gravels were deposited during the Cretaceous by a proto-Harts system, that was during the late Cretaceous extended far to the north of the continental divide, now separating the Limpopo drainage from the Orange/Vaal system.

The Post-African landscape cycle in the interior

The African cycle of landscape evolution was brought to an end by uplift of the subcontinent during the Miocene (Partridge, 1997b). The uplift was concentrated along several axes, of which the Ciskei-Swaziland axis was the most prominent with an uplift between 200 and 300 m (Partridge and Maud, 1987). The Ciskei-Swaziland axis is located 80 to 100 km inland of the coastline and is thus situated well seaward of the line of the Great Escarpment (Partridge and Maud, 1988). On the interior plateau the uplift was concentrated along the western extension of the Griqualand-Transvaal axis and was estimated at 150 m. This resulted in a slight westward tilt of the subcontinent. The uplift along the interior axis further accentuated the Kalahari basining, and probably resulted in increased sedimentation rates. The Budin and Eden formations, which form the middle section of the Kalahari succession, were probably related to the Miocene and Pliocene periods of uplift in the interior (Partridge and Maud, 1987). More or less contemporaneous with the Miocene uplift . subsidence of 400 m of the Bushveld Basin occurred. The upper most terraces of the Vaal River near Kimberley have yielded terrestrial fossils of Pliocene age, and are found along-side the Post-African surface (Partridge and Maud, 1987). Partridge and Maud further stated that, as a result of the Miocene rise of the Griqualand-Transvaal axis, the drainage patterns of the interior were affected. Rejuvenation in the vicinity of the axis resulted in the capture of the pre-existing upper proto-Orange system, by the Koa tributary of the lower proto-Orange system. This interpretation differs from Dingle and Hendey (1984), who identified two separate periods of capture. de Wit (1993) suggested that the integration of the upper and lower proto-Orange systems, rather, was due to two humid climatic pulses, and was not related to any axial uplift. During the latter part of the Tertiary the Molopo River was

already largely established in its present position, and due to the reintegration of the Nossob/Molopo drainage on top of the sedimentary filling of the Kalahari Basin, the river system managed to breach the southern rim of the Kalahari Basin (Partridge and Maud, 1987). The Kalahari-Zimbabwe axis (du Toit, 1933) situated to the north of the Griqualand-Transvaal axis, was likewise affected by Tertiary uplift, and greatly influenced the Zambezi and Limpopo drainage systems (Thomas and Shaw, 1990, 1991, 1992, 1993).

Geomorphic aspects of the Pliocene uplift in the interior

During the Pliocene both tectonic uplift and substantial climatic changes occurred. The Pliocene uplift (between 5 and 3 Ma) was concentrated along the same axes as in the Miocene, but the Pliocene movement was of much greater amplitude (Partridge and Maud, 1987; Partridge, 1997b). The uplift along the Ciskei-Swaziland axis varied from 600 to 900 m, while the uplift along the Griqualand-Transvaal axis was indicated between 50 and 100 m. Although uplift and marginal tilting also affected the southern and western margins (Partridge and Maud, 1988), the movements were not as great as in the eastern margins. This unequal uplift resulted in a further accentuation of the westward tilt of the subcontinent. The uplift of the Kalahari Basin again lagged behind, which caused a further deepening of the basin, and probably initiated renewed sedimentation. Partridge and Maud (1987) cited du Toit (1933) to state that the eastern part of the Griqualand-Transvaal axis originated only as a result of the Pliocene uplift. Marshall (1991) contended this with evidence of uplift along an incipient form of that section of the Griqualand-Transvaal axis, that could date back to Cretaceous times.

The position of the axis of the Permo-Carboniferous Cargonian Highlands as proposed by Visser (1987), coincides to a large extent with that of the proposed Cainozoic Griqualand-Transvaal axis of du Toit (1933). A gross structure of alternate interior basins and upwarped zones or 'swells' has characterised considerable areas of Gondwana from Proterozoic times onward (Partridge and Maud, 1987). From late Precambrian to early Palaeozoic times a major part of the African continent has been affected by a

thermotectonic event (Truswell, 1977), which was termed the Pan African Event. Cahen *et al.* (1984) indicated two separate periods of uplift dated at 1150-1100 Ma and 600-560 Ma. Truswell (1977) cited Clifford (1970) who stated that the Pan-African orogeny culminated at 550±100 Ma. The erosive feature of the Cargonian Highlands as proposed by Visser (1987), may initially have been formed as part of the 'Pan-African' episode of convection-generated orogenesis within intercratonic belts. The highlands could already have been in existence as a relative high during the deposition of the sediments of the Cape Supergroup, during the lower Palaeozoic. Extensive erosion during the Palaeozoic and Mesozoic would have levelled the Cargonian Highlands, and this ancient structure might have become rejuvenated during the Neogene in the form of the Griqualand-Transvaal axis, as the subcontinent was affected by the Great African Plume (Su *et al.*, 1994). Cahen *et al.* (1984:416) also stated that most rifting and warping in Mesozoic and Cainozoic times occurred on the site of former late Precambrian to early Palaeozoic tectono-thermal activity.

Partridge and Maud (1987) suggested, in agreement with Mayer (1983), that the Pliocene axis was in places located a considerable distance to the south of its Miocene counterpart. Partridge and Brink (1967) indicated that the alluvial gravels of the older 30-60 m terraces of the lower Vaal River were largely deposited in response to the Pliocene rejuvenation of the Griqualand-Transvaal axis. Incision in the Post-African II cycle resulted in the development of lower terraces along the major rivers (e.g. Vaal River) in response to a variety of structural, geomorphic and climatic factors (Partridge and Maud, 1987). Pliocene uplift along the Kalahari-Zimbabwe axis resulted in further modification of the Limpopo drainage.

Two periods of climatic amelioration, which may have resulted in Pliocene warming, have been inferred to have occurred successively between 5.1 and 4.2 Ma and 3.5 and 2.8 Ma (Partridge, 1997b). Partridge suggested that during this period the re-established exoreic Molopo drainage managed to break through the southern rim of the Kalahari Basin. A major decline in temperature characterised the late Pliocene. Concurrent

progressive aridification during the latter part of the Pliocene (~2.8 Ma) and further during the Pleistocene period caused the lower part of the Nossob/Molopo system to become defunct and to be transgressed by dunes of Kalahari Sand (Partridge and Maud, 1987; Partridge, 1993, 1997b). The Kalahari Sand (Gordonia Sand Formation) forms the terminal unit of the Kalahari Group. The extensive mantle of Kalahari Sand was deposited over distances well beyond the boundaries of the Kalahari Basin. The flat sediment-filled Kalahari Basin had thus become a sedimentary plain or 'sedi-plain' (Twidale, 1968) with a flatness similar to well-developed erosion surfaces.

2.1.2 Regional geology

The surface geology of the Molopo drainage basin area is dominated by the sediments of the Cainozoic Kalahari Group. More than 74% of the area has a sediment cover with a thickness of 15 m or more. Only the hills are devoid of sand cover and in addition the rim area has limited occurrences of bedrock near the surface which are partially or only thinly covered with Kalahari Group sediments. Three parallel ranges of hills protrude through the sediment cover in the rim area. In the east is the Asbestos-Kuruman-Makhubung hills range, then the Langeberge-Korannaberg, and further to the west the Skurweberge-Witberg. Towards the centre of the Kalahari Basin these ranges of hills disappear under the Kalahari Group sediments.

Discrepancies between the Botswana and South African geological maps

The Molopo River, which serves for over 650 km as an international border, divides the drainage basin into two sections, that are covered by the 1:1000 000 geological maps of both Botswana and South Africa. Although there is general concordance on the geology of the area (Table 2.1), discrepancies and uncertainties with respect to correlations do exist along the border. These discrepancies can largely be contributed to the extensive, obscuring, sediment cover of the Kalahari Group.

Table 2.1 Comparative legend for the South African and Botswana sections of the geological map of the Molopo drainage area. Based on the 1 : 1000 000 geological maps of Botswana and South Africa.

SOUTH AFRICA					BOTSWANA		
		KALAHARI GROUP		T-Qk			
KAROO SEQUENCE	ECCA GROUP	Pietermaritzburg Formation	Shale	Pp	Kl	Orange, red or white aeolian sandstone, mudstone	
		Dwyka Formation	Tillite, sandstone, mudstone, shale	C-Pd	K	Arkose, carbonaceous, mudstone, coal, shale and tillite	KAROO SUPERGROUP
	NAMA GROUP	Kuibis Formation	Quartzite	Nku			
	KORANA-LAND SEQUENCE	Toeslaan Formation	Kinzigite	Mto			
		Gnede Hoop Formation	Pink gneiss, quartzite, schist, amphibolite	Mgo			
		Groblershoop Formation	Schist, quartzite, metalava	Mg	Ng	Norite, gabbro	MOLOPO FARMS ULTRA-BASIC COMPLEX
OLIFANTS-HOEK SEQUENCE		Brulsand Formation	Quartzite, shale, subgreywacke	Mbr	Ub	Diorite, pyroxenite, harzburgite	
	VOLOP GROUP	Matsap Formation	Subgreywacke, quartzite, conglomerate	Mm			
		Lucknow Formation	Quartzite, limestone	Vlu	W	Red conglomerate, coarse sandstone, shale	WATERBERG SUPERGROUP
GRIQUA-LAND WEST SEQUENCE	COX GROUP	Ongeluk Formation	Andesite	Vo			
	GRIQUATOWN GROUP	Asbestos Hills Formation	Iron formation, jaspilite	Va	Tu	Shale, sandstone, quartzite, conglomerate	
	CAMPBELL GROUP	Ghaap Plateau Formation	Dolomite, limestone, chert	Vgh	Tm	Banded ironstone	TRANSVAAL SUPERGROUP
TRANSVAAL SEQUENCE	CHUNIESPOORT GROUP	Malmani Subgroup	Dolomite, chert	Vm	Tl	Dolomite, chert, shale, quartzite	
	VENTERSDORP SUPERGROUP	Allanridge Formation	Andesite	Val			
		DOMINION GROUP	Andesite, quartz, porphyry, quartzite	Rd			
		BASEMENT COMPLEX	Granite, gneiss	ZA	0G4	Monzonite, tonalite, granite, adamellite	
			Chert, iron formation, jaspilite, schist, lava	Zk	X	Quartzo-feldspathic gneiss, gneissic rocks	BASEMENT GNEISS COMPLEX

Gould and Rathbone (1985), for example, discussed a case of changing interpretations of purple to brown sandstones at Makopong. These rocks were initially interpreted as belonging to the Waterberg System (Smit, 1974a). Subsequently they were included in the Volop Group of the Olifantshoek Sequence (SACS, 1980). Gould and Rathbone (1985), however, stated that the rocks were undoubtedly equivalent to the Waterberg rocks in the north of the area. Ludtke (1985) further discussed the case of red to maroon glassy quartzites near Tshabong. These rocks had traditionally been interpreted as belonging to the Matsap-Waterberg Group (du Toit, 1954; Boocock and van Straten, 1962), but Mallick *et al.* (1981) suggested that they be of Griqualand West-Transvaal age. There are further discrepancies in the matching of the maps along the border with the Karoo Sequence. The subdivision of the Karoo Supergroup (K and K1 on the Botswana geological map) does not entirely match the subdivision of the Karoo Sequence (SACS, 1980) into Pietermaritzburg Formation and Dwyka Formation on the South African map. Furthermore, the mainly sediment-covered rocks of the Transvaal Supergroup in Botswana, are found in an intermediate position between the rocks of the Transvaal Sequence of the Bushveld Basin to the east, and the similar rocks of the Griqualand West Sequence to the south of them in south Africa (Gould and Rathbone, 1985). They are, however, physically continuous with both sequences in South Africa.

For the purposes of this study the description of the geology of the drainage basin area is subdivided into four parts: the rocks of the Basement Complex, the late-Archaeozoic to Proterozoic rocks, the rocks of the Karoo Sequence, and the sediments of the Kalahari Group.

Basement Complex

The Basement Complex is made up of the Kraaipan Group rocks (Zk) and ancient granites and gneisses (ZA) on the South African side of the border and quartzo-feldspatic gneiss and gneissic rocks (X) as well as monzonite, tonalite, granite and adamellite (f/G4) of the Basement Gneiss Complex on the Botswana side. The Basement Complex rocks form a kidney-shaped body in the eastern section of the drainage basin (Fig. 2.10) and are

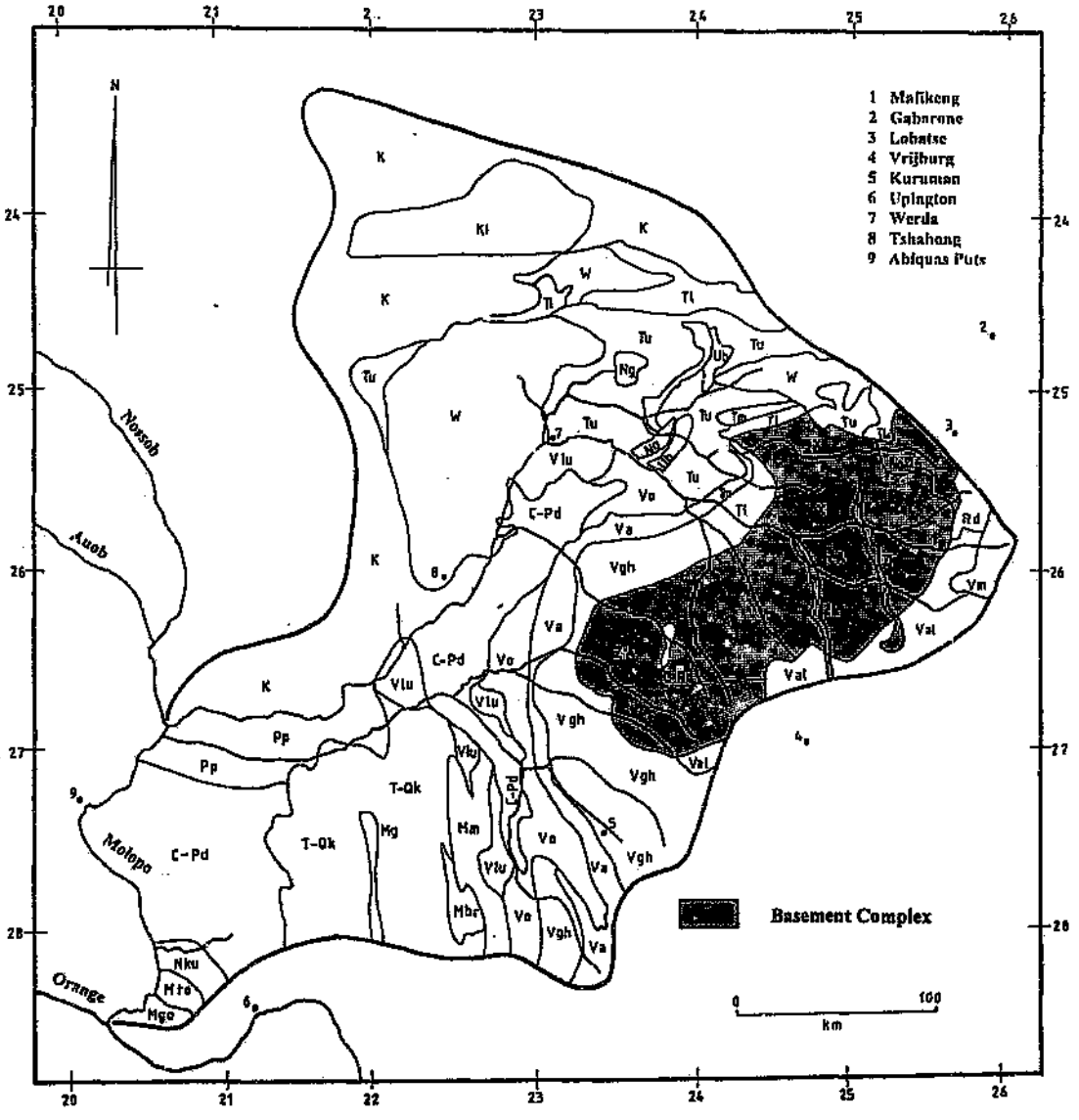


Figure 2.10 Pre-Kalahari geology, emphasising Basement Complex rocks in the Molopo drainage area. See Table 2.1 for the legend.

generally covered by thin sediments (< 15 m) of the Kalahari Group. The Kraaipan Group rocks are made up of ironstone, banded iron stone, jaspillite, banded and massive chert, and are interbedded in a succession of metavolcanic and metasedimentary rocks (Michaluck and Moen, 1991). Exposed outcrops of the Kraaipan Group occur in two north-south trending belts. The only natural outcrops of the Archaean granites and gneisses (ZA) occur within the Molopo Valley and its tributaries in the area.

Late Archaeozoic and Proterozoic rocks

Late Archaeozoic and Proterozoic rocks are these formed subsequent to the Basement Complex, but prior to the Karoo Sequence. They include a variety of rocks which have exerted a great influence on the geomorphology of the area. The near-horizontal dolomites of the Malmani Subgroup (Vm) and the Ghaap Plateau Formation (Vgh) form almost flat areas with Karst features to the east of Mafikeng and on the Ghaap Plateau between Vryburg and Kuruman (Fig. 2.11). The more erosion-resistant rocks of the Griqualand West, Olifantshoek and Korannaland Sequences form three sub-parallel ridges on the southern rim of the Kalahari Basin, which have strongly influenced the development of the Molopo drainage. A third area of rocks, belonging to this assembly, is found in the south-west section of the drainage basin. Except for the above-mentioned ridges, the rocks in the eastern half of the drainage area are covered by a substantial cover of Kalahari Group sediments, exceeding 210 m in places. In the western section they are generally covered by thick deposits of both the Karoo Sequence and the Kalahari Group.

The area to the east of the basement complex is dominated by outcrops of the andesites of the Allanridge Formation (Val) and the dolomites of the Malmani Subgroup (Vm). To the west and south of the Basement Complex lie the outcrops and suboutcrops of the Griqualand West Sequence and the Olifantshoek Sequence. The oldest formations are next to the Basement Complex and the younger formations occur successively further away of the complex. The dolomites of the Ghaap Plateau Formation (Vgh) form the near-horizontal Ghaap Plateau. The more resistant rocks of the Asbestos Hills Formation (Va) form the prominent, curved ridge of the Kuruman Hills and the Makhubung Hills.

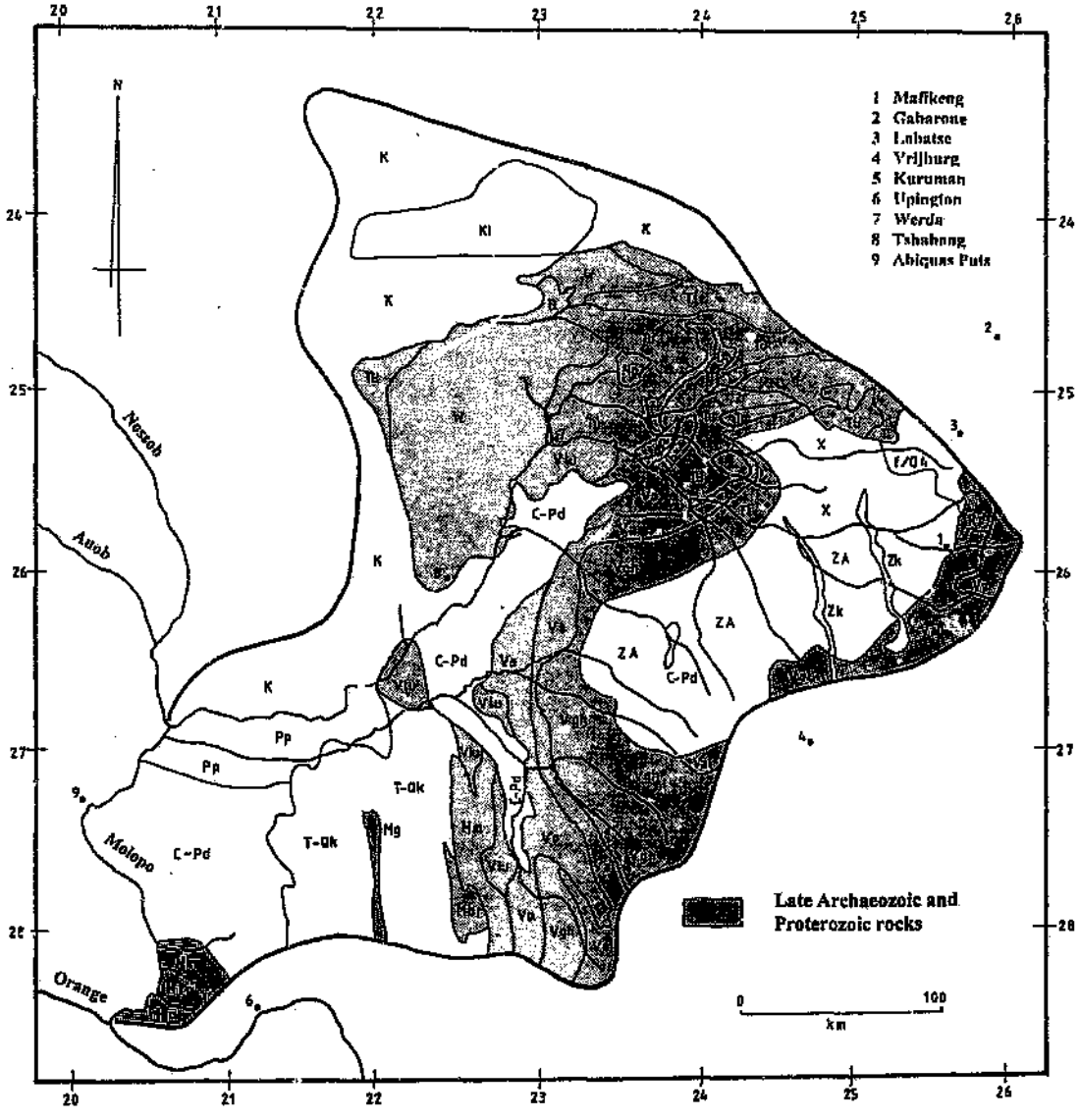


Figure 2.11 Pre-Kalahari geology, emphasising late Archaeozoic and Proterozoic rocks in the Molopo drainage area. See Table 2.1 for the legend.

The Kuruman Hills have a general SSE-NNW orientation, while the Makhubung Hills have a SSW-NNE orientation. Further to the west lies the topographically lower belt of the andesites of the Ongeluk Formation (Vo). Smit (1974b) indicated that the ancient glaciers of the Dwyka glaciation carved deep valleys into the andesites of the Ongeluk Formation. Adjacent to the Ongeluk Formation occur the Lucknow (Vlu), Matsap (Mm), and Bruisand (Mbr) Formations of the Olifantshoek Sequence. The erosion-resistant rocks of the Matsap Formation give rise to the Langeberge and Korannaberg north-south range of hills which disappears towards the north under the sediments of the Kalahari Group. A third and smaller north-south range of hills occurs further west where quartzites of the Groblershoop Formation (Mg) outcrop beyond the sediments of the Kalahari Group. In the south-western extremity of the drainage area there are minor outcrops of the Korannaland Sequence and the Nama Group.

In Botswana, to the north of the Molopo River, the rocks of the Griqualand West and Olifantshoek Sequences are correlated with the rocks of the Transvaal and Waterberg Supergroups (Table 2.1). The rocks are there generally covered by substantial deposits of the Kalahari Group and correlation is in places uncertain. In the Molopo Farms area north of the Molopo River, between $23^{0}00'$E and $24^{0}30'$E and an area from $24^{0}30'$S to the Molopo River, are the intrusive rocks of the Molopo Farms Ultrabasic Complex (Rathbone and Gould, 1982; Gould and Rathbone, 1985; Gould *et al.*, 1987, 1989). The complex has a lopolithic form and is very similar in both age (post-Transvaal, but pre-Waterberg) and geological setting to the Bushveld Igneous Complex in South Africa (Fig. 2.12)(Gould *et al.*, 1987).

Karoo Sequence

Outcrops and suboutcrops of the Karoo Sequence (RSA) and the Karoo Supergroup (Botswana) are found generally in the western half of the drainage basin (Fig. 2.13). The outcrops of the Dwyka Formation (C-Pd) are limited to the most south-westerly section, while the suboutcrops of both the Dwyka formation and the Pietermaritzburg Formation (Pp) are generally found underneath the sediments of the Kalahari Group. The Dwyka

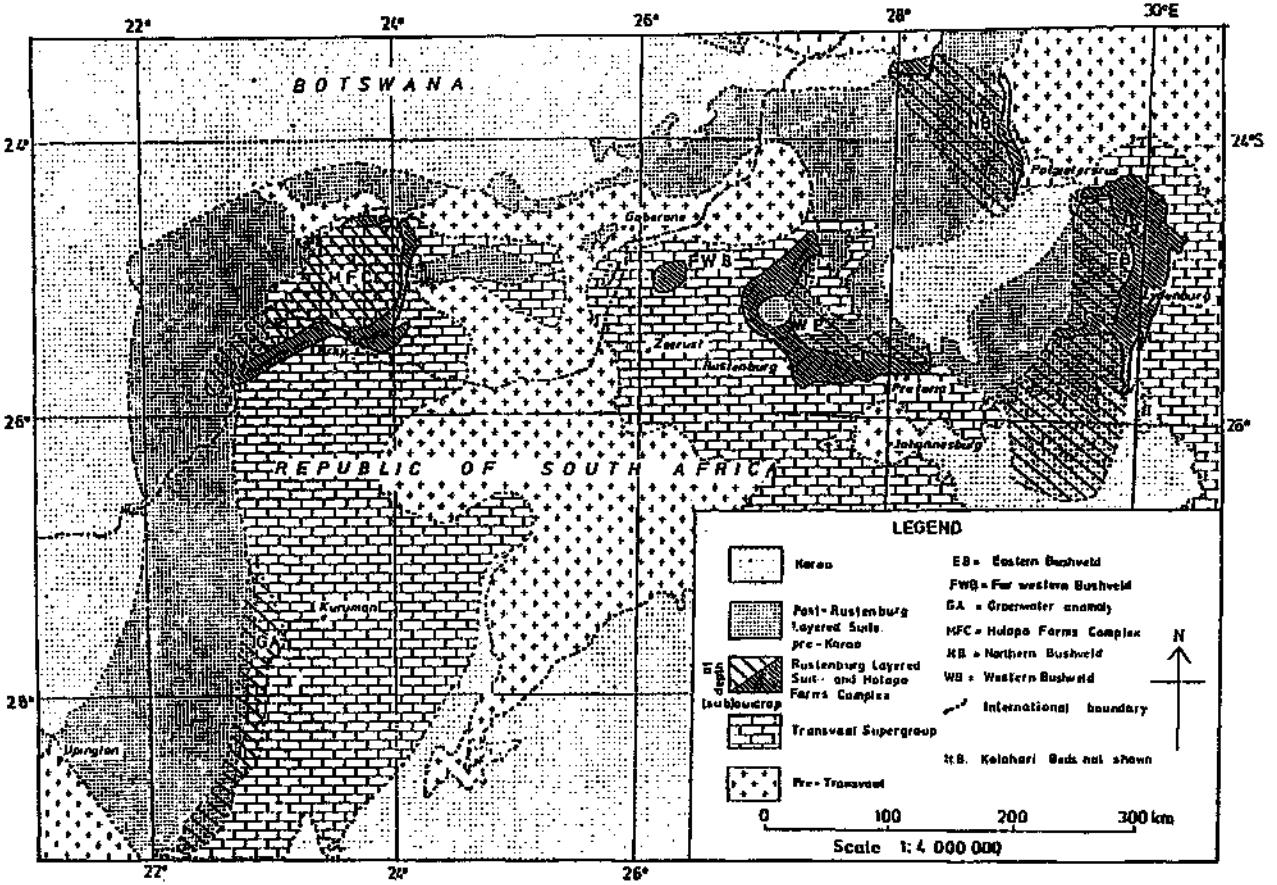


Figure 2.12 Geological sketch map showing extent of basic and ultrabasic rocks correlated with the Bushveld Complex (after Gould *et al.*, 1987).

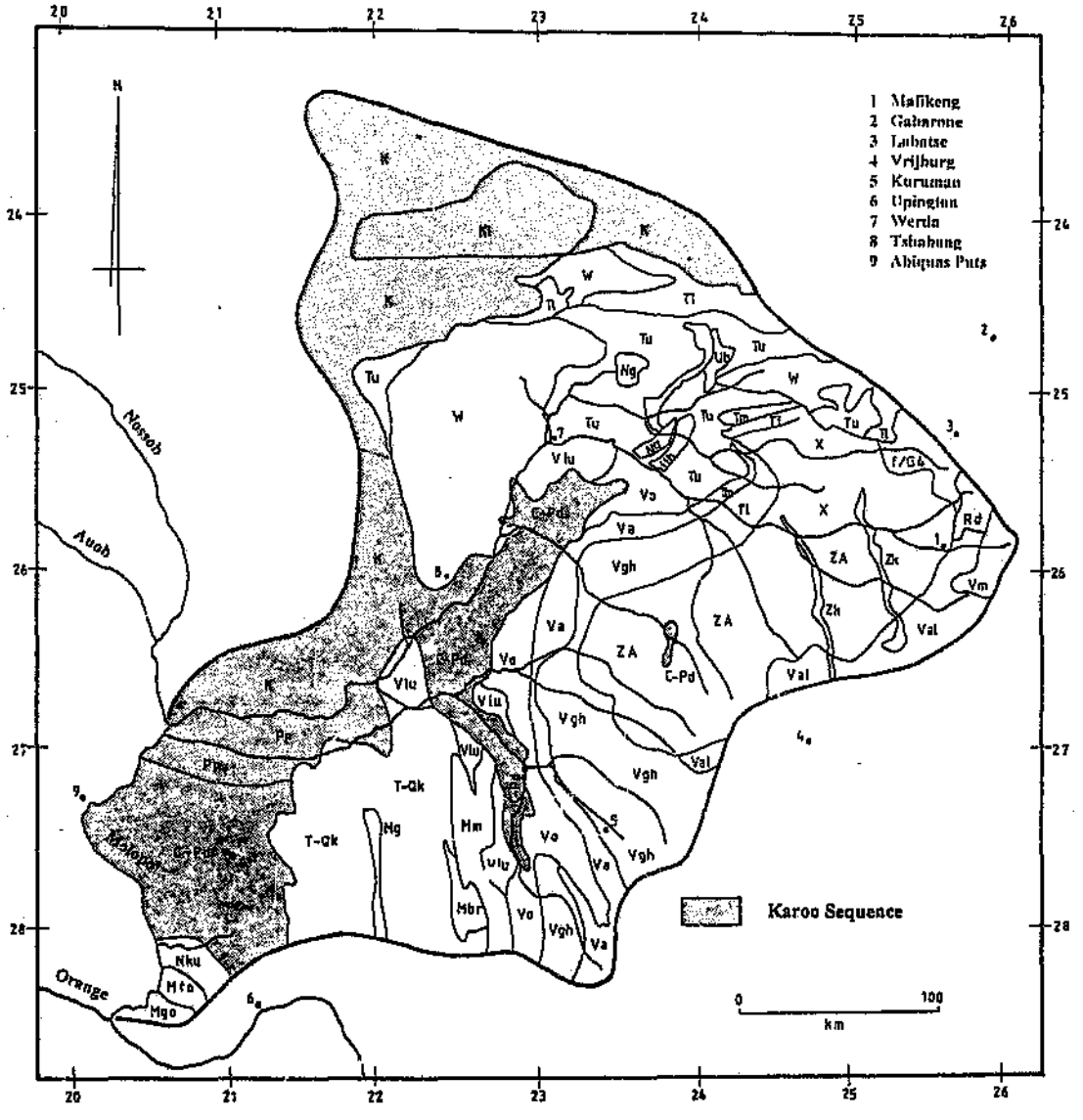


Figure 2.13 Pre-Kalahari geology, emphasising Karoo Sequence rocks in the Molopo drainage area. See table 2.1 for the legend.

Formation consists of tillites, sandstone, mudstone and shale, while the Pietermaritzburg Formation consists mainly of shale. Some of the thicker deposits of the Dwyka have been interpreted as suboutcrops, that fill ancient, pre-Karoo drainage lines (Smit, 1974b, 1977; Visser, 1983a; section 3.2.1). Across the Molopo River on the Botswana side of the drainage basin the rocks are correlated with the subdivision (K and K1) of the rocks of the Karoo Supergroup. Arkose, carbonaceous mudstone, coal, shale and tillite are grouped in the K subdivision, while orange, red to white sandstone and mudstone make up the K1 subdivision.

Kalahari Group

The development of the stratigraphy of the sediments of the Kalahari Basin has been reviewed by Thomas (1988), Thomas and Shaw (1990, 1991). In particular the sediments of the area of the South African section of the basin have been described by Smit (1977), SACS (1980), Thomas *et al.* (1988), and Thomas and Thomas (1989). The Kalahari Group has been subdivided into the Wessels, Budin, Eden, Mokalanen, Gordonia, and Lonely Formations by Thomas and Thomas (1989). The Wessels and Budin Formations do not crop out and are known mainly from boreholes. The Eden Formation consists predominantly of sandstones which are well exposed along the Molopo River. The Mokalanen Formation, which consists of calcrete and diatomaceous limestone, as well as the Lonely Formation, which consists of diatomaceous limestone (Thomas and Thomas, 1989), are not yet approved by the South African Committee for Stratigraphy. The Gordonia Formation is predominantly made up of red aeolian sand.

Figure 2.14 shows the distribution of the Kalahari Group sediments with a thickness that exceeds 15 m within the drainage area. It appears that the sediments have buried the pre-Kalahari topography of which only the higher sections still protrude. The deposits are generally thinner at the easterly and southerly rim of the Kalahari Basin. An understanding of the thickness variation of the sediments is important for the

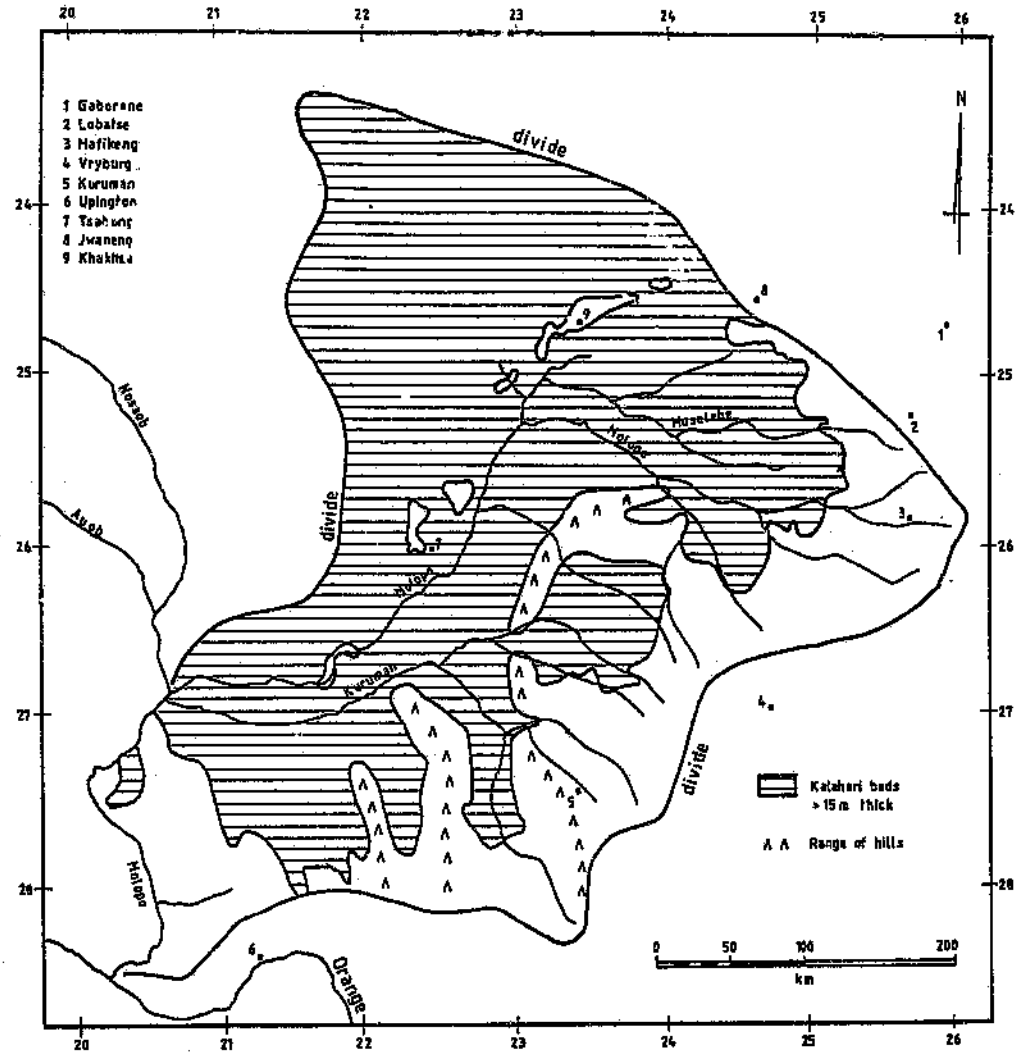


Figure 2.14 Distribution of Kalahari Beds (> 15 m) in the Molopo drainage area.

reconstruction of the evolution of the Molopo drainage and a detailed discussion is included in section 4.2.1.

2.1.3 Regional structural influences

Tectonic activity has played a crucial role in the deposition and development of the Kalahari Group sediments according to Thomas and Shaw (1990). They identified three major aspects of tectonic activity: the establishment of the continental setting for deposition, contributions to the drainage evolution of the area, and the creation of local conditions that favour thick sedimentation. Below follows a brief discussion of basement structures, Karoo structures, the Morokweng meteorite impact structure, and neotectonics. The possibility of resurgence of earlier structures ' -ing the later geological history of the area has been suggested (Green, 1966).

Basement structures

The basement of the drainage area is made up of the pre-Karoo rocks of the Basement Complex and the late Archaeozoic and Proterozoic. The southern African subcontinent comprises several early Precambrian to Cambrian structural provinces (Tankard *et al.*, 1982). The use of the non-genetic term 'province' rather than the concepts of 'cratons' and 'mobile belts' has been suggested, since some cratons include rocks that have the structural and metamorphic characteristics of 'mobile belt' terranes. Figure 2.15 shows the tectonic provinces as indicated by Pretorius (1985). The major part of the Molopo drainage occurs within the Kaapvaal structural province, with only the final downstream part within the much smaller Gordonia structural province. The north-south Tshane structural hinge-line is situated near to the contact between the Kaapvaal and Gordonia provinces. Pretorius (1985) further suggested that the Rhodesia and Kaapvaal stable blocks are the two core components of the Southern African Craton, and that the key to understanding the configuration of the Southern African Craton lies beneath the Kalahari of Botswana.

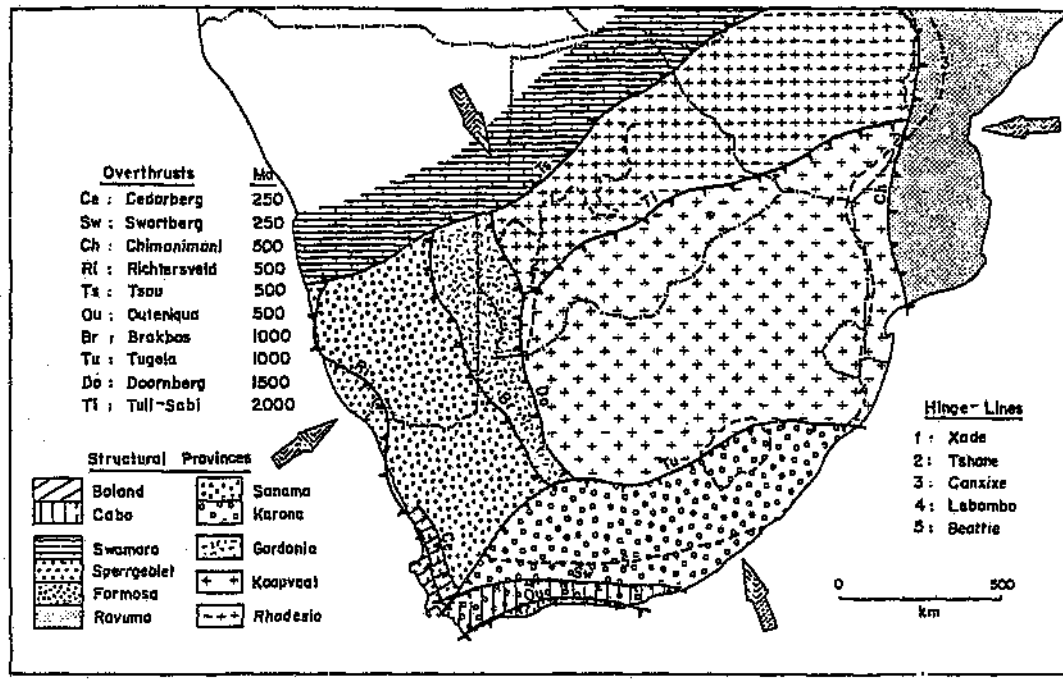


Figure 2.15 Tectonic provinces of southern Africa (after Pretorius, 1985).

A synopsis of basement features for Botswana (Fig. 2.4) has been provided by Meixner *et al.* (1985). Two major structural lines together divide the basement geology of Botswana: they are the Kalahari and Makgadikgadi Lines. The Archaean Basement, which is covered by Kalahari and Karoo sediments, is situated to the east of the structural lines, and is made up of the Rhodesian and Kaapvaal cratons. The Zoetfontein Fault separates the two cratons. To the west of the Kalahari-Makgadikgadi Lines lies the Medial Rift and the Ghanzi-Chobe Foldbelt. In the Medial Rift the basement appears highly magnetic and deeply buried. A thickness of up to 15 km of non-magnetic cover in three basins associated with the rift have been indicated by Meixner *et al.* (1985): they are the Passarge, Ncojane and Nossob Basins. The cover is probably of Nama and Karoo age, with possibly even Precambrian sediments in the Passarge Basin. Folding and faulting patterns in the area of the Molopo Farms Complex have been described by Gould *et al.* (1987). Major pre-Waterberg folding occurred along E-W axes, and faults inferred from LANDSAT imagery show a general NW-SE as well as some other fault directions. The Cargonian Highlands (Visser, 1987) which may have formed in association with the tectonism of the late Precambrian-early Cambrian Pan African Event were a major structure, which strongly influenced sedimentation of the Karoo Supergroup.

Karoo structures

The Karoo Supergroup within Botswana has been affected only by epeirogenic movements and not by major fold-structures (Green, 1966). Two main phases of epeirogenic movement have been indicated: an earlier one between the Stormberg Series and older strata of the Karoo Supergroup, and a second, post-Stormberg phase, which produced the post-Karoo structural features. The earlier phase of epeirogenic movement consisted mainly of regional uplift accompanied by broad warping (Green, 1966). The post-Stormberg movements resulted in a large number of faults and possibly synclinal warping south of the Makgadikgadi. All the faults appear to be normal faults of tensional origin. Green (1966) further suggested that the fault-pattern may have been controlled by pre-existing structural trends and could represent 'resurgent tectonics'. He further mentioned that in eastern Botswana the post-Karoo structural trends already existed, at

least, as long ago as the post-Transvaal and post-Waterberg structural epoch. Migratory rift faulting has also been a controlling force of Karoo sedimentation and volcanism (Reeves, 1973).

Morokweng meteorite impact structure

A large almost circular subsurface structure near Morokweng (centred at 26^0 31' S and 23^0 32' E) has been inferred on the basis of gravity and magnetic investigations (Fig. 2.16) and has recently received much attention (e.g. Corner *et al.*, 1997; Hart *et al.*, 1997; Hecht, 1997; Köberl, Armstrong and Reimold, 1997; Köberl, Reimold, Armstrong, Brandstätter and Kruger, 1997; McKinnon, 1997). The Morokweng meteorite impact structure beneath the Kalahari Sands was first recognised in 1994 (Corner, 1994) and later confirmed by the discovery of shocked rocks (Andreoli *et al.*, 1995). The impact structure appears as a 70 km diameter feature on the regional aeromagnetic map of southern Africa and Hart *et al.* (1997) suggested that the diameter of the total structure may be as large as 340 km (Fig. 2.17). It has been proposed that the 70 km diameter inner structure (Fig. 2.18) would represent the central uplift area, which would most likely have had a peak-ring structure (Corner *et al.*, 1997). The evidence for an impact origin of the Morokweng structure is based on the following points: (1) the circular geophysical images, (2) anomalous Ir concentrations in putative impact melt rock, (3) the identification of planar deformation features (PDFs), and (4) other shock deformation phenomena in surface pebbles and rocks recovered from the boreholes (Hart *et al.*, 1997). Both Corner *et al.* (1997) and Hart *et al.* (1997) give a date of about 145 ±2 Ma for the crystallisation of the impact melt. Circumstantial evidence on the linking of the Morokweng impact to the break-up of Gondwana and causal relationships with J-K extinctions (Köberl *et al.*, 1997a) have also been advanced.

In order to consider the influence of this potentially, large impact structure on the evolution of the drainage lines in the area, the analogy of the Vredefort Impact Structure is discussed briefly. This impact structure is situated about 390 km to the east of the Morokweng impact structure and about 110 km south-west of Johannesburg. The

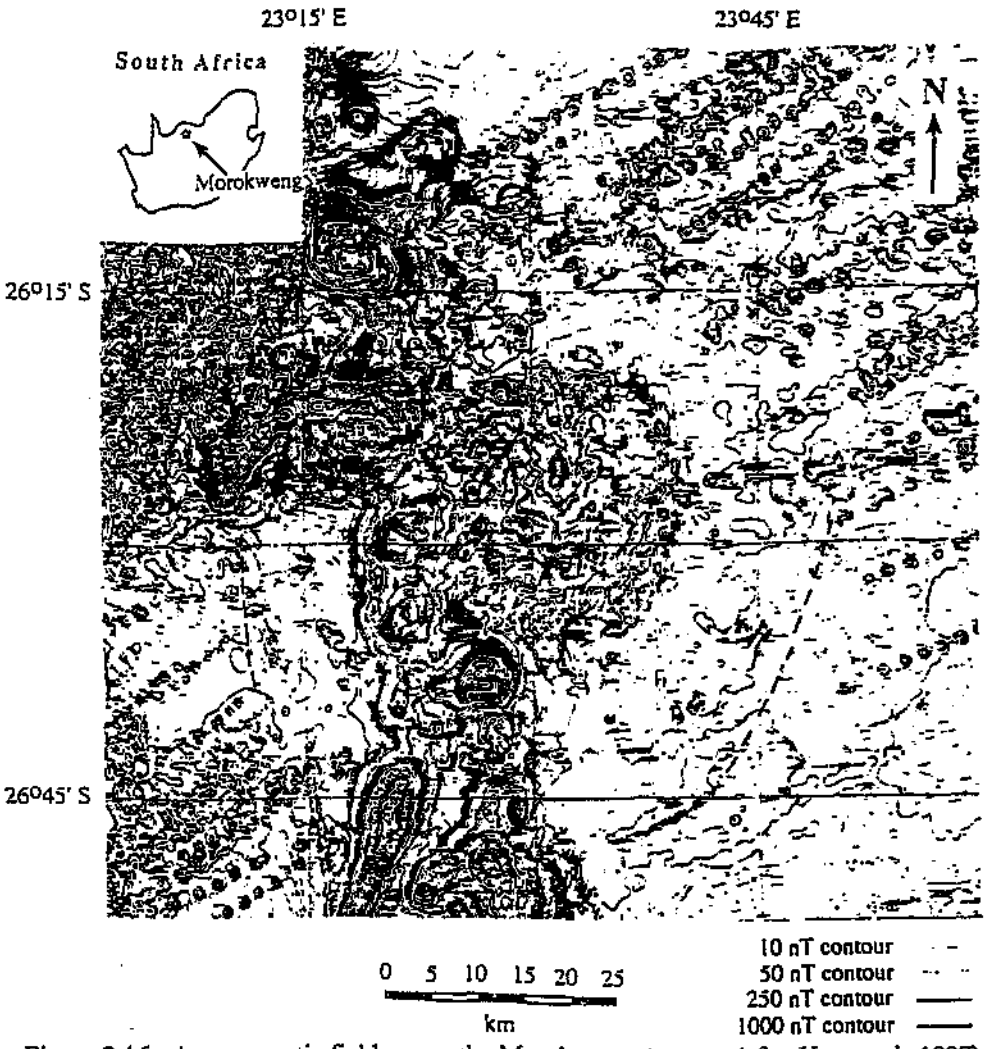


Figure 2.16 Aeromagnetic field across the Morokweng structure (after Hart *et al.*, 1997). Dashed line denotes the outer margin of the weakly magnetised halo.

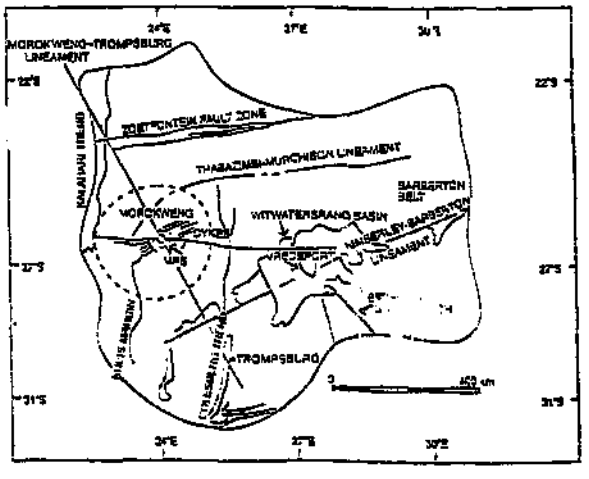


Figure 2.17 Kaapvaal province showing the Morokweng impact structure (after Corner *et al.*, 1997). MRS = Morokweng Ring Structure.

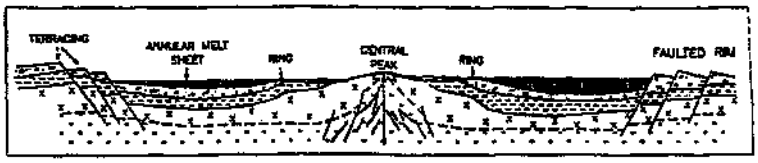


Figure 2.18 Schematic cross section through a complex impact structure (after Reimold and Gibson, 1996). Erosion is indicated by the lack of impact melt and/or fallback breccia on top of the central uplift region.

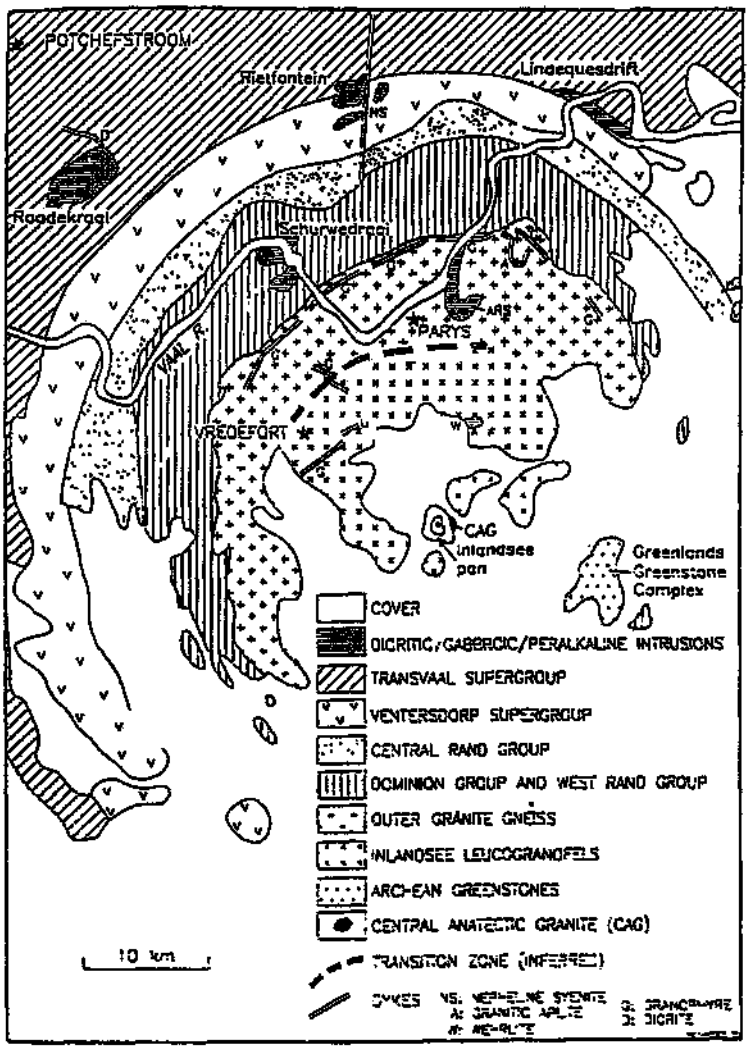


Figure 2.19 Schematic geological map of the Vredefort Dome (after Reimold and Gibson, 1996).

Vredefort impact event is much older and dated at 2023 ± 4 Ma (Reimold and Gibson, 1996). Reimold and Gibson (1996) further described the structure as having a 50 km-wide core, surrounded by a 15-20 km-wide collar of up and over turned supracrustal strata. The southern part of the collar is not closed but the remaining part is extensively faulted. They further stated that the formation of the Vredefort structure caused significant deformation at a radial distance of 120 km from the dome. The central impact melt body has subsequently been eroded. The Vredefort structure has just like the Morokweng structure a westward tilt (Reimold, pers. comm., 1997). The course of the Vaal River is partially guided by the ring structure, as well as by tangential faults, of the Vredefort impact structure (Figure 2.19). It is suggested here that the Morokweng impact structure will have similarly affected the drainage lines, within a large radius of the point of impact in the Molopo drainage area.

Neotectonics

The term neotectonic refers to post-Gondwana tectonic activities. The influence of the intra-continental Griqualand Transvaal axis has generally been accepted as an important factor in the development of the drainage lines in central southern Africa (e.g. du Toit, 1933; Partridge and Brink, 1967; Mayer, 1978; McCarthy, 1983; Partridge and Maud, 1987). Thomas and Shaw (1991) also stressed that tectonics have played a major role in the development of the Kalahari Basin and its drainage systems. They suggested that an endoreic drainage system, of which the Okavango is the only remaining major river from this system, developed from the marginal continental hingeline to the interior due to isostatic uplift of that hingeline. Thomas and Shaw (1990) emphasised the concept of endoreic drainage as one of the few feasible mechanisms which could account for the initial deposition of the Kalahari Group sediments. They further stressed that the drainage evolution since the late Jurassic, has been dominated by the progressive capture of the endoreic systems by surrounding, more aggressive, exoreic systems, for which they cited evidence from the Cunene River in Angola, the Fish River in Namibia, the Orange-Vaal in South Africa and the section of the Zambezi system in Botswana. They further referred to the role of the Pliocene uplift of the Kalahari-Zimbabwe axis in the drainage

developments of the Zambezi drainage. Thomas and Shaw (1990) thought it likely that by the late Pleistocene the middle Kalahari had been deprived of its fluvial inputs. They also referred to local neotectonic activity during and after deposition. The fault-aligned Okavango Delta (e.g. Kunyere and Thamalakane faults) has been described as a slowly subsiding rift structure linked to the east African rift system (Shaw and Thomas, 1993). Post-Gondwana (neotectonic) subsidence in the rift has been estimated between 300 m (Greenwood and Carruthers, 1973) and 1000 m (Hutchins *et al.*, 1976) as cited by Thomas and Shaw (1991). Truncation of a linear dune system adjacent to the western Okavango and the activity of distributary streams have also been associated with recent neotectonic activity (Shaw and Thomas, 1993). A major change in the hydrological regime of the lower Okavango in 1952 has been attributed to earthquake activity (Hutchinson and Midgeley, 1973). Reeves (1973) determined a 'Kalahari Seismicity (the 'regression' line in Fig. 2.20) for Botswana on the basis of statistical analysis of twenty Kalahari earthquake epicentres. This seismicity axis has a familiar NE-SW trend through the centre of Botswana, and closely parallels the linear features of the Okavango Delta, The Ghanzi Ridge, and the Makgadikgadi Salt Pans. Reeves further suggested a relation between the seismicity axis and regional subsidence, as the lowest area of the Kalahari (the Makgadikgadi Basin) lies over the axis, and also the axis of the dry Okwa/Mmone drainage has a distinct north-east strike. An association of the Kalahari seismicity axis with the possible influence of earlier rifting on Karoo sedimentation and volcanism was also suggested by Reeves (1973).

2.1.4 Climate, soils and vegetation of the Molopo Basin

Climatic characteristics

The climate of the Molopo drainage area has been characterised as semi-arid to arid (Poynton, 1971; Schulze and McGee, 1978). A similar picture emerged according to the classification of arid environments by Meigs (1953) as indicated by Thomas and Shaw (1991). The north-eastern section is characterised as semi-arid while the south-western

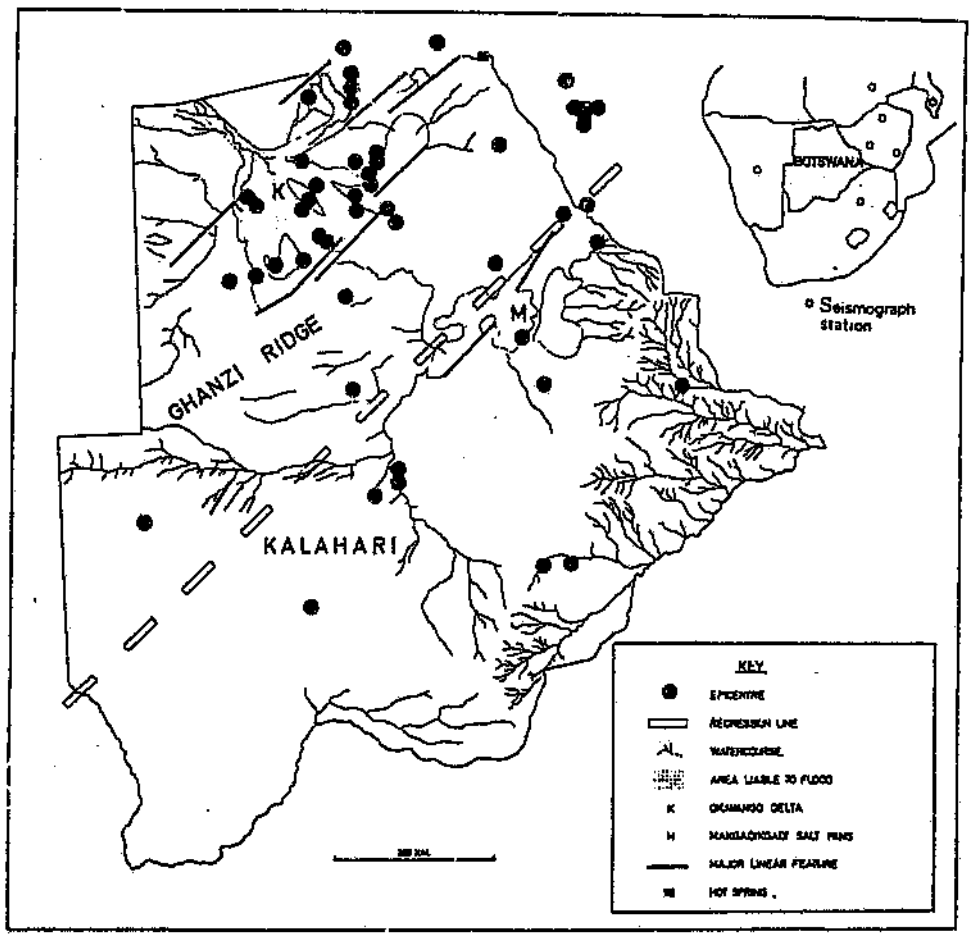


Figure 2.20 Earthquake centres in Botswana, September 1965 to August 1971 (after Reeves, 1973).

part is more arid. The annual precipitation varies from more than 500 mm in the north east to less than 200 mm in the south-west of the drainage area. There is a strong seasonality in the rainfall. For most of the Kalahari region on average over 80 per cent of the annual rainfall occurs between October and April (Thomas and Shaw, 1991). The annual potential evapotranspiration exceeds by far the annual precipitation (Makara, 1987). The average annual water deficiency, which according to Carter (1954) indicates the moisture required by vegetation, that soil storage and precipitation have failed to meet, varies from less than 600 mm in the north-east to more than 800 mm in the south-west (Schulze and McGee, 1978).

The mean annual temperature within the study area ranges from 22^0 C in the northern section to 18^0 C in the southern section. The mean annual temperature range varies between 12^0 C in the north-east to 16^0 C in the south-west. The summer (January) mean daily temperatures ranges between 24^0 and 26^0 C, with a mean monthly maximum between 30^0 and more than 35^0 C. The winter (July) mean daily temperature is around 10^0 C, with a mean monthly minimum between 0^0 and 2.5^0 C (Schulze and McGee, 1978).

Soils

The sandy soils of the central and southern Kalahari have an extreme low fertility (Thomas and Shaw, 1991). Compared with the sandy soils of the of central Australia, the dunefield soils of the southern Kalahari are at least three times as deficient in phosphorous and two to ten times as deficient in extractable calcium (Buckley *et al.*, 1987; Thomas and Shaw, 1991). Where the soil cover is thin and the pre-Kalahari bedrock protrudes the soil, characteristics are modified according to the bedrock. Most of the sandy soils have poor profile development, are moderately acid, and are categorised as arenosols in the FAO-UNESCO classification used in southern Africa (Thomas and Shaw, 1991).

On the schematic soil map of southern Africa (von M. Harmse, 1978) almost the entire Molopo drainage area (Fig. 2.21), except for the south-eastern rim section, is indicated as dominated by mainly yellow and red arenosols (6c and 6b). The arenosols are generally red in upland sites and on dune ridges, yellow in level areas, and grey in bottomland areas (von M. Harmse, 1978). The dry river beds of the Molopo and Kuruman have compact sandy clays or silts (Werger, 1978). Lithosols (Oa) can be found on the rim area, where topography is the dominant soil forming factor. The soils are shallow (30 cm and less), with weak profile differentiation and coarse fragments. Lithosols developed on dolomite bedrock have been indicated as calcrete and calcareous rocks (Oe). Only a very small area south of Mafikeng has some more fertile fersiallitic yellow and grey sands and loams (2c) with incipient soil horizons, weakly to moderately developed structures, and mainly black montmorillonitic clays (3a) with a varying degree of argillo-pedoturbation have formed on basic (Ventersdorp Lava) rocks (von M. Harmse, 1978).

Vegetation

Much of the vegetation of the Kalahari is commonly described as savanna, with the most extreme south-western Kalahari being a desert grassland (Thomas and Shaw, 1991). The vegetation of most of the Molopo drainage area ranges from a bush savanna in the north to a more arid shrub savanna in the south-west (Weare and Yalala, 1971). The shrub savanna occurs in conjunction with the dunefield, and has a groundcover of clumped grasses of often less than 35 per cent (Thomas, 1988) and even some bare areas. Plant densities and diversity generally increase in a north-easterly direction (Thomas and Shaw, 1991). The vegetation along the dry Molopo River valley upstream from about Watersend Farm has been described as 'Molopo Thornveld' (Weare and Yalala, 1971). The dry valley generally comprises the same trees as neighbouring areas, but they attain greater heights within a kilometre or so from the valley bottom (Thomas and Shaw, 1991). The more humid, most north-easterly area of the Molopo drainage, has been described as supporting arid sweet bushveld (Weare and Yalala, 1971). The eastern boundary of the southern Kalahari is, according to Werger (1978), situated somewhat near Kuruman, although the characteristic Camelthorns (*Acacia erioloba*) of the Kalahari still occur

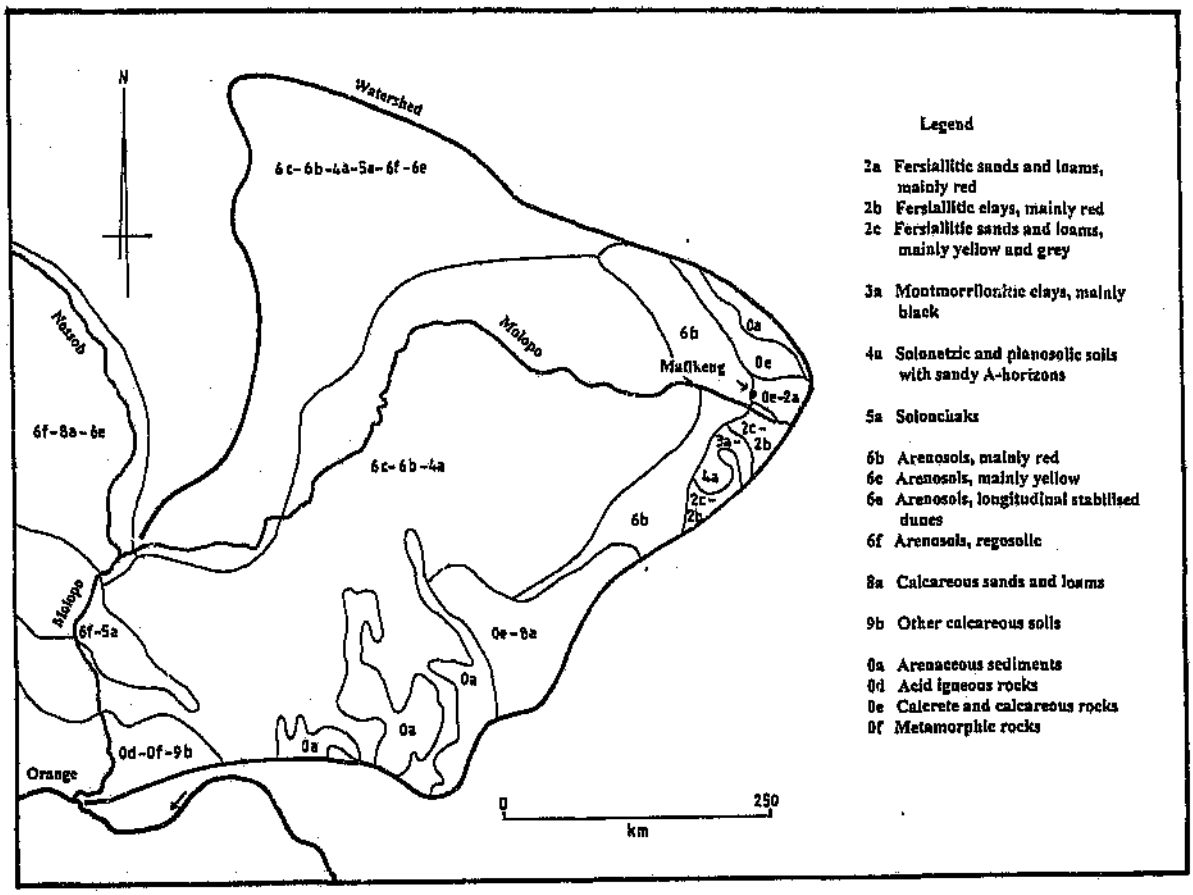


Figure 2.21 Schematic soil map of the Molopo drainage area (based on von M. Harmse, 1978).

between Kuruman and Vryburg on the Ghaap Plateau on deep sandy patches, where the surrounding calcrete substrate generally supports a grassland. Zambezian and Karoo floral elements generally become more dominant towards the south-easterly boundary and in the mountainous areas like the Langeberge and Asbestos Hills.

2.2 Topography of the Molopo Drainage Basin

The area of the Molopo drainage basin is generally characterised by a lack of relief. Most of the sediment-filled Kalahari Basin is extremely flat terrain with wide, shallow, poorly defined dry river valleys, and dunes in the south-western dunefield with a wave length of about 250 m and amplitudes of not more than 15 m. The only substantial relief features are the ranges of hills on the south-eastern rim of the Kalahari Basin.

2.2.1 Areal extent and description of divides

As outlined in section 1.2, the Molopo drainage comprises the Molopo, Moselebe and Kuruman Rivers and their tributaries. The entire basin covers an area of approximately 193 000 km^2 and extends across the international border between Botswana and South Africa. The north-western section of the Molopo drainage area is situated in the south of Botswana, and covers parts of the Kweneng, Southern, and Kgalakgadi districts. In South Africa the drainage basin is situated in the western section of the North West Province as well as the northern section of the Northern Cape Province. This section of the drainage area is covered by the following 1:250 000 topographical maps: 2522 Bray, 2524 Mafikeng, 2621 Twee Rivieren, 2622 Morokweng, 2624 Vryburg, 2720 Noenieput, and 2722 Kuruman. The integrated drainage area has a general NNE-SSW slope, from just over 1100 m AMSL at the Bakalahari *Schwelle* in the north, to about 850 m AMSL in the area near to the Orange/Molopo confluence in the south, over a straight distance of about 600 km (0.42 m per km). Except for the southern Auob/Nossob tributary system, the Moselebe is the only substantial right-bank dry valley system which joins the Molopo River in its northern section. The more than 400 km distance between the Moselebe/Molopo and Nossob/Molopo confluences, is except for the minor dry Shabe Valley, generally characterised by an absence of right-bank tributaries originating from

the arid, sediment-filled Kalahari Basin. Both the Molopo and Kuruman Rivers have several well-developed left-bank tributaries, of which some may have episodic floods during above-average rainfall years (section 1.2.1).

The areal extent of the drainage area is bounded by vague divides between the Nossob and the Molopo River in the west, the Bakalahari *Schwelle* in the north, and the generally better defined eastern divide between the Molopo drainage and the Harts, Dry Harts, Vaal, and Orange Rivers (Fig. 1.2). The potential divide between the Nossob and Molopo Rivers is an almost indistinguishable rise about halfway between the two drainage lines. Uncertainties about recharge and groundwater flows in that area makes this divide truly a potential divide. The Bakalahari *Schwelle* is slightly better defined, and where the divide is situated on outcrops of the Basement rocks (between Jwaneng and Lobatse) the northern divide is generally well defined. Where the eastern divide does not lie across the flat Karst terranes of the Malmani and Ghaap Plateau Dolomites, it is generally well-defined.

2.2.2 Description of the Molopo valley system, drainage patterns and thalweg

Molopo valley descriptions

The Molopo River valley has various forms between its origin at the Molopo Eye, and its confluence with the Orange River. The section between the Molopo Eye and Mafikeng on the Malmani Dolomites is characterised by a very shallow (less than 5 m) valley with a width of about 750 m. Beyond Mafikeng up to Phitshane/Makgobistad (100 km downstream from Molopo Eye), where several tributaries join the Molopo River, the valley is incised in granitic-gneissic rocks and has a depth of about 20 m and a width of over 1.5 km, and has two remnant rock-cut terrace levels at 6 m and 14 m (Plate 2.1) as mentioned by Goudie (1973). About 8 km west of Phitshane the Molopo River has cut an almost 30 m deep gap (*poort* in Afrikaans) into banded ironstone rocks of the Kraaipan Group (Wellington, 1955). The floor of the gap was further described by Wellington as about 50 m wide, flat and bush-covered, with a stream channel about 2.5 m wide. From



Plate 2.1 Rock-cut terrace remnant with calcretised sediment along the upper Molopo west of Mafikeng. Water is accumulating (late 1994) in the new Modimola dam.

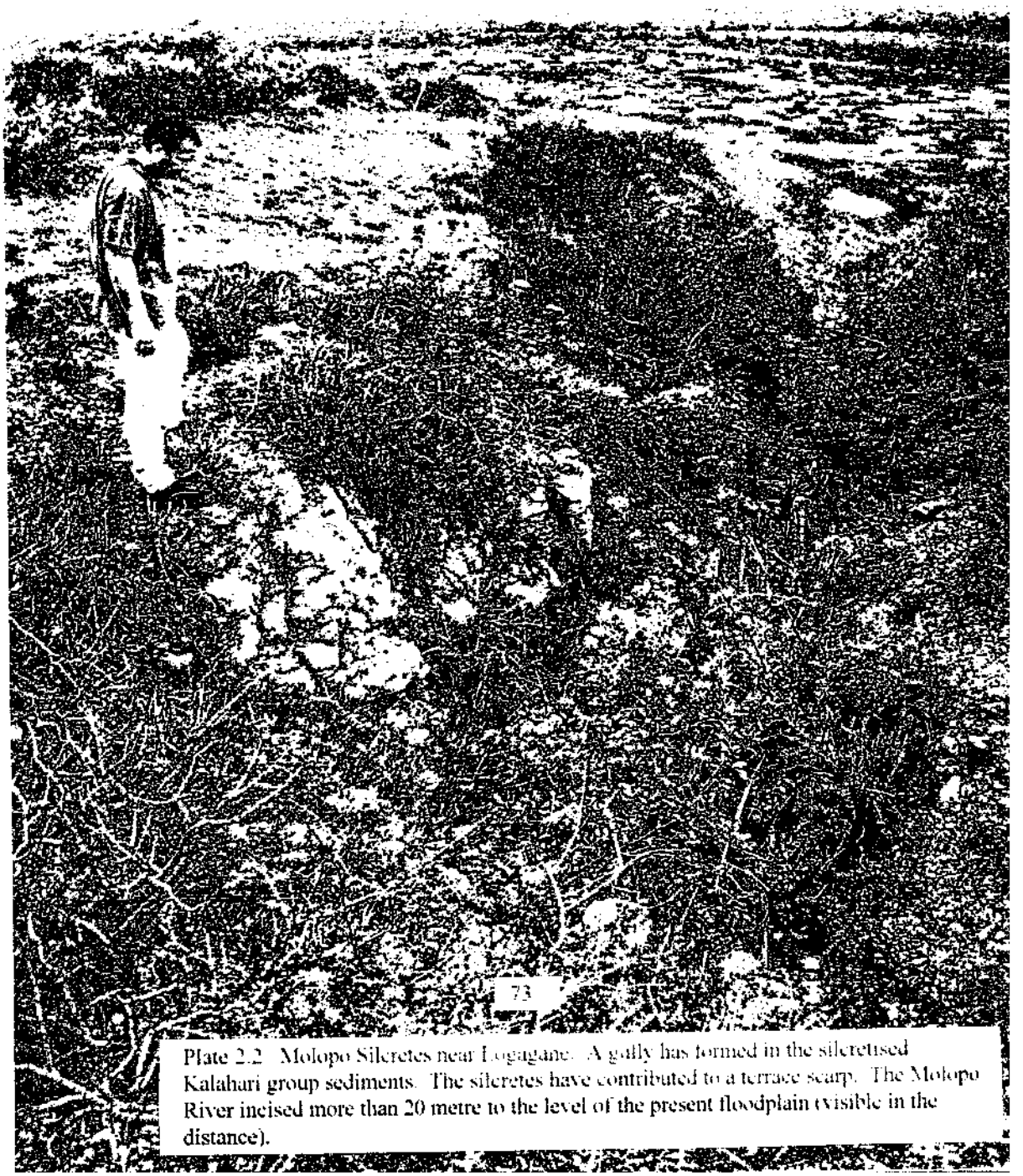


Plate 2.2 Molopo Silcretes near Logagane. A gully has formed in the silcretised Kalahari group sediments. The silcretes have contributed to a terrace scarp. The Molopo River incised more than 20 metre to the level of the present floodplain (visible in the distance).

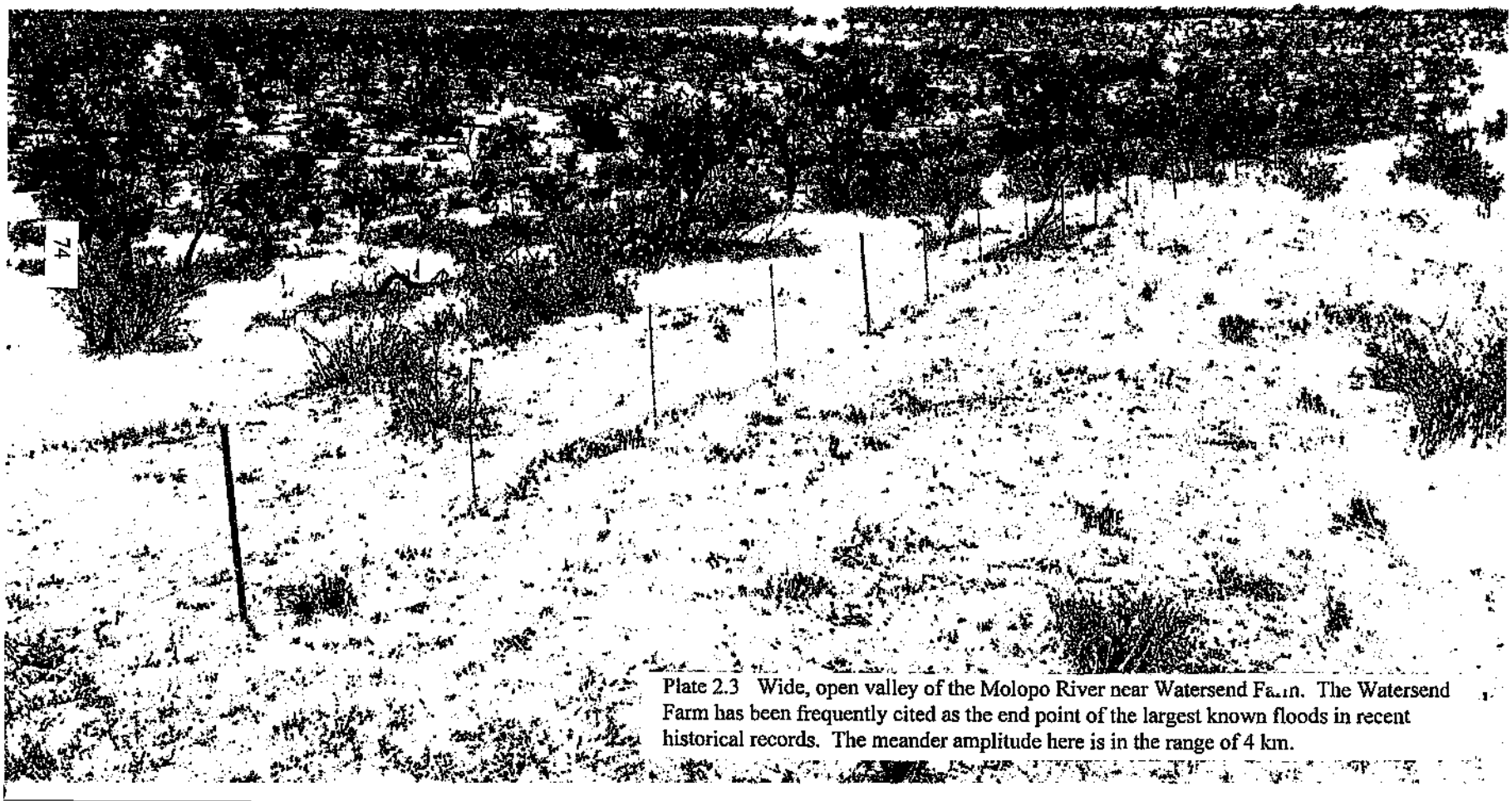


Plate 2.3 Wide, open valley of the Molopo River near Watersend Farm. The Watersend Farm has been frequently cited as the end point of the largest known floods in recent historical records. The meander amplitude here is in the range of 4 km.

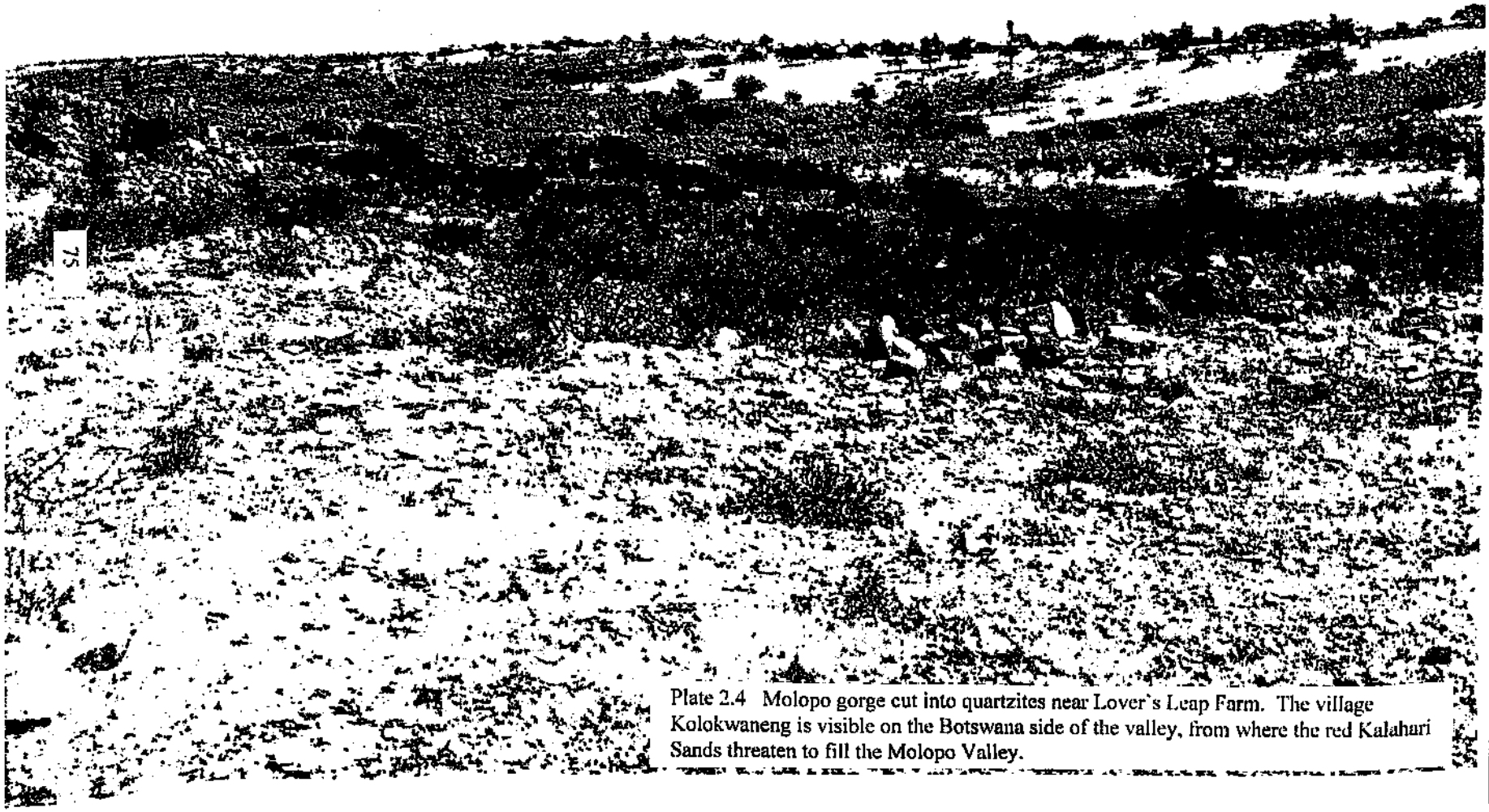


Plate 2.4 Molopo gorge cut into quartzites near Lover's Leap Farm. The village Kolokwaneng is visible on the Botswana side of the valley, from where the red Kalahari Sands threaten to fill the Molopo Valley.

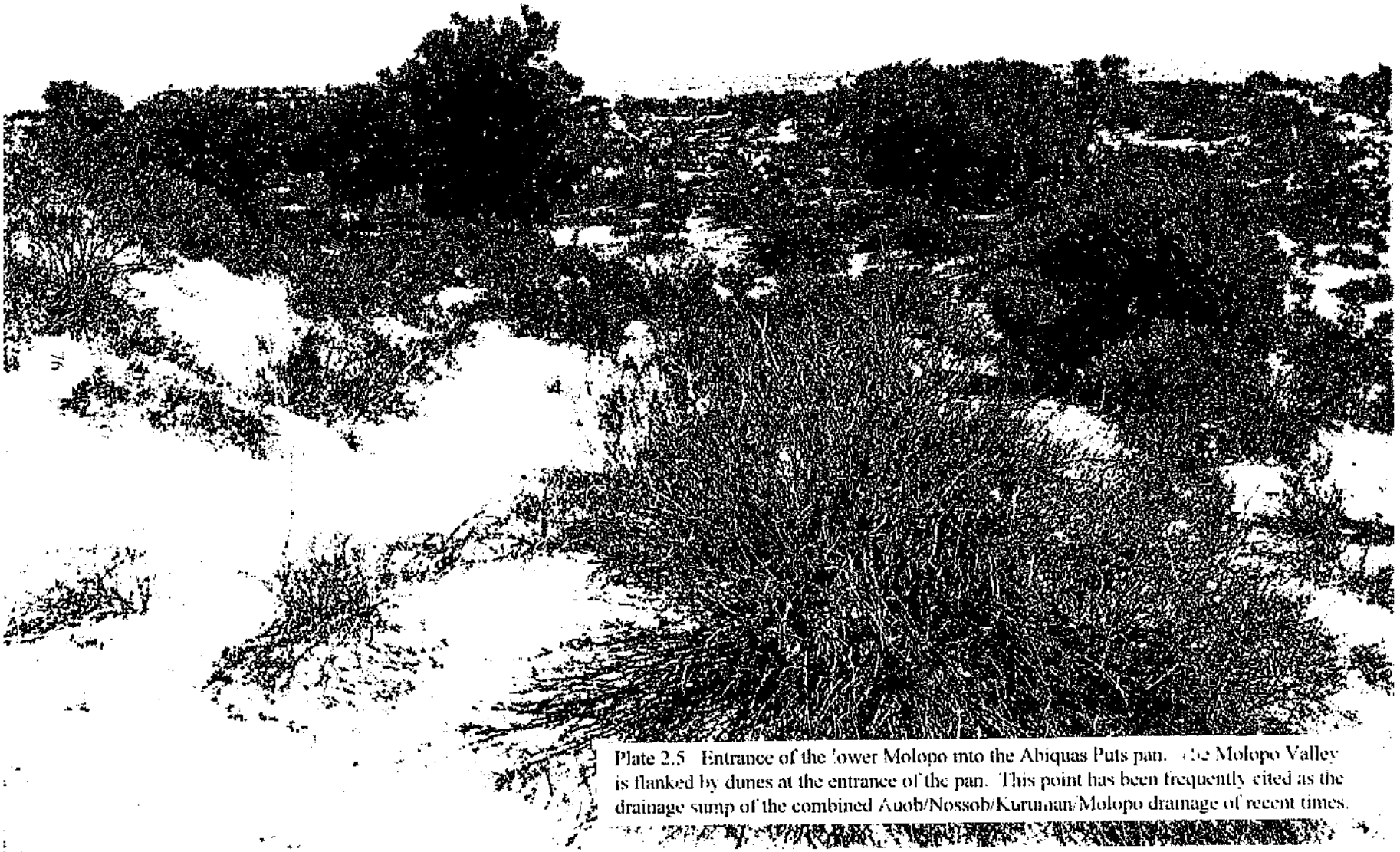


Plate 2.5 Entrance of the lower Molopo into the Abiquas Puts pan. The Molopo Valley is flanked by dunes at the entrance of the pan. This point has been frequently cited as the drainage sump of the combined Auob/Nossob/Kuruman/Molopo drainage of recent times.

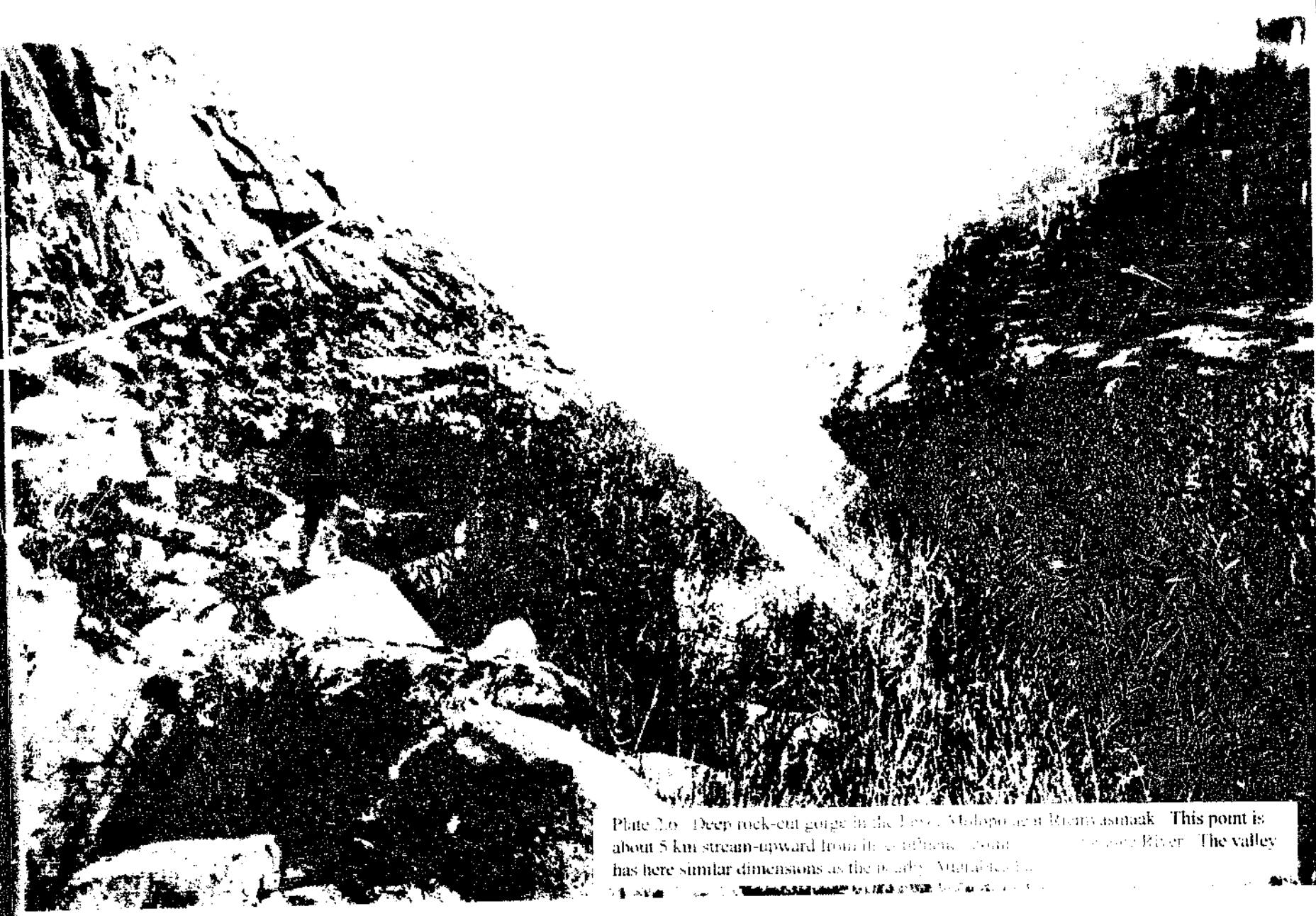


Plate 2.6 Deep rock-cut gorge in the [illegible] Molopo near Riemvasmaak. This point is about 5 km stream-upward from its confluence [illegible] River. The valley has here similar dimensions as the nearby [illegible]



Plate 2.7 Molopo/Orange confluence area. The Orange River flows [illegible]. Irrigated vineyards are visible along the Orange River.

Phitshane to Moreson Farm (390 km downstream) the Molopo River is incised up to 15 m into sediments of the Kalahari Group. Where the sediments have been cemented by duricrusts (e.g. 'Molopo Silcretes', as mentioned by Michaluck and Moen, 1991) the valley sides can be relatively steep (Plate 2.2) and sometimes gorge-like (e.g. between Thsidilamolomo and Mabule). The section between Moreson Farm and Watersend Farm (455 km downstream) has been described by Nash (1992), as being 150 m wide and having a relief of less than 6 m, and having some pans with evidence of long-standing water (Plate 2.3). The valley from McCarthy's Rust (565 km downstream) towards Khuis (708 km downstream) was further described by Nash as having a width of around 250-300 m and a total depth of 10-12 m, as well as having sections with extensive outcrops of grey crystalline *terrazo* silcretes (Summerfield, 1982). Where the Molopo River valley is cut into heavily calcretised quartzites and Dwyka Tillites between Lover's Leap Farm (695 km downstream) and Middelputs/Khuis, its valley is well-defined and forms a gorge (Plate 2.4). The channel was described to be nearly 30 m deep, where cut into quartzites (Wellington, 1955). Beyond Khuis, where the Molopo Valley is incised into the sediments of the Kalahari Group the valley becomes increasingly shallow, until it becomes ill-defined beyond Abiquas Puts (995 km downstream) and is blocked by sand dunes (Plate 2.5).

Beyond Swartmodder Farm (1024 km downstream) the lower-Molopo becomes more deeply incised into Dwyka Tillites, quartzites, and granitic bedrock to form a gorge, nearly 60 m deep and less than 40 m wide near Riemvasmaak (1126 km downstream), before it joins the Orange River downstream from the Augrabies Falls. An imaginative description of the last 50 km of the lower Molopo from Swartmodder to the Orange confluence has been provided by Haughton (1927). He characterised the section below Riemvasmaak as having a valley of variable grade, pieces of river-cut terraces, two distinct (dry) waterfalls, a stretch of rapids, a narrow gorge section with high perpendicular sides (Plate 2.6), and a final low-grade sandy stretch bordered by low hills up to the Orange River (Plate 2.7). Lateral tributaries near the gorge section have hanging valleys where they meet the Molopo. He suggested that the gravel terraces of the

lower Molopo antedate the recession of the Molopo Falls to their present position, and that they indicate an uplift of the area of about 15 m. Haughton (1927) further indicated that the dry Molopo Falls near Riemvasmaak are related to the Augrabies Falls of the Orange, and that they are an indication that the Molopo River had still sufficient erosive power for the waterfall to migrate headwards 12 km, while the Orange River cut its waterfall back by about 17 km. Haughton estimated that the active erosion and cutting-back of the Molopo Falls extended into times more recent than late Cretaceous or early Tertiary.

The Valley of the Kuruman River, where cut into the sediments of the Kalahari Group, is like that of the Molopo River, with similar wide and shallow characteristics, but generally of smaller dimensions. The Kuruman River valley near Groot Drink Farm ($26^0 55' 16''$S, $22^0 44' 33''$E) was described by Nash (1992) as having an average depth of 10-15 m beneath the surrounding Kalahari Sand surface, with two sedimentary terrace levels at respectively 2.5 to 3.0 m and 7.5 to 8.0 m above the valley floor. Near Bella Vista Farm ($26^0 42' 26''$S, $22^0 26' 49''$E) Nash further described the valley dimensions as having a 30-38 m wide channel and a 10 m depth, as influenced by flood scouring in 1988. The influence of episodic recent flooding (e.g. 1988) on the Kuruman drainage system was discussed by Shaw *et al.* (1992), and they suggested a correlation to the high-phase events of the Southern Oscillation.

The Moselebe system has generally even shallower dimensions than the Kuruman River. In its upstream section, where the Moselebe crosses the Ramatlhabama-Lobatse road, it was described by Nash (1992) as only a shallow trough of 5 m depth with a poorly defined channel less than 10 m across,. Further downstream near Mmathethe Nash described the valley as 700-800 m wide, reaching a width of 1.5 km and a depth of over 10 m, incised into the sediments of the Kalahari Group. A sediment terrace level of 2.5 m above the valley base, with a 90 m wide channel was further identified. Yet further downstream near the Bray-Khakhea road the Moselebe was described as having a depth

of 7-8 m and a width in excess of 1 km, with a sedimentary terrace level at 3.5 to 4.0 m above the valley base and a 200 m wide channel.

Drainage patterns

The Molopo drainage displays a variety of drainage patterns and evidence of channel metamorphosis. Sub-parallel patterns occur between the Moselebe and the upper Molopo over a stretch of approximately 300 km, as well as between the Kuruman and the middle section of the Molopo over about 400 km. A similar trend can be observed between the dry valleys of the Phepene, Kgokgole, Moshaweng, Matlhwaring tributaries, and the upper section of the Kuruman River. Large valley meanders occur in the middle section of the Molopo River between Phitshane and Khuis where the slope of the thalweg is very gentle (approximately 0.33 m per km). The amplitudes and wavelengths of the valley meanders are in the range of 4 km. Evidence of meander cut-offs, as well as down-stream meander shifting can also be found in this section. A meander-in-meander morphology occurs where small-size channel meanders, with an amplitude and wavelength of about 50 to 100 m, have formed in a sandy-clay back-fill in the valley of the upper Molopo. A dendritic pattern is dominant where the Basement rocks (granites, gneisses) have only a thin (< 15 m) sand cover.

The same pattern is visible within the Auob/Nossob system. Where the headwaters are located on the bedrock, or where the bedrock has only a thin sediment cover a dendritic drainage pattern has developed. Further downstream, on flatter terrain, the Nossob, Olifants, and Auob Rivers display a sub-parallel pattern over several hundreds of kilometres. The Nossob River also has similarly large valley meanders as the Molopo River in its middle section between Union's End and Twee Rivieren. The Nossob River must have received enough water at one stage to develop into a mature river, as it formed meanders and even some oxbow lakes (Thomas *et al.*, 1988). Similar patterns on the extremely low gradient alluvial plain of the Murray/Darling drainage system in Australia have been described by Williams *et al.* (1993). They commented on three major kinds of channels on the 'Riverine Plain', as studied by Schumm (1977) and Bowler (1978), and

discussed river metamorphosis between a presently highly meandering course, a much larger variety, and a straighter wide, low-sinuosity palaeochannel, associated with complex environmental changes. It is suggested that the Molopo drainage underwent similarly complex environmental changes.

Thalweg profiles

The length of the Molopo River thalweg between the Molopo Eye and the Molopo/Orange confluence exceeds 1135 km. The straight line NE-SW distance between the two points is about 630 km. The 500 km longer path of the thalweg of the Molopo River is due to its meandering nature and the influence of the Kalahari basining superimposed upon the general NNE-SSW slope of the Molopo drainage basin.

Table 2.2 General thalweg data of the Molopo, Moselebe and Kuruman Rivers. (based on 1 : 500 000 topographical maps)

Molopo River		Kuruman River		Moselebe River	
Altitude m.	Thalweg km.	Altitude m.	Thalweg km.	Altitude m.	Thalweg km.
1485	0	1450	0	1295	0
1400	17	1400	7	1250	3
1300	35	1300	28	1200	39
Mafikeng	40	1200	50	1150	79
1200	60	1100	74	1100	114
1100	150	1000	103	1050	173
1000	415	900	289	1000	291
900	720	Molopo confl.	387	Molopo confl.	307
800	1004				
700	1104				
600	1122				
500	1126				
~450	1135				
Orange confl.					

The Molopo River starts off in a generally east-west direction for the first 150 km, between the Molopo Eye and Tshidilamolomo. From Tshidilamolomo the meandering thalweg of the Molopo River continues in a generally north-westerly direction, towards the centre of the southern Kalahari, until the confluence point with the Moselebe River at

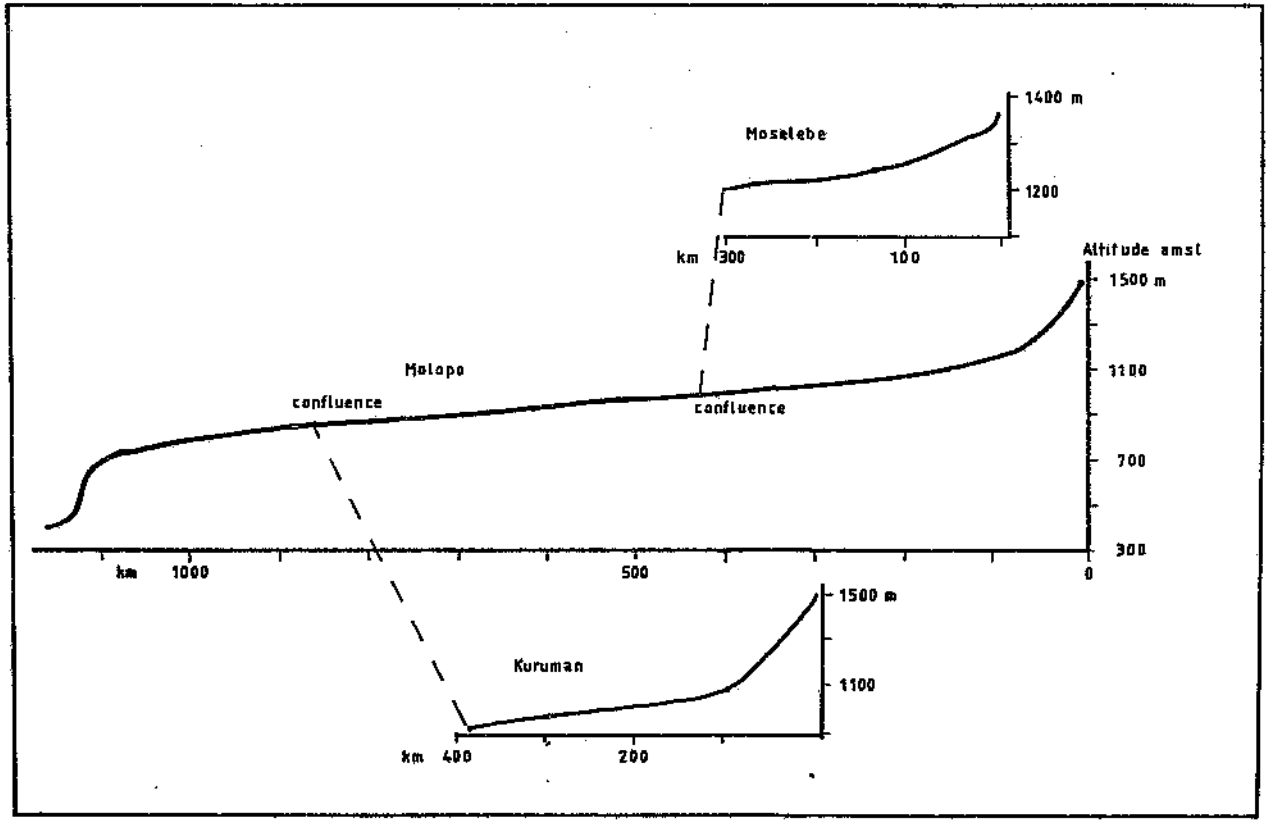


Figure 2. 22 Thalweg profiles of the Molopo, Moselebe, and Kuruman Rivers.

Makopong, just over 400 km downstream from the Molopo Eye. By way of a large bend, which already started just beyond Bray, the thalweg turns then into a southerly direction towards Khuis, about 708 km downstream from the Molopo Eye.

Beyond Khuis, the Molopo river turns again into a westerly direction until the confluence point with the Nossob River, about 845 km downstream from the Molopo Eye. From the Nossob confluence to the Orange confluence the thalweg continues with a large S-shape in a southerly direction. It is suggested here that the two westerly sections of the thalweg are due to Kalahari basining.

The initial slope of the thalweg over the first 35 km downstream from the Molopo Eye is relatively steep at about 5.3 m per km (Table 2.2). The thalweg profile then quickly becomes much less steep (Fig. 2.22) to reach its most gentle slope of about 0.33 m per km, not far from the halfway mark. The thalweg profiles of the Kuruman and Moselebe are very similar to the headward section of the Molopo River. Beyond the middle section the extremely gentle slopes of the Molopo thalweg slowly increase again to become rather abruptly very steep over the last 50 km or so, with a final slope of about 15 m per km over the last 12 km to the Orange River (based on data by Haughton, 1927). This final steep section represents a major knickpoint, which had migrated from the level of the Orange River just below the Augrabies Falls.

* * * * *

The physical background to the study of the Molopo drainage evolution has been reviewed in this chapter. The macro-scale geomorphology of southern Africa has been outlined, with specific emphasis on the drainage developments within the interior of the subcontinent. The development of the Kalahari Basin has been indicated as a major influence on the drainage evolution of the area. The regional geology and major regional structural influences have been outlined with the view that tectonics have played a major role in the history of the Molopo drainage. Resurgence of old geological structures has

been indicated as a potential factor. The recent studies of the Morokweng meteorite impact structure have revealed an added, important influence on the evolution of the drainage lines of the area. The general climatic characteristics, soils, vegetation, and ultimately the topography of the area were described in the final section of the chapter. The ancient Permo-Carboniferous drainage indicators, discussed in the next chapter, give evidence of very old fluvial drainage lines in this area, prior to the Cainozoic evolution of the Molopo drainage.

PART 2 DRAINAGE INDICATORS

CHAPTER 3 PERMO-CARBONIFEROUS DRAINAGE INDICATORS

3.1 Introduction

The aim in this chapter is to identify indicators to the earliest evidence of drainage lines within the present Molopo drainage area on the basis of a review and interpretation of existing data on the Permo-Carboniferous Dwyka glaciation. The indicators of the Permo-Carboniferous drainage evolution will later form the starting point in the discussion of the various stages in the development of the Molopo drainage (chapter 7). A rather well-developed picture of the palaeogeography during the pre-glacial and glacial stages has emerged from the work of Visser (e.g. 1983a, b; 1987; 1989).

The beginning of the glacial deposition in the Karoo Basin has been estimated to have occurred at the Stephanian-Westphalian stage boundary (296 Ma: Harland *et al.*, 1982) during the Upper Carboniferous period (Loock and Visser, 1985; Visser, 1987). The beginning of the glacial deposition had been preceded by an erosional period of about 30 Ma, which had occurred after the deposition of lower Palaeozoic sedimentary rocks in the southern Cape (e.g. Witteberg Group) at the end of the Visean (352-333 Ma: Harland *et al.*, 1982). The end of Permo-Carboniferous glaciation is estimated to have occurred towards the end of the Sakmarian stage (282-269 Ma: Harland *et al.*, 1989) during the early Permian period (Visser, 1987). The duration of the Permo-Carboniferous glaciation is thus estimated to have been about 20-30 Ma.

Some very ancient landscape features have been preserved through the infilling and protection of pre-Dwyka landscape features with sediments of the Karoo Sequence. In several places they have been exhumed through removal of the generally easily erodible Karoo sedimentary rocks. The case of the exhumed pre-Dwyka Dry Harts Valley ('Kaap Valley'), immediately to the east of the Molopo drainage, was discussed by du Toit (1910). Evidence for the antiquity of headwater sections of the Nossob some distance south of Gobabis, towards the west of the Molopo drainage was presented by Nash (1992), citing Hegenberger and Seeger (1980). Nash suggested that since the Permo-Carboniferous glaciation a valley has existed along the headwater section of the Nossob. There is evidence that within the Kalahari Basin some more pre-Dwyka landscape features are still filled with the Karoo sediments, and covered with Kalahari Group sediments. Extensive suboutcrops of sediments of the Karoo Sequence are known to occur beneath Kalahari Group sediments in the central area of the Molopo drainage (see geological maps 1:1000 000 RSA; 1:250 000, sheets 2622 Morokweng and 2522 Bray).

3.2 Dwyka Tillite Formation

The Dwyka Formation is the lowest lithostratigraphic unit of the Karoo Sequence (SACS, 1980) and is followed by the Ecca Group, Beaufort Group, Stormberg Group, and the Drakensberg Basalt Group. Deposition of the Karoo sediments has occurred in southern Africa in two major basins: the main Karoo Basin of South Africa and the Kalahari *(Karoo)* Basin which is centred on Botswana (Thomas and Shaw, 1991). The Kalahari (*Karoo*) Basin has been described as a fault-controlled, prominent valley system (Visser, 1989). This interpretation becomes obvious from the isopach map (Fig. 3.1) of the glacial deposits in the Kalahari Basin (Visser, 1987). The major NE-SW valley system has a maximum thickness of about 400 m of tillite. The isopach pattern indicates a prominent tributary system towards the south-east, which extends into the area of the Molopo drainage (Fig. 3.2). The occurrence of Dwyka Tillite within the Kalahari Basin has also been termed the 'Botswana Block' by Frakes and Crowell (1970). They further described the Botswana Block as a shallow half-graben that contains a Palaeozoic and Mesozoic

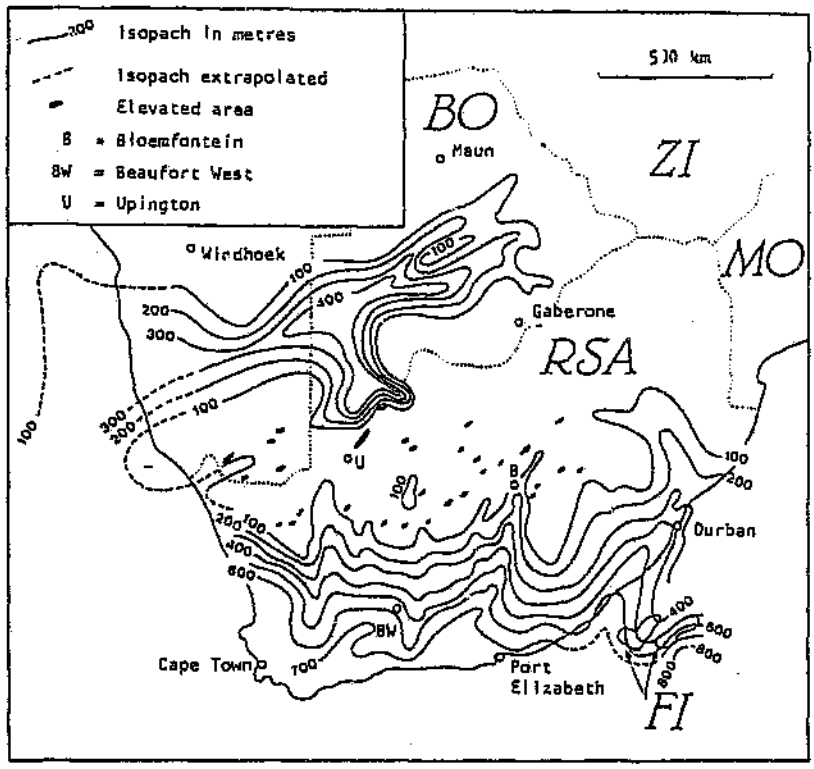


Figure 3.1 Isopach map of the glacial sequences in the Karoo and Kalahari Basins and on the Falkland Islands (after Visser, 1987).

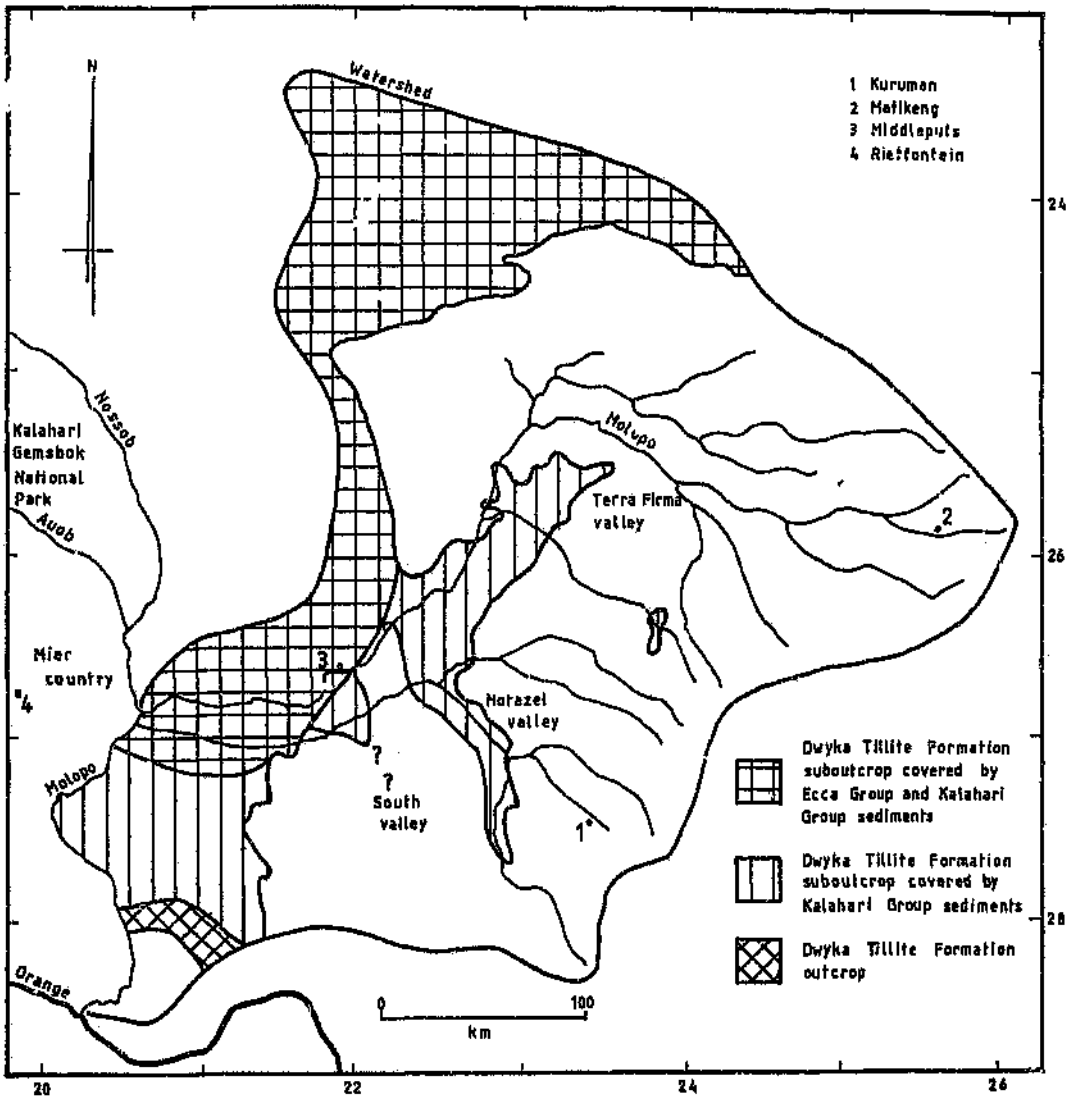


Figure 3.2 Dwyka Tillite Formation outcrops and suboutcrops in the Molopo drainage area. Based on the 1 : 1000 000 geological maps of South Africa and Botswana.

sedimentary sequence. Three sub-basins within the Kalahari (*Karoo*) Basin have been recognised (Green, 1966). These are the North-East Botswana Basin, the Central Kalahari Basin , and the South-West Botswana Basin. The Dwyka Tillite outcrops and suboutcrops in the Molopo drainage area form part of the South-West Botswana Basin.

3.2.1 Dwyka Tillite outcrops

Outcrops of Dwyka Tillite occur mainly near a major pan belt in the southern section of the Molopo drainage (Fig. 3.2). Outcrops in the relatively sand-free areas west of the lower Molopo River, and in and around many of the pans in the region (Plate 3.1) have been reported by Thomas and Thomas (1989). The provenance of the clasts in the tillite indicates a wide range of exotic rock types from the east and north-east. The clasts include basic vesicular lava, dolomite, banded ironstone, jasper, quartzite, red sandstone, granite and gneiss (Thomas and Thomas, 1989). Besides tillite the Dwyka also comprises blue-green shales and mudstones with some erratics. Sedimentary rocks of the Prince Albert Formation overlie the Dwyka Formation and are exposed in an area just north of Noenieput. The deposits of the Prince Albert Formation have been interpreted as representing an initially shallow-marine sequence with, lower in the sequence towards the south, varved sediments indicating glacial meltwater lakes (Thomas and Thomas, 1989). Towards the top of the formation the sediments are more arenaceous and indicate a rapid influx of clastic material and indicate the onset of fluvio-glacial conditions. Karoo dolerite dykes and sill-like bodies are further present in the Karoo Sequence.

Outcrops of the Dwyka and Prince Albert Formation have also been described to the north of the Noenieput region in the Rietfontein area, in the 'Mier country' west of the lower Nossob (Thomas *et al.*, 1988), as well as in the valley of the Molopo River from Katakura Farm almost continuously for about 30 km eastwards towards Middelputs. The sections in the valley of the Molopo River have been described by Rogers (1907), du Toit (1916), Boocock and Van Straten (1962), Gerrard (1965), Green (1966), and Frakes and

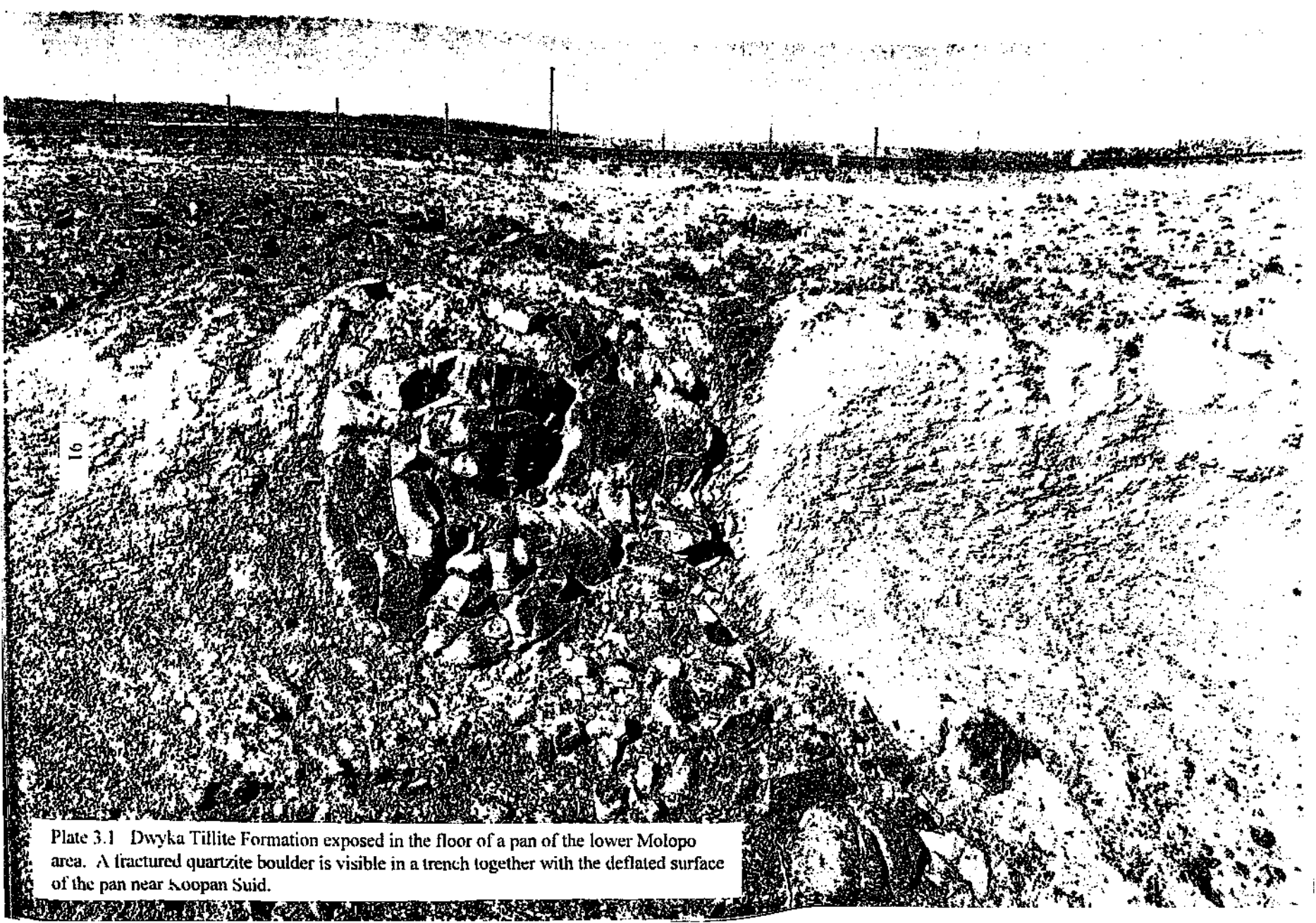


Plate 3.1 Dwyka Tillite Formation exposed in the floor of a pan of the lower Molopo area. A fractured quartzite boulder is visible in a trench together with the deflated surface of the pan near Koopan Suid.

Crowell (1970). There has been some discussion about the calcareous nature of some of the tillite deposits in the Molopo Valley. Calcification (calcretisation) was indicated as a typical feature of the outcrops of the Dwyka series in the Molopo River (Boocock and Van Straten, 1962), but Gerrard (1965) pointed out that there are also superficial deposits resting unconformably on the Dwyka Tillite which are calcified *in situ* and seem only indirectly related to the tillite. The tillite in Gordonia was interpreted as being of glacial-marine origin, with a landmass suggested somewhere to the north, and deeper-sea conditions prevailing to the south (Thomas *et al.*, 1988). Sedimentary rocks belonging to the Prince Albert Formation overly the Dwyka Formation also in this area, and outcrops occur between Rietfontein and the Kalahari Gemsbok Park. The Prince Albert Formation is interpreted as representing a marine environment with slow deposition, which was periodically interrupted by periods of rapid influx of coarser arenaceous material (Thomas *et al.*, 1988). The stratigraphy and nomenclature of the sediments of the Karoo Sequence differs in Botswana somewhat from that of the South African as agreed upon by the Committee for Stratigraphy (SACS, 1980). In south-west Botswana the sediments of the Dwyka Group have been subdivided into the basal massive tillites of the Malogong Formation overlain by the pebbly mudstones and shales of the Khuis Formation, and the siltstones of the Middelputs Formation (Thomas and Shaw, 1991).

3.2.2 Dwyka Tillite suboutcrops

Except for minor outcrops the Dwyka Formation occurs mainly as suboutcrops in the western region of the present North West Province (Smit, 1977). Suboutcrops of Dwyka tillite have been described by Smit (1974a, 1974b, 1977) and Visser (1983a). The suboutcrops of the Terra Firma area indicate the presence of tillites under up to 200 m of sediments of the Kalahari Group. The presence of the Dwyka Tillite is based on data from boreholes, and a thickness of about 950 m of the tillite was indicated by gravity profiles (Smit, 1974a). Smit described the presence of an elongated basin with a NE-SW trend, here called the 'Terra Firma Valley', filled with Dwyka sediments, to be of post-Waterberg age. He further suggested that subsequently, during post-Karoo times, a drainage basin was cut into the relatively soft Dwyka shale and tillite. The valleys of the

post-Karoo drainage system were later filled in with Tertiary (Kalahari Group) sediments. The present superimposed drainage on the Kalahari Group sediments originated according to Smit (1974a) only during late Tertiary times.

Further to the south, in the area of the 2522 Morokweng geological map, these suboutcrops of the Dwyka Formation in the Terra Firma Valley consist mainly of dark shales and tillite (Smit, 1974b). A maximum thickness of 610 m was penetrated in a single borehole without reaching the base of the formation, while gravity measurements indicated a thickness of about 660 m about 30 km to the west of Heuningvlei. In an area south-east of Morokweng several separated outliers of Dwyka Tillite occur underneath the Kalahari Beds on Basement granite. Borehole data indicate a thickness exceeding 300 m in places (Smit, 1974b). Similar glacial valley fills (e.g. in the 'Hotazel Valley') occur in the area covered by the 2722 Kuruman geological map. The groundwater in the tillite formation has generally very high concentrations of Na, Cl and SO_4 which according to Smit (1977) is a result of deposition in an epicontinental sea and subsequent ineffective draining of the salt. The groundwater in the overlying Kalahari Group sediments is generally of a similar saline nature.

3.2.3 Kalahari Basin glacial stratigraphy

The overall glacial stratigraphy of the Kalahari Basin is considered to be fairly simple, with five recognised stratigraphic units in the south and three units toward the north (Visser, 1983b). The five units of the southern region, of most relevance to the Molopo drainage, are described briefly below.

(v) An upper tillite of up to 150 m faintly bedded very fine tillite. Probably formed by basal melt-out of a small floating ice shelf fed by outlet glaciers from the southern highlands Visser (1983b).

An upper shale/siltstone unit of about 80 m thick. The basal shale is interpreted as clays deposited from suspension while the siltstone probably represents a distal pro-delta deposit.

(iii) A main tillite unit of up to 255 m thick made up of a monotonous succession of massive fine tillite with isolated erratics measuring up to 30 m across. This tillite is interpreted as a basal melt-out from a large floating ice shelf.

(ii) A bedded heterolithic diamictite with a maximum thickness of 230 m. It consists of an alternation of diamictite, small-pebble conglomerate, immature sandstone, siltstone, rhytmite shale and dark-coloured mudstone with ice rafted debris.

(i) A basal tillite interpreted as a lodgement till of up to 100 m thick occurs in the valleys. It is generally a coarse tillite with abundant locally derived debris.

3.2.4 Depositional model for the Dwyka Formation

It has been suggested that the ice sheet probably first developed sometime later in the Carboniferous, reached its full extent in the late Carboniferous and disappeared towards the end of the early Permian (Frakes and Crowell, 1970). Wet-based valley glaciers would have advanced in the pre-glacial valley system during the valley glaciation stage, followed by the advance of the ice sheet during the ice-sheet stage. There is overwhelming evidence of glaciomarine conditions in the southern Kalahari (Visser, 1983a) during the final deglaciation stage. The presence of free-swimming, planktonic and neritic forms of marine fossils in the glacial beds towards the central part of the Kalahari Basin has been reported (Martin and Wilczewski, 1970; Visser, 1983a). As the wet-based ice retreated following valleys on to the land, streams deposited fluvioglacial sediments which formed deltas into the drowned valleys. The model of a partly grounded retreating ice mass with down-valley slumping of till deposited near the ice front has been favoured by Visser (1983b). Icebergs produced by calving of the retreating ice front

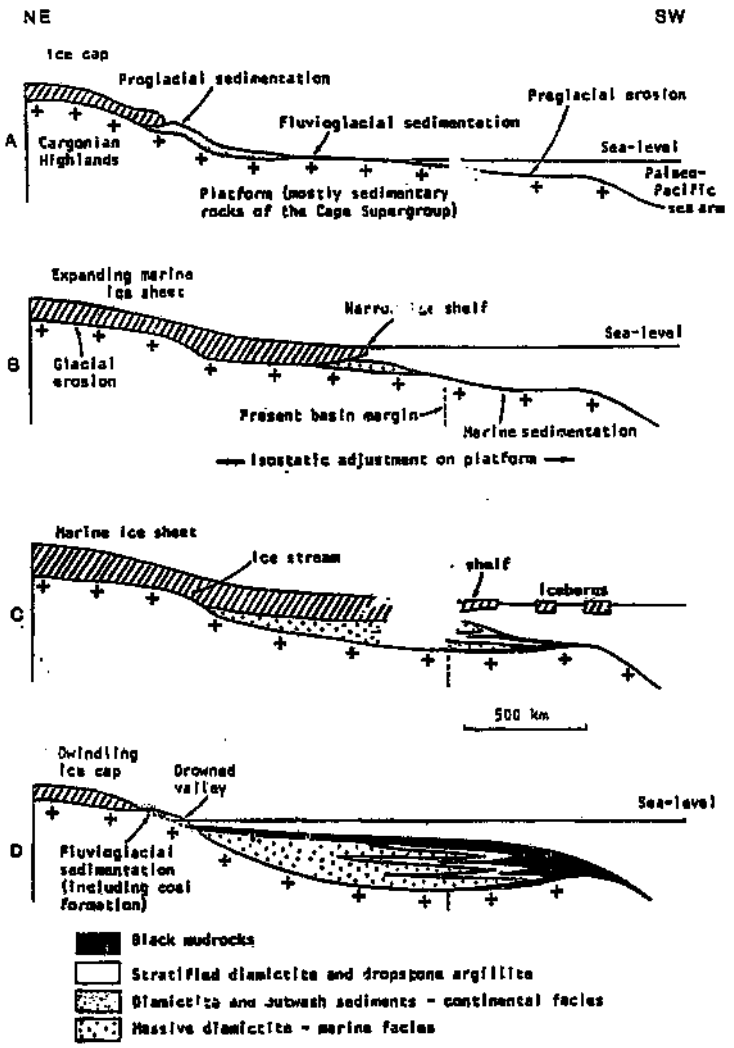


Figure 3.3 Stages in the growth and disintegration of the Permo-Carboniferous marine ice sheet in the Dwyka basin (after Visser, 1989).

transported debris down-valley, while fine sand, silt, and mud were deposited by low-density turbidity currents in the more distal marine environment. Basal melting of the icebergs released the debris which then settled among the mud on the valley floor (Fig. 3.3).

Glacial stages

A model with three stages for the Loeriesfontein area in the north-western part of the main Karoo Basin has been suggested (Visser, 1981). In his sequential model there would first be a stage with valley glaciers, then the stage of the formation of an ice sheet, followed by a deglaciation stage. The recognition of these stages allows for a more realistic interpretation of local and regional palaeo-ice flows based on a study of the palaeo-topography.

The first stage is the 'valley glaciation stage'. The palaeo-topography of his study area in the main Karoo Basin is interpreted as representing a fluvial landscape modified by later glaciation Visser (1981). The palaeo-ice flow at this stage would follow the palaeo-topography quite strictly, where temperate glaciers with basal meltwaters would erode the valleys during an advancing glacier stage. Deposition in the valleys would indicate: breccia ⇒ bedded deposits ⇒ massive tillite (= units (v), (iv), and (iii) of Visser's glacial stratigraphy of the Kalahari *(Karoo)* Basin).

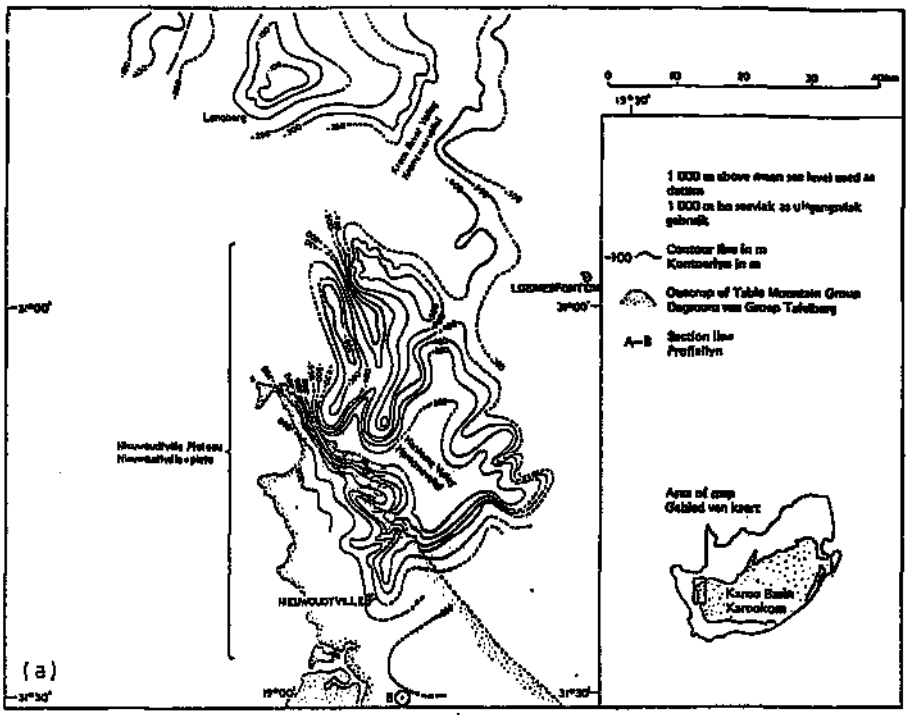
The second stage is the 'ice-sheet stage'. The entire area became covered by an ice sheet, and former valleys now acted as outlet glaciers. The flow of the ice sheet over the larger part of the area was now unconstrained by the basement topography, and the degree of bedrock erosion depended on the thermal regime of the basal ice. The spreading centre for the ice would have been the inland area of southern Africa. This inland area was a high-lying plateau ('Transvaal Highlands') according to Truswell (1977, as cited by Visser, 1981). Visser further mentioned that repeated advances and retreats of the ice front could have breached some watersheds with the formation of 'cols'.

The third stage is the 'deglaciation stage'. During this stage the palaeo-topography became again a controlling factor in the ice flow and in the mountainous area the ice sheet reverted to conditions of valley glaciation (Visser, 1981). With further amelioration of the climate, abundant meltwater would be available and this would have resulted in the deposition of transition beds at the top of the Dwyka Formation.

The palaeo-ice flows and glacial stages, discussed above, were determined for two valley systems in the area around Loeriesfontein (Fig. 3.4) in the north-western part of the main Karoo Basin (Visser, 1981). The local and regional palaeo-ice flows determined for the Hotazel Valley area within the Molopo drainage show a similar trend as for the Loeriesfontein area (Visser, 1987, Fig. 8). The Hotazel Valley is located on the southern edge of the Kalahari (*Karoo)* Basin (Fig. 3.5), and shows a palaeo-ice flow direction towards the north, that was determined by the pre-glacial fluvial valley. The regional ice sheet flow is almost at right angles to the flow direction of the valley glacier, curving from ENE to WSW and W, and appears less constrained by the basement topography (Fig. 3.6). Both the Loeriesfontein Valley systems and the Hotazel Valley region were influenced by the same (Transvaal) ice spreading-centre. It is therefore suggested that a similar model with the sequence of stages as described for the Loeriesfontein area can apply to the evidence of the Dwyka glaciation in the Hotazel Valley.

3.3 Dwyka Palaeogeography

The thickest deposits of the Karoo Sequence (> 9 km, Dingle *et al.*, 1983) occur in the intracratonic main Karoo Basin. In the Kalahari Basin the Karoo Sequence is much thinner, but still widespread (Tankard *et al.*, 1982). Within the Kalahari the rocks of the Upper Palaeozoic and Mesozoic Karoo Sequence lie unconformably upon Precambrian lithologies, and have a poor surface exposure as most of them are covered by sediments of the Kalahari Group (Nash, 1992). The Dwyka sedimentation has been strongly influenced by the pre-Karoo topography (Thomas and Shaw, 1991). In the south-west of the Kalahari (*Karoo*) Basin the sedimentation has been greatly influenced by a high but



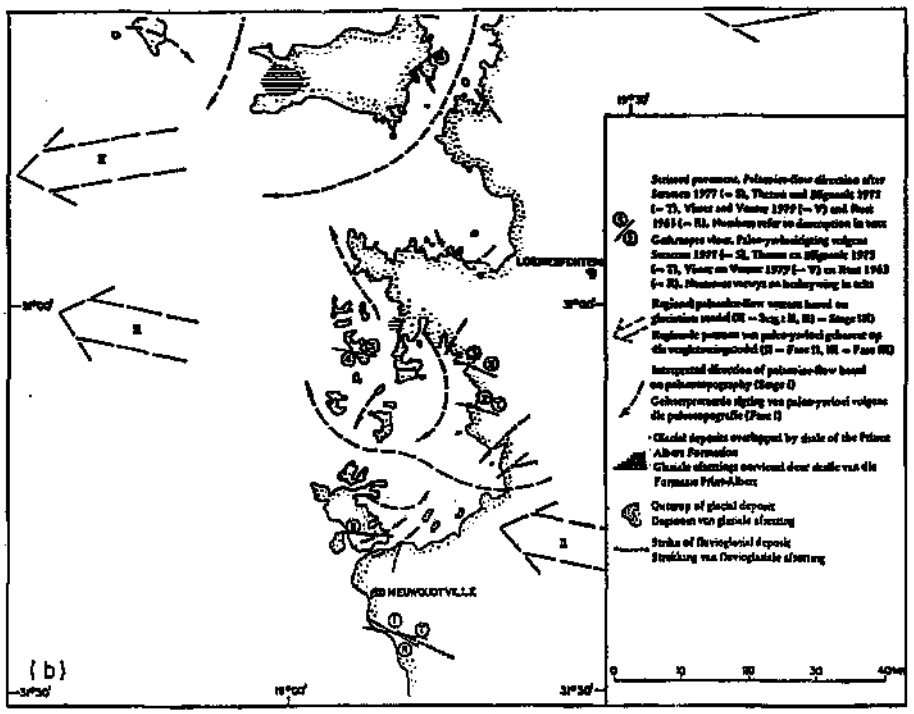


Figure 3.4 Palaeotopography (a) and interpretative palaeoice flow map (b) of the Dwyka Formation around Loeriesfontein (after Visser, 1981).

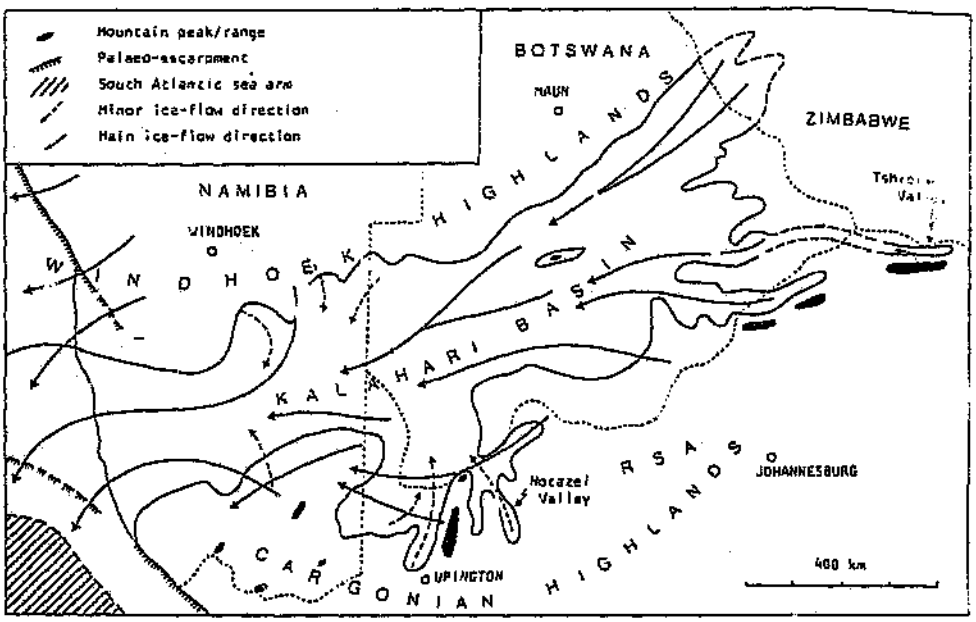


Figure 3.5 Outline of the Kalahari Basin during the Maximum glaciation (after Visser, 1987).

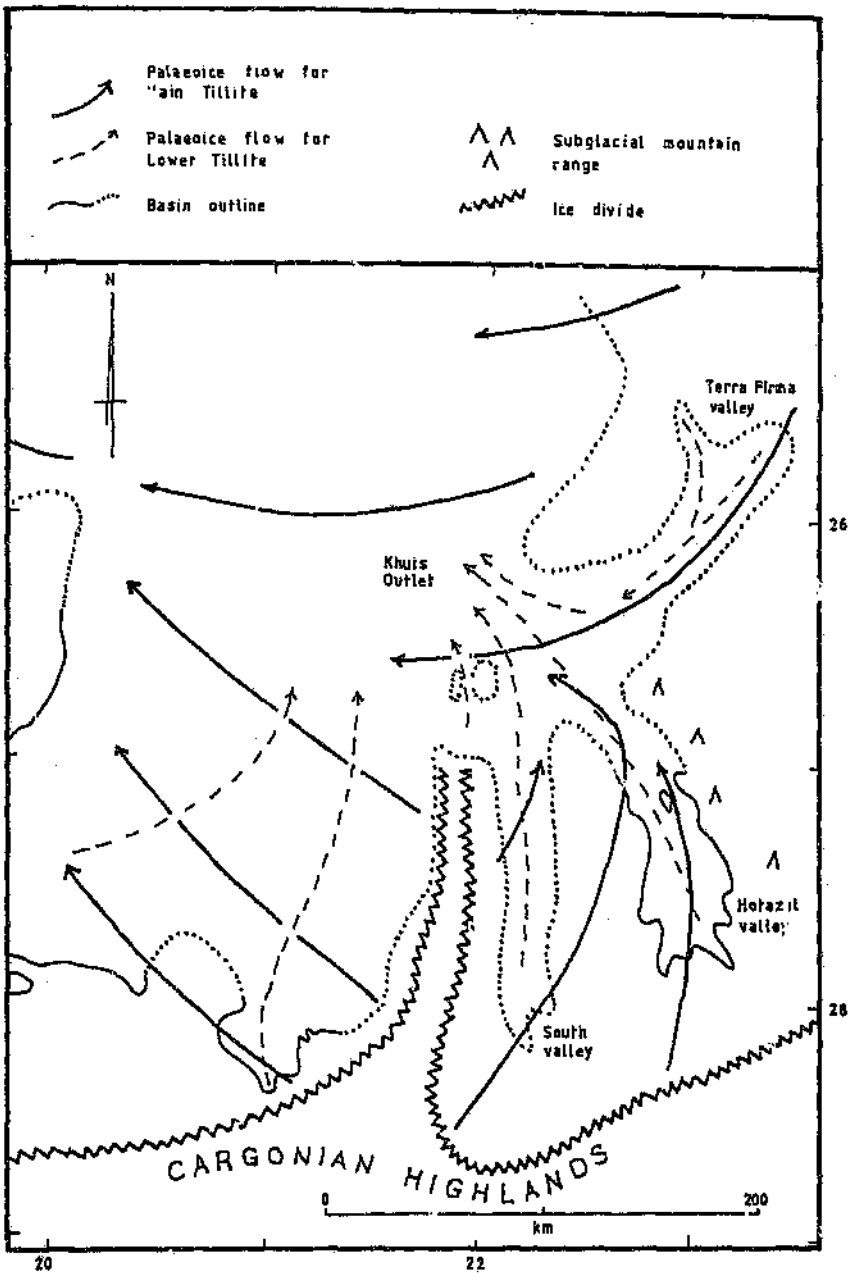


Figure 3.6 Palaeoice flow directions in the Molopo drainage area. Based on Visser (1983a).

strongly dissected upland (Visser, 1983a as cited by Thomas and Shaw, 1991). It was suggested that the effect of this upland diminished as the topography was lowered during the progression of the Karoo.

3.3.1 Preglacial topography

The general palaeogeography of southern Africa has been reconstructed, based on a study of the re-exposed glaciated landscape, buried glaciated surfaces, and the glacial deposits (Visser, 1987). The preglacial landscape was that of an extensive continental highland dissected by river systems (Fig. 3.7). The pre-glacial valley system of the present Molopo drainage area formed part of a major river system which exited into a marine embayment in the west (Visser, 1983b). The rivers drained prominent mountainous regions which were partly surrounded by lower Palaeozoic sediments of the Cape Basin in the south. The highlands may have been bounded by a southern palaeo-escarpment during the Carboniferous. The regional palaeo-slope was towards the south and the west, with a sea present in the south-west. The origin of the sea could be attributed to a drowned abortive rift, connected with the palaeo-Atlantic Ocean, which formed during an incipient break-up of Gondwana (Visser, 1987).

3.3.2 Palaeogeography during maximum glaciation

The morphological elements at the end of the Carboniferous (Stephanian) represent mostly preglacial features that were modified and eroded as well as isostatically depressed during the glaciation (Visser, 1987). Towards the north-west of the Kalahari Basin existed a massive inland plateau ('Windhoek Highlands') which may have extended into South America (Fig. 3.8). The highlands attained altitudes of about 1500 m above sea level (Martin, 1981 as cited by Visser, 1987). The Windhoek Highlands represented a major ice-spreading centre, from which there were major ice-flows into South America and minor flows into the Kalahari Basin (Frakes and Crowell, 1969).

The Kalahari Basin was a very irregular, glacially modified, fluvial valley or trough, with a width of between 200 and 500 km and a length of about 1400 km (Visser, 1987). The

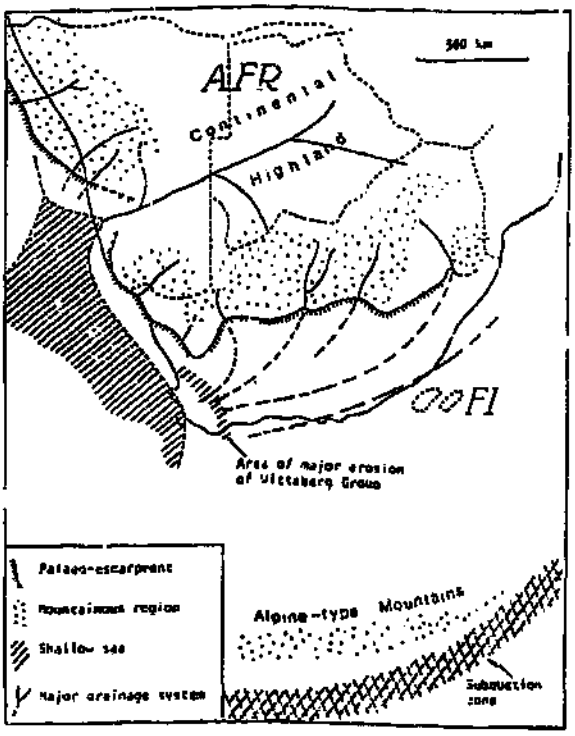


Figure 3.7 Schematic pre-glaciation palaeogeography of southern Africa (after Visser, 1987).

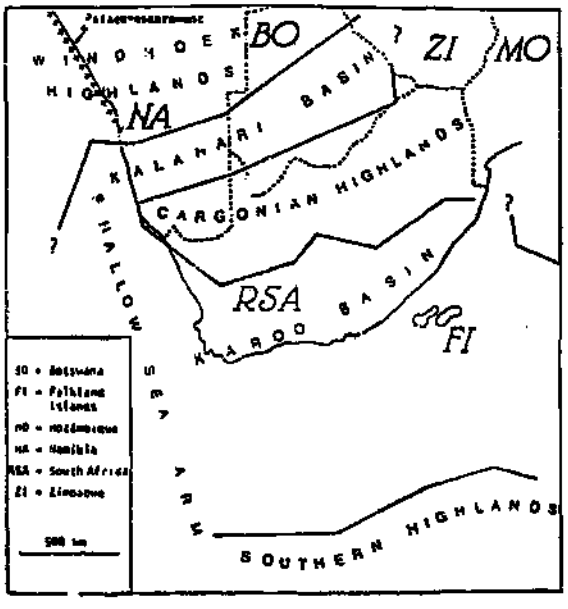


Figure 3.8 Geomorphic elements of part of south-western Gondwana during maximum glaciation (after Visser, 1987).

western extension of the basin probably opened into a sea. A water depth of about 600 m at the outlet of the Kalahari Basin towards the end of the glaciation was postulated by Martin (1973). Visser (1987) suggested therefore on the basis of the thickness of the glacial deposits in the upper reaches of the basin that glacial overdeepening lowered the Hotazel Valley to between 800 and 1000 m below sea-level.

To the south of the Kalahari (*Karoo*) Basin existed the Cargonian Highlands (*Cargonian* was derived from *Car*boniferous and *Gon*dwan*ian*: Visser, 1987). The Cargonian Highlands were a generally E-W mountainous region, which could have connected with the Windhoek Highlands in the north-east. The relief of the Cargonian Highlands has been described as having elevations of between 2000 and 2500 m above sea-level in the north and elevations of about 1500 m in the south where it formed a palaeo-escarpment in places (Visser, 1987). The terrain also sloped towards the east with suggested heights of 300 m above sea-level in northern KwaZulu-Natal. Visser (1987) estimated that during maximum glaciation the Cargonian Highlands were covered by about 3000 m of ice. The valleys in the highlands were described as partly sediment-filled, winding with uneven floors. The 'Hotazel Valley' (Fig. 3.9), now occupied by the Kuruman and GaMogara Rivers, had a length of about 140 km, a depth of about 750 m and a width that varied between 2 km in its upper reaches to 26 km down-valley. The 'Kaap Valley', presently occupied by the Dry Harts and lower Vaal Rivers, was broad and shallow (200-300 m deep), and had a length of about 350 km. Some of the valleys became glacially overdeepened basins in their lower reaches.

The Cargonian Highlands formed a ice-spreading centre with major ice flows towards the south and south-west. The Transvaal ice sheet on the Cargonian Highlands spread from the present Northern Province into the Karoo Basin (du Toit, 1921) . A massive ice lobe, the 'Botswana Lobe' (Frakes and Crowell, 1970) invaded the Kalahari Basin. Frakes and Crowell recorded at least three advances for the Botswana Lobe. An additional minor Griqualand Ice Sheet with a tiny ice lobe south-east of Khuis was identified by du Toit (1921). This ice sheet apparently originated on the Cargonian Highlands between the

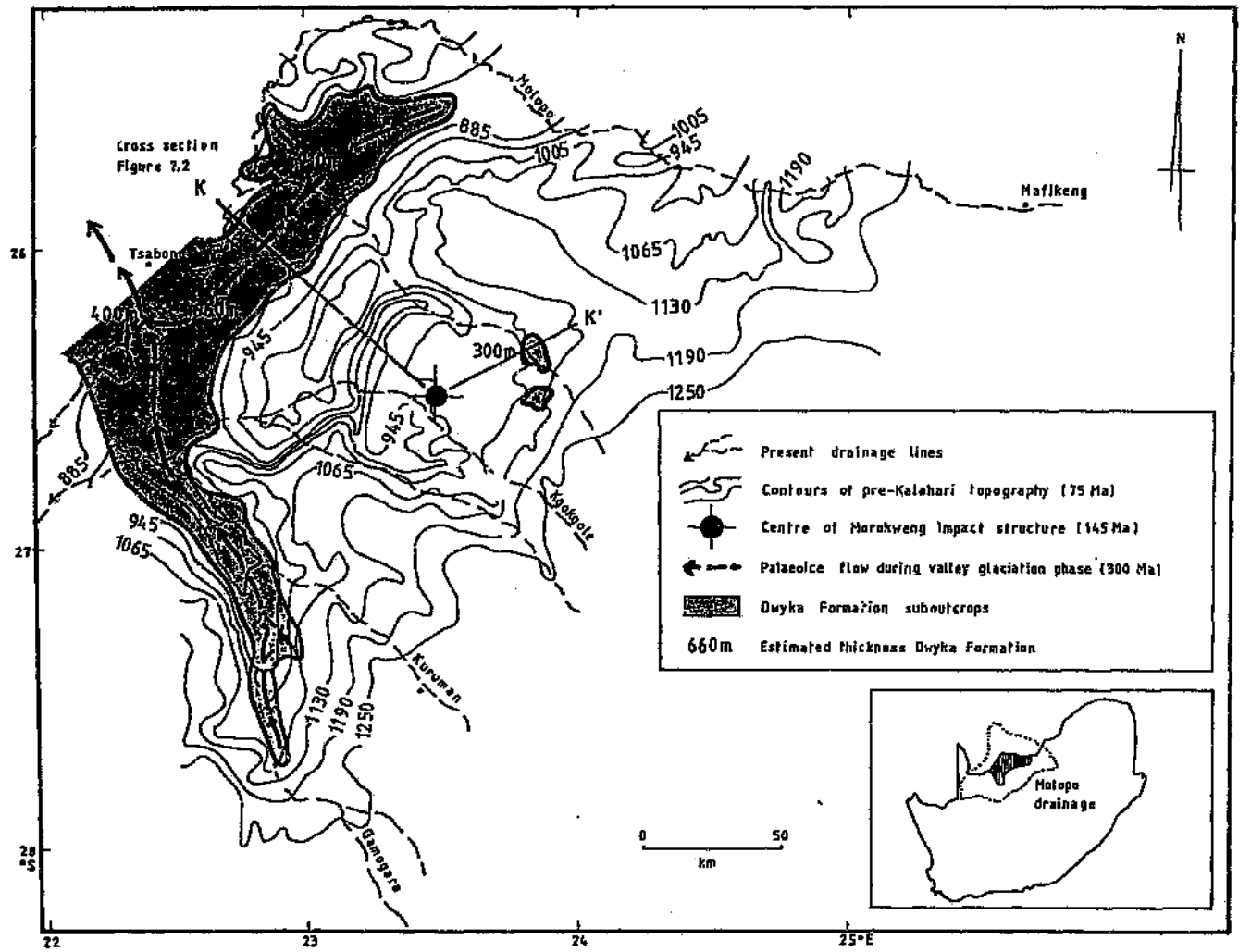


Figure 3.9 Pre-Kalahari topography and Dwyka Formation suboutcrops in relation with the Morokweng impact structure.

Botswana Block and the main Karoo Basin. During periods of rapid expansion of the major ice sheet the ice also flowed across the high interfluves towards the west. Almost all basement rock was submerged under the ice (Visser, 1987), but an ice divide existed (Visser, 1983) west of 'South Valley'. This ice divide was made up of a mountain range as well as a separate mountain peak near Khuis-Middelputs. Such places had Ecca shales overlapping the Dwyka beds. Extensive palaeovalleys have been identified on the area of the exhumed remnants of the Cargonian Highlands (Fig. 3.10) by Visser (1983a).

3.3.3 The Hardap Interglacial and Tses Glaciation

At least three major ice advances occurred in the upper reaches the Hotazel Valley during the glacial history of the Kalahari Basin (Frakes and Crowell, 1970). After a maximum glaciation phase a deglaciation phase occurred possibly towards the end of the Carboniferous due to warmer conditions and a rise in sea-level (Visser, 1983a). A dissected upland with an intricate system of valleys and broad lows was created by ice stream erosion. The deglaciation culminated into the Hardap Interglacial, when small ice caps were left on the higher mountains of the Cargonian Highlands. The presence of fish remains and silicified wood serve as an indicator of benign conditions. Outwash fans formed in the upper reaches of the valleys (Visser, 1983b) and ice-rafted debris was deposited were ice tongues entered the fjords (e.g. Hotazel Valley).

During the early Permian the Tses Glaciation (Visser, 1983a) took place and the ice caps expanded again. A marine ice sheet with ice domes, outlet glaciers, ice streams and ice shelves formed (Fig. 3.11). One major ice dome within the Cargonian Highlands was centred more or less over Mafikeng. Ice flow radiated from the ice domes and resulted in northerly and westerly ice flows in the area of the present Molopo drainage. Nunataks were suggested to have existed on either side of the Hotazel Valley. Amelioration of the climate at the end of the Tses Glaciation caused a rapid disintegration of the ice sheet, and only small ice caps remained for some time along the Cargonian Highlands. Visser

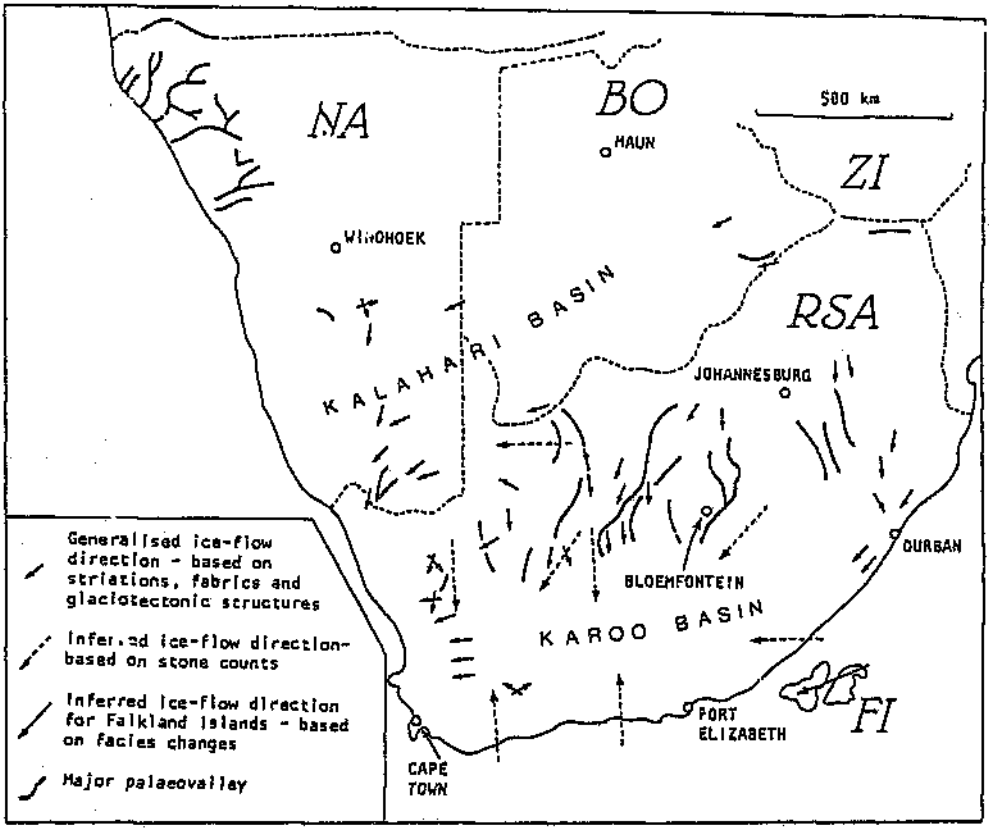


Figure 3.10 Major palaeovalleys in southern Africa (after Visser, 1987).

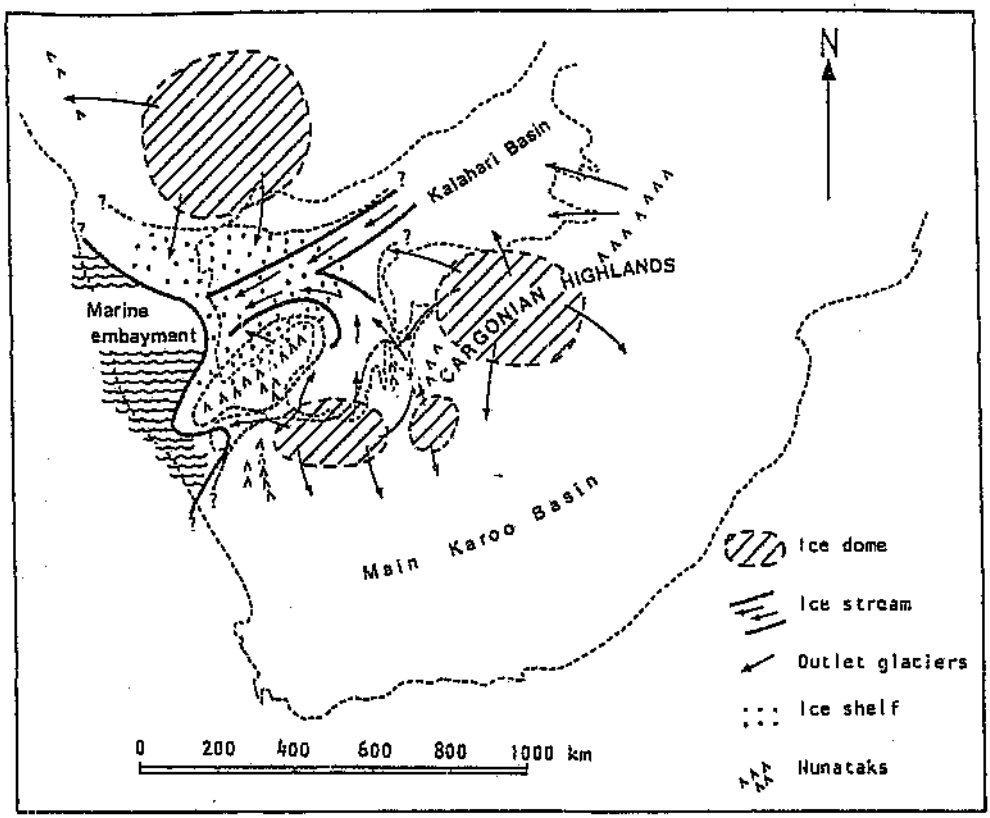


Figure 3.11 The ice extent during the Tses Glaciation (Early Permian) at its maximum and at an advanced stage of disintegration (after Visser, 1983a).

(1983a) further suggested that fine-grained sediments were subsequently deposited over the entire area, with the exceptions of major peaks, as it was inundated by the Ecca sea.

3.3.4 Postglacial flooding

Towards the end of the Permo-Carboniferous during the Sakmarian stage (282-269 Ma; Harland *et al.*, 1982), within the early Permian period, large parts of Gondwana became flooded (Visser, 1987). Visser contributed this to a combination of sea-level rise and isostatic depression. Postglacial rebound was apparently not sufficient to elevate the sediment-water interface above sea-level. A sea arm extended into the southern Kalahari and the Whitehill Sea was formed in the main Karoo Basin, between the remnants of the Cargonian Highlands and the Southern Highland (Fig. 3.12). Peripheral fluvioglacial systems existed along the basin margin and Tankard *et al.* (1982) suggested that large areas of floating ice developed through the annual thaw from the reducing ice caps. After the melting of the last ice regional subsidence apparently cancelled out the effects of isostatic rebound along the western parts of the Cargonian Highlands (Visser, 1987). Areas which were less affected by the regional subsidence experienced extensive erosion of the glacial sediments and basement rocks. The entire Molopo area is estimated to have been inundated and fine-grained sediments of the Ecca Group were deposited on both the glacial deposits and the exposed basement. The basement rocks of the higher areas of the Molopo drainage were thus most probably covered with sediments of the Ecca Group.

* * * * *

In this chapter the evidence of the Permo-Carboniferous Dwyka glaciation in the southern Kalahari and more specifically in the Molopo drainage has been reviewed. Outcrops and suboutcrops of the Dwyka Tillite Formation together with exhumed glacial features have provided significant information on the pre-Dwyka topography as well as Dwyka palaeogeography. There are indications that a fluvial valley system existed prior to the Permo-Carboniferous glaciation (~300 Ma). The primeval Molopo drainage system originated on a continental highland of Gondwana situated in central southern Africa and

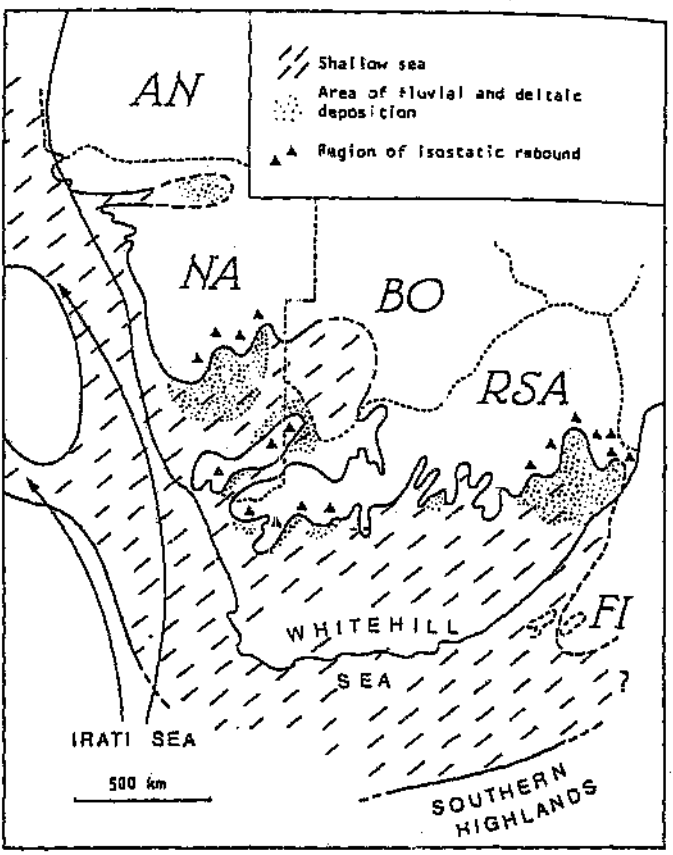


Figure 3.12 Distribution of post-glaciation seas over southern Africa (after Visser, 1987).

was part of a larger Kalahari (*Karoo*) drainage line, that exited towards the south-west in an incipient form of the South Atlantic Ocean. The main valley of this ancient Molopo drainage was situated somewhat to the east of the present Molopo River. A major tributary, the Hotazel Valley, was situated where the present GaMogara River is situated, and a third valley system with a north-south trend between the Langeberge and the Skurweberge. These three valleys drained through the 'Khuis Outlet' in a north-westerly direction into the southern Kalahari (*Karoo*) Basin and further into an incipient sea arm. During the Permo-Carboniferous glaciation they became remoulded and possibly deepened. In the later stages of the glaciation and after the glaciation the western half of the Molopo drainage area became flooded by seawater. It is suggested that after a long period of erosion the drainage conditions of the Molopo drainage area were dramatically affected by the Morokweng meteorite impact (145±2 Ma). The next chapter deals with the Kalahari Group sedimentary sequence formed within the Molopo drainage area, and discusses the beginning of the developments of the Molopo drainage during the Cainozoic.

CHAPTER 4 CAINOZOIC MOLOPO KALAHARI SUB-BASIN

4.1 Introduction

It is the aim in this chapter to identify drainage indicators that can be deduced from a scrutiny of the spatial and other aspects of the Kalahari Group sediment body within the sub-basin, the Kalahari Group stratigraphy as well as pre-Kalahari palaeo-topography of the region. Emphasis will be on the integrated drainage basin in contrast with some earlier studies of the Kalahari Group sediments which have concentrated on particular sections of the Molopo drainage. It is believed that such an investigation of this sub-basin is vital to the development of the model of the long-term drainage evolution of the Molopo.

4.2 Kalahari Group Stratigraphy

The development of the knowledge concerning the sediments of the mega-Kalahari, now known as the Kalahari Group (SACS, 1980) has been reviewed by Thomas and Shaw (1991). The lack of natural exposures has been a major limiting factor on the investigations of the stratigraphy of the Kalahari Group and the development of a unifying stratigraphy is still far from complete. After a review of some of the general developments and problems in the mega-Kalahari the situation in the Molopo sub-basin is reviewed below.

4.2.1 Mega-Kalahari stratigraphy

Because of the lack of natural outcrops most of the data gathered in the mega-Kalahari have been from the surficial and immediate sub-surface units, while the thickness of the Kalahari Group deposits is generally based on borehole data. Thomas and Shaw (1991) discussed the problems of correlation, stratigraphy and age of the deposits. They argued that a major problem is the fact that not all identified stratigraphical units are present everywhere and that it is impossible to assign a deposit to a stratigraphical position on lithological factors alone. They also cautioned against the chronostratigraphical interpretation of the lithological units of the Kalahari sediments. This refers especially to the cemented sandstones which are duricrusts and not necessarily the product of primary deposition. These duricrusts can from not only on a regional scale but also on very local scales. Their age determination has proven to be extremely difficult. The absence of diagnostic fossils is a further handicap to the development of a stratigraphy, and dating of the various events in the mega-Kalahari during the Cainozoic is in general still very uncertain (Thomas and Shaw, 1991).

Six major lithological units in the Kalahari group have been identified for the mega-Kalahari by Thomas and Shaw (1991): conglomerate and gravel; marl; sandstone; alluvium and lacustrine deposits; Kalahari Sand; and duricrusts. The Kalahari Sand and the duricrusts were indicated as most important in terms of their areal extent. In addition they identified four minor lithological units: diatomaceous earths; shell beds; pan sediments and evaporites; and cave deposits. Although these minor units have limited spatial distributions, but they are deemed important as palaeoenvironmental markers.

4.2.2 Molopo drainage stratigraphy

du Toit (1954) had suggested a classification for the Kalahari Basin of a lower sequence of 'Kalahari Beds', which were primarily mostly consolidated, and an upper sequence of unconsolidated 'Kalahari Sands'. Knowledge about the stratigraphy and palaeoenvironments of the Molopo sub-basin has progressed through various studies in

different sections of the Molopo sub-basin. The central, southern and northern sections are discussed briefly below.

Central section

The first detailed data concerning the Kalahari group sediments within the Molopo drainage became available during the early and mid-1970s through the work of Smit (1974a, 1974b, 1977) For over a decade Smit analysed the data of over two thousand boreholes for water in the Bray-Morokweng-Kuruman area (Smit, P. J., pers. comm., 1995). On the basis of his analyses he determined the pre-Kalahari surface, and the thickness variation and stratigraphy of the Kalahari Group sediments. Smit (1977) identified 4 four units:

(iv) aeolian sands, rarely > 20 m thick (top);

(iii) fine-grained clayey and calcareous sandstones. Silcrete and calcrete are common in the sandstones. Up to 80 m thick;

(ii) red calcareous clay containing layers of fine gravel or coarse sand, with sandstone lenses. Up to 100 m thick;

(i) clayey gravel and gravelly clay. Up to 100 m thick (bottom).

In a number of cross-sections across filled-in pre-Kalahari drainage lines (Figs. 4.8 and 4.9) Smit (1977) indicated the vertical and horizontal extent of the units in the Bray-Morokweng-Kuruman area. The three lower units of Smit's subdivision corresponded with du Toit's (1954) Kalahari Beds.

SACS (1980) created a formal Kalahari Group for the South African section of the Kalahari Basin which included the four units as suggested by Smit (1977):

(iv) the Gordonia Sand Formation (top) which included the cover sands and fossil dunes of the Gordonia District in the Northern Cape Province, the western Free State, the western section of the North West Province, and the Molopo valley;

(iii) the Eden Sandstone Formation for a clayey calcareous sandstone as exposed on the farm Eden 703/46;

(ii) the Budin Clay Formation for calcareous gravelly clays as exposed on the farm Budin 495 on the Moshaweng River north-east of Kuruman; and

(i) the Wessels Gravel Formation (bottom) for a basal clayey gravel unit as exposed in a shaft of the Wessels manganese mine.

The Gordonia, Eden, and Budin Formations are all exposed in the Mamatwan manganese mine, halfway between Kuruman and Hotazel, and are further discussed in section 6.3.1.

Southern section

In a discussion of the Tertiary and Quaternary periods in the Gordonia District, Botha *et al.* (1986) presented an extended version of the SACS (1980) Kalahari Group stratigraphy (Table. 4.1). In between the Eden and Gordonia Formation they suggested in addition the Mokalanen Formation for a duricrust of up to 20 m thick, consisting of nodular and hardpan calcrete and silcrete as exposed on the farm Mokalanen 175 along the Molopo River. Above the Gordonia Formation they suggested the Lonely Formation for a buff grey diatomaceous limestone found in some pans and in bluffs of the Kuruman River. They cited Heine (1978) that the Lonely Formation represented a wetter period from about 19 000 to 12 000 BP in a generally dry cycle. Both newly suggested formations have not yet been approved by the SACS.

Table 4.1 Stratigraphy of the Kalahari Group in Gordonia (Botha *et al.*, 1986).

GROUP	PERIOD	FORMATION	LITHOLOGY
			Fine-grained, brown to white pan sediments
	Quaternary	Lonely[#]	Buff-grey diatomaceous limestone found in some pans and in bluffs of the Kuruman River
		Gordonia	Reddish aeolian sand
KALAHARI			Gravels derived from weathering of Dwyka Tillite
		Mokalanen[#]	Nodular and hardpan calcrete, silcrete
	Tertiary	Eden	Red, brown and green sandstone, calcareous grit, conglomerate
		Budin	Red and brown calcareous clay and marl
		Wessels	Clayey gravel

[#] Not yet approved by the South African Committee for Stratigraphy (SACS)

Northern section

Drilling investigation of the basic and ultrabasic rocks of the Molopo Farms Complex also led to an improved knowledge of the distribution and isopach patterns of the overlying Kalahari Group sediments in the northern section of the Molopo drainage area (Gould *et al.*, 1987). In addition to a brief, general description of the Kalahari Group in the area (Gould *et al.*, 1987 and Gould and Rathbone, 1985), detailed descriptions of the borehole logs of Kalahari Group sediments above the rocks of the Molopo Farms Complex were also made available (Gould *et al.*, 1989). Although the stratigraphy of Smit (1977) can easily be recognised, a slightly more complex stratigraphy was revealed in some of the borehole logs. Most of the SACS (1980) recognised Kalahari Group formations can readily be identified in a partial section across the major southward trending pre-Kalahari, proto-Molopo palaeovalley (Fig. 4. 1). The Gordonia Formation is the top formation and attains a thickness of about 45 m in borehole log MF 10A. The Eden Formation, sandwiched between the Gordonia and Budin formation varies in thickness between about 70 m (MF 6) and 30 m (MF 10A). The Budin Clay Formation is present in all logs and is about 50 m thick in log MF 9. Gould *et al.* (1987) commented that the red clay horizon occurs from approximately 90 to 120 m below the surface, and in places more than one clay horizon is developed. The stratigraphy below the Budin Formation, with well developed calcretes and silcretes, appears more complex than reported from the other sections of the Molopo drainage. Sporadic gravel horizons occur in the calcretes and silcretes underneath the red clay (Gould *et al.*, 1987). Only log MF 10A has a basal gravel conglomerate which is comparable with the descriptions of the Wessels Gravel Formation. The basal conglomerate in the Molopo Farms area seems less widespread (Gould *et al.*, 1987) than in the Molopo drainage area south of the present Molopo River (Smit, 1977).

A comparison of the stratigraphy of the three sections discussed above suggests that the stratigraphy, as developed for the southern section of the Molopo drainage, seems to be quite satisfactory for the entire Molopo sub-basin at this stage, and no further

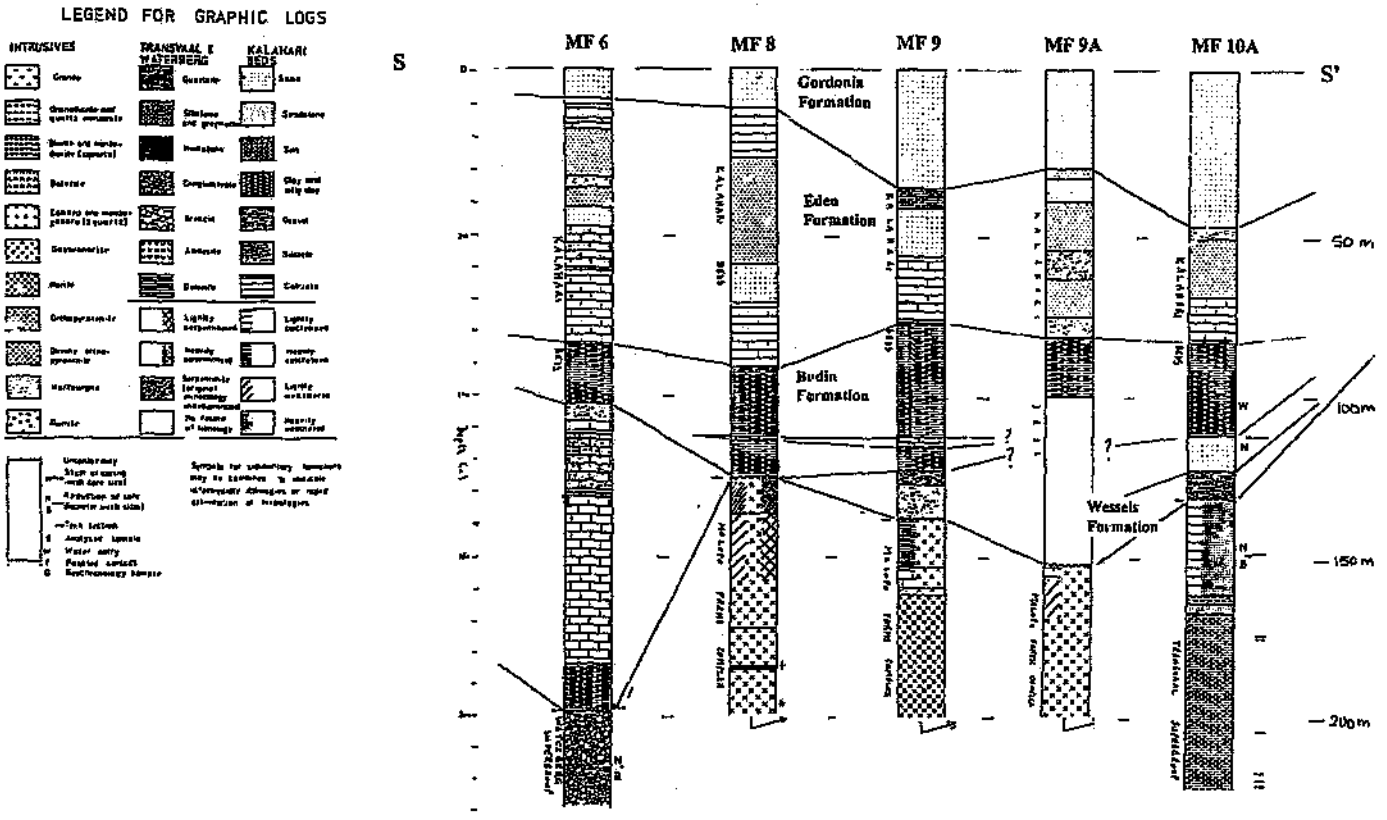


Figure 4.1 Graphic borehole logs across a pre-Kalahari palaeochannel. Based on data from Gould *et al.* (1989). See Fig. 4.3 for the location of the section.

stratigraphic investigation has been undertaken for this large-scale study of the Molopo drainage.

4.3 Palaeoenvironmental Reconstructions for the Molopo Drainage Area

During Tertiary times northward flowing pre-Kalahari river-valleys became choked with gravel, clay and sand eroded from nearby highlands by periodic floods (Smit, 1977). Smit further suggested that the blockage of the old valleys eventually forced streams to cut new channels. After the channels had been choked and flooding became less intense calcretisation progressed. During progressive aridification desert conditions were established and fine sands spread over the entire area. Smit (1977) thus explained the changing sedimentation patterns during the infilling of the Kalahari Basin mainly in terms of climatic changes. In addition, Smit also constructed detailed maps of the thickness variation of the Kalahari Group sediments, as well as a map of the pre-Kalahari topography in the Kuruman-Bray area. The pre-Kalahari topography (Fig. 1.5 a) has no corrections for Kalahari basining, that occurred since the initiation of the Kalahari Basin, and should thus not be called just 'pre-Kalahari' but 'pre-Kalahari as affected by basining'. Although Smit (1977) clearly stated in his text that the pre-Kalahari was towards the north, his drawing of the pre-Kalahari drainage revealed a southward flowing pattern. This pattern was extended further towards the south-west, into the Gordonia District (Fig. 1.6) in the work of Levin *et al.* (1985).

Botha *et al.* (1986) also presented a Kalahari Group isopach map with a palaeodrainage reconstruction for the Gordonia area (Fig. 4.2). It appears that the northward flowing palaeodrainage interpretation was based on a pre-Kalahari topography, that had also not been corrected for subsidence due to lowering of the Kalahari Basin. The late Cretaceous and the early Tertiary were according to Botha *et al.* (1986) fairly wet periods during which north-flowing palaeorivers were established. They suggested that the initial courses of the palaeorivers could have followed valleys formed during the Dwyka glaciation. The basining during Kalahari times resulted then in a reversed direction of flow towards the central part of Botswana. It can thus be inferred that (although not specifically

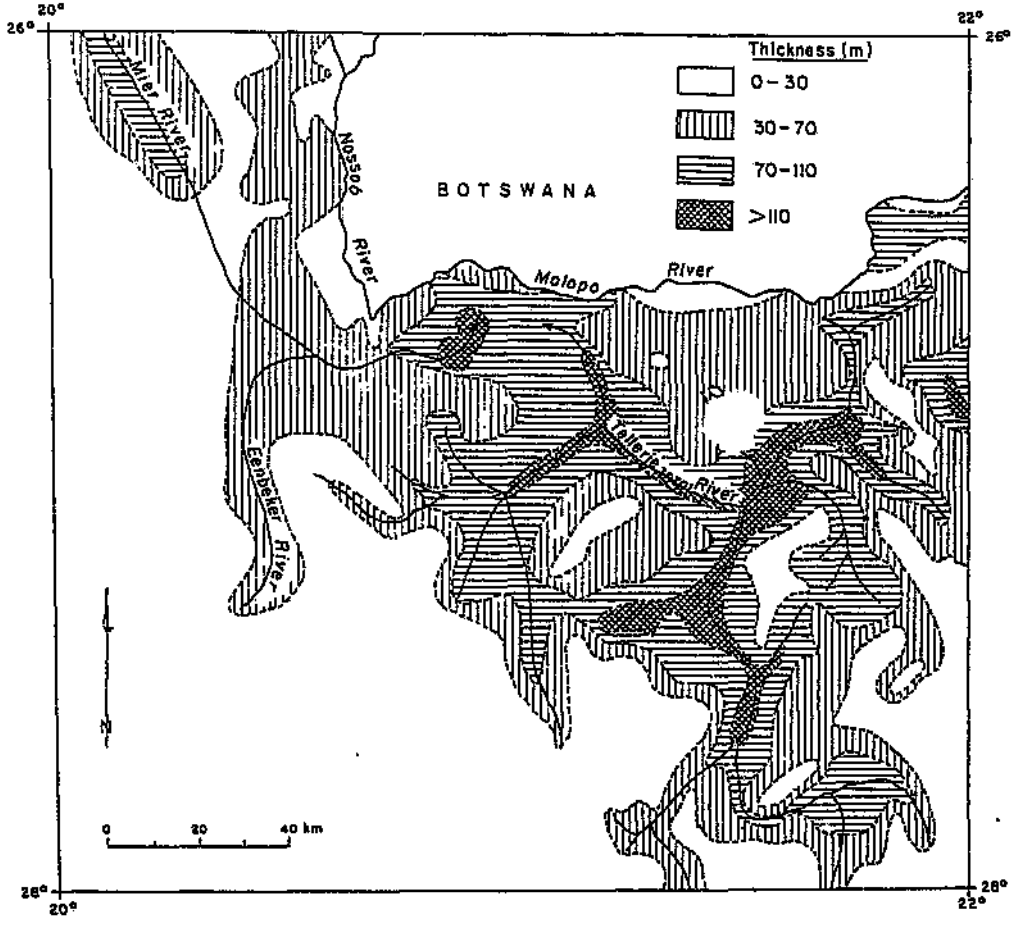


Figure 4.2 Isopach map of the Kalahari Group in Gordonia (after Botha *et al.*, 1986).

mentioned) they envisaged a southward flowing palaeodrainage prior to the start of the Kalahari basining, which became reversed during basining. Deposition of fluvial sediments in the palaeovalleys (Wessels and Budin Formations) followed during a wet cycle in the early Miocene (Botha *et al.*, 1986). Subsequent deposition of the Eden Formation extended beyond the palaeovalleys as deposited by a north-flowing braided-river system on a fairly even surface. The subsequent fairly dry, semi-arid conditions during the late Miocene and early Pliocene resulted in the formation of cretes (Mokalanen Formation) on the African and end-Tertiary Surface. A more humid period during the middle Pliocene resulted in the accumulation of Dwyka-derived gravels as well as the incision of the modern, westward flowing rivers such as the Molopo, Kuruman and Orange. Arid to extreme-arid conditions during the late Pliocene resulted in the widespread deposition of aeolian sands. Fluvial conditions during the Quaternary finally resulted in the widespread development of pans, and the establishment of the present-day drainage pattern in the Gordonia district. Botha *et al.* (1986) also indicated that dating of the geological events after the break-up of Gondwana is highly speculative due to the lack of fossils and/or suitable rocks for radiometric dating. They emphasised the influence of climatic change on the deposition of the sediments of the Kalahari Group

4.4 A Discrete Molopo Sedimentary Sub-basin

A number of smaller basins within the contiguous Kalahari Basin have been identified by Thomas (1988). South of the Ghanzi Ridge Thomas identified three sub-basins (Fig. 1.4), of which the basin that he identified in the south-western section of the Kalahari Desert is termed here the 'Molopo sub-basin'. Thomas (1988) indicated specifically a lower area around the 'Great Bend' of the Molopo River where the thickness of the Kalahari Beds is between 201-300 m. With the problems that are encountered in developing an integrated stratigraphy for the mega-Kalahari, it is suggested that the Molopo sub-basin should be seen as a discrete sub-basin of the larger Kalahari Basin with its own typical stratigraphy such as proposed by Botha *et al.* (1986). Similar discrete

sub-basins each with their own set of regional influences and their own typical stratigraphy could exist for example in the Auob-Nossob area and in the Okavango area.

4.4.1 Kalahari Group isopach pattern

In more detail the thickness of the sediments of the Kalahari Group in the Molopo sub-basin can be read from the isopachs on the latest issues of the 1:250 000 maps of the Geological Series (2522 Bray, 2524 Mafikeng, 2620 Twee Rivieren, 2622 Morokweng, 2624 Vryburg, and 2720 Noenieput). Further information has been provided by Smit (1977) for a section of the drainage area south of the Molopo River in the combined Bray-Morokweng-Kuruman region. The thickness variation of the sediment in the Molopo Farms area north of the Molopo River has come forward from investigations towards the ultra-basic Molopo Farms Complex rocks (e.g. Gould and Rathbone, 1985; Gould *et al.*, 1987). The isopachs in the Molopo Farms Project area are based on many fewer boreholes and show much less topographical detail than the area south of the river. Behr (1989) briefly also discussed the thickness variation, mainly based on data from the Botswana side of the Molopo River. More recently digital data of the sediments of the Kalahari Group within the area of Namibia, Botswana, and South Africa have recently become available on CD-ROM from the Council for Geoscience.

4.4.2 Pre-Kalahari surface

The topography of the pre-Kalahari surface in the Molopo sub-basin can easily be determined by subtracting the Kalahari Group isopach values from the altitude values of the present Kalahari surface topography. Since the present Kalahari surface is nearly flat, though slightly tilted towards the west, the pre-Kalahari topography can generally be read from the isopach values, where the thickest Kalahari Group deposits indicate the presence of former drainage lines. It should be born in mind that this pre-Kalahari surface is nothing more than the theoretical Kalahari Basin surface with the Kalahari Group sediments removed. To obtain a true pre-Kalahari topography, corrections for the basining should be made to the surface. The Kalahari Basin deepened progressively as the sediments accumulated and Neogene tectonics influenced the subcontinent.

Weathering characteristics

The pre-Kalahari surface outside the Kalahari Basin has generally been greatly eroded during the Cainozoic (Partridge and Maud, 1987). It is, however, from the surface beneath the Kalahari Group sediments within the Kalahari Basin that some more information about the character of the pre-Kalahari surface may become known. The borehole descriptions of the Molopo Farms area provide details of various degrees of weathering and calcretisation (Gould *et al.*, 1989). Weathering and calcretisation depths vary greatly within the area. The weathering mantle appears to be better preserved in the topographically higher areas and more strongly eroded in the palaeovalleys. Calcretisation seems to be primarily associated with rocktype (e.g. serpentinite, serpentinised dunite, and harzburgite).

Palaeotopography

The palaeotopography of the pre-Kalahari surface of the Molopo Farms Project area exhibits more relief than the present Kalahari surface. Gould *et al.* (1987) indicated measured slopes of 105 m in 2 km and suggested the presence of even steeper slopes. In general, the thickest deposits occur in the southern section along the Molopo River within three southward-orientated palaeovalleys. The palaeovalleys are separated by palaeoridges with isolated palaeohills. Some of these palaeohills protrude through the sediment cover of the Kalahari Group. A major NE-SW palaeovalley, which contains a maximum thickness of 220 m of Kalahari Group sediments near the Molopo River is situated in the eastern part of the Molopo Farms Project area (Fig. 4.3). This palaeovalley connects with the major valley system as identified by Smit (1977). A second NNE-SSW palaeovalley is a tributary to the major palaeovalley. The thickest sediment infilling of just over 180 m occurs near an acute set of bends in that palaeovalley just before it joins the major system. The orientation of these two valley systems seems strongly influenced by two major sets of faults within the area. One set has a NW-SE alignment, while a second set is almost perpendicular, NE-SW. These fracture zones were apparently in existence before the deposition of the Kalahari Group sediments and predate neotectonic

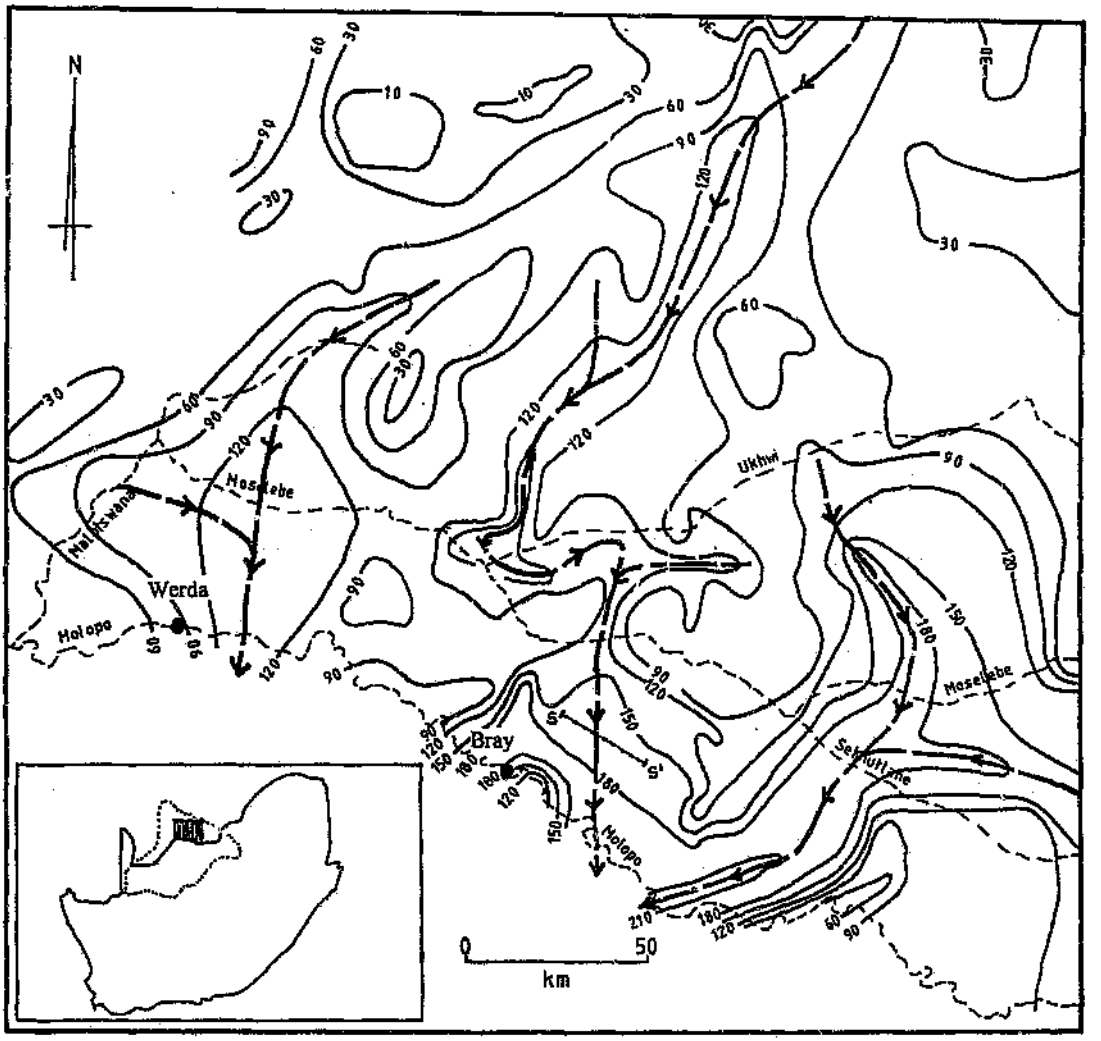


Figure 4.3 Isopachs of Kalahari Beds and proto-drainage lines in the Molopo Farms Project area (after Gould *et al.*, 1987).

structures. A similarly close relationship between fracture zones and valley orientation has been indicated elsewhere in the Kalahari by Shaw and de Vries (1988) and Nash (1992). A third, much smaller tributary N-S palaeovalley is situated just east of Werda. All three palaeovalleys connect well with the pre-Kalahari topography reconstructed by Smit (1977) for the area immediately south of the Molopo River (Fig. 1.5a).

The palaeotopography of the pre-Kalahari surface south of the Molopo River in the larger Bray-Morokweng-Kuruman area has been reconstructed by Smit (1974a, 1974b, 1977). The general pattern is that of a northward deepening valley system on a westward tilted-surface. The deepest part of the valley system seems to be aligned with the underlying suboutcrops of the tillites of the Dwyka Formation. Smit (1977) interpreted the existence of a northward-flowing pre-Kalahari drainage system (section 1.2.2). A tributary valley system from the surroundings of the Morokweng impact structure appears to be post-impact date (145±2 Ma), but pre-Kalahari (75 Ma), since there are no Dwyka Tillite suboutcrops present below this palaeovalley.

The area described by Smit (1977) is bounded in the south-west, near Khuis-Middelputs, by a basement high where the thickness of the Kalahari Group sediments is much less than in the areas to the north-east and further to the south-west of it. This zone of thinner Kalahari Group deposits coincides with the Tshane hinge-line at the western edge of the Kaapvaal structural province (Fig. 2.15). The region to the south-west of the basement high has been described by Levin (1980), Levin *et al.* (1985), Botha *et al.* (1986), Thomas *et al.* (1988), and Thomas and Thomas (1989). The palaeotopography of the area between the Orange and the Kuruman River was first described by Levin (1980). Levin noted that sediments of the Kalahari Group were deposited in a basin structure that seems much older than its suggested Cainozoic age. Evidence from borehole data indicated that sediments of the Nama Group, the Karoo Sequence, and the Kalahari Group were all deposited in almost the same depositional structural basin (Levin, 1980; Smit, 1964). The filling of palaeovalleys with Karoo and Kalahari sediments has created the present-day flat appearance, but the pre-Kalahari topography had much greater relief than the

present-day surface (Levin, 1980). Levin described the occurrence of hollows or troughs in the pre-Kalahari topography which are filled with sandstone and/or clay (Fig. 4.4). He also identified some additional palaeodrainage lines through a study of ERTS-imagery. One such palaeodrainage line, which runs from the Korannaberg to the Kuruman River he called the 'Madala Drainage'. He further suggested that pans in Gordonia are located along certain palaeodrainage lines on the pre-Kalahari erosion surface. Levin *et al.* (1985) put forward an interpretation of pre-Kalahari drainages for most of the Molopo drainage south of the Molopo River, based on the work by Smit (1977) and Levin (1980). He extended Smit's southward palaeodrainage to end somewhere in the middle of the Gordonia Block (Fig. 1.6).

Botha *et al.* (1986) deduced the Tertiary drainage in Gordonia from an isopach map of the Kalahari Group (Fig. 4.2). They described the pre-Kalahari palaeotopography as fairly rugged with palaeovalleys cut into the assumed African surface. They stated that the age of the African surface is uncertain but that it had some relation to Gondwana, its drift onset, and an unconformity in the offshore stratigraphy dated at about 125 Ma. A complex pre-Kalahari drainage pattern with two major northflowing palaeoriver systems (e.g. Eenbeker River and Tellerie River) and one south-flowing river (Mier River) was identified by Botha *et al.* (1986). They indicated that all the rivers drain towards an area north-east of Askham ($26^{0}49'40''$S, $20^{0}45'$E), where the pre-Kalahari topography has at present its lowest point above sea level. Thomas and Thomas (1989) also agreed that the pre-Kalahari drainage was mainly to the north. All authors cited above did not indicate how much the deepening of the Kalahari Basin has influenced the pre-Kalahari drainage. Substantial lowering of the basin in relation to the rim would have strongly influenced the initial pre-Kalahari topography.

Pre-Kalahari structural drainage control

Shaw and de Vries (1988) indicated that the relationship between the drainage network of the present Molopo and the pre-Kalahari geology was not fully understood. It has become apparent from the work of Gould *et al.* (1987) in the Molopo Farms Complex

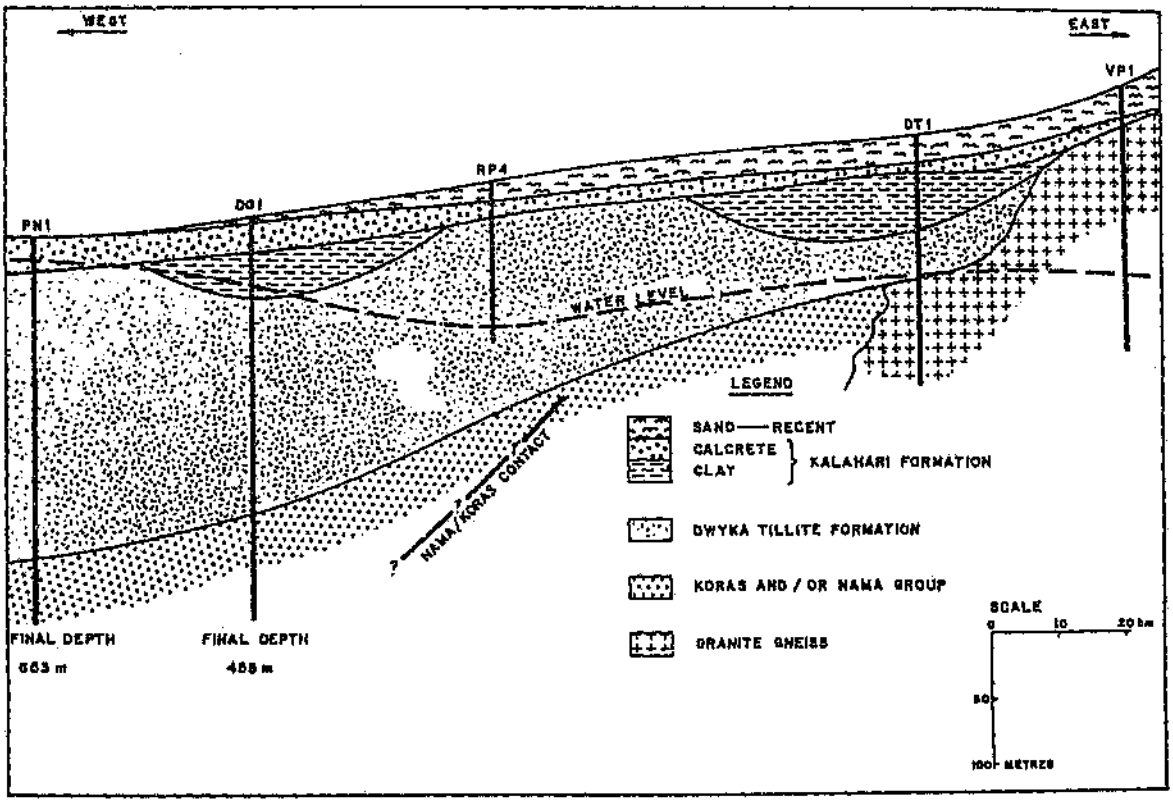


Figure 4.4 East-west geological profile in the area between the Orange and Kuruman Rivers (after Levin, 1980).

area that the course of the pre-Kalahari proto-Molopo was strongly determined by the pre-Kalahari geology. A comparison of the mapped structures of the Molopo Farm Complex area with the pre-Kalahari drainage patterns clearly shows the strong affinity. Dominant faults have NNE-SSW, NE-SW, NW-SE, and WNW-ESE trends (Fig. 4.5) which can easily be recognised in the palaeodrainage lines. With the filling of the sub-basin this affinity has been strongly reduced, and the present course of the Molopo shows a very limited relationship with the pre-Kalahari geology.

4.5 Back-tilted Drainage Basin

The pre-Kalahari topography of the combined area of the Molopo Farms Complex and the area discussed by Smit (1977) forms a long, semi-circularly shaped basin with a centripetal valley system (Fig. 4.6). The length of the entire basin exceeds 400 km. The Molopo Farms area clearly shows a buried southward draining valley system, while the area south of the Molopo indicates a northward-flowing system. The combined picture suggests that this is in fact a back-tilted drainage system that became filled with Kalahari Group sediments (Bootsman, 1997). It appears that prior to the initiation of the Kalahari Basin a proto-Molopo River existed with a largely north-south trend, which drained in a southward direction between the erosion resistant ridges of the banded ironstone towards the east and the quartzites in the west, through the Hotazel Valley, to join the upper proto-Orange (de Wit's, 1993 Karoo River) just south of the present Vaal-Orange confluence. Since most of the topography of southern Africa was covered with Karoo Sequence rocks, the valley would have initially developed into the softer Ecca Group and Dwyka Tillite Formation rocks between the harder ridges. Evidence further indicates (chapter 6) that the proto-Molopo crossed the then presumably Karoo-covered Asbestos Hills between Danielskuil and Griquatown. This proto-Molopo would thus have formed a major artery of the upper proto-Orange. The dramatic change in general drainage direction from north-westerly during the Permo-Carboniferous, to the southerly direction as is evident during the Upper Cretaceous, coincides with the Morokweng meteorite impact (Bootsman and Reimold, 1998).

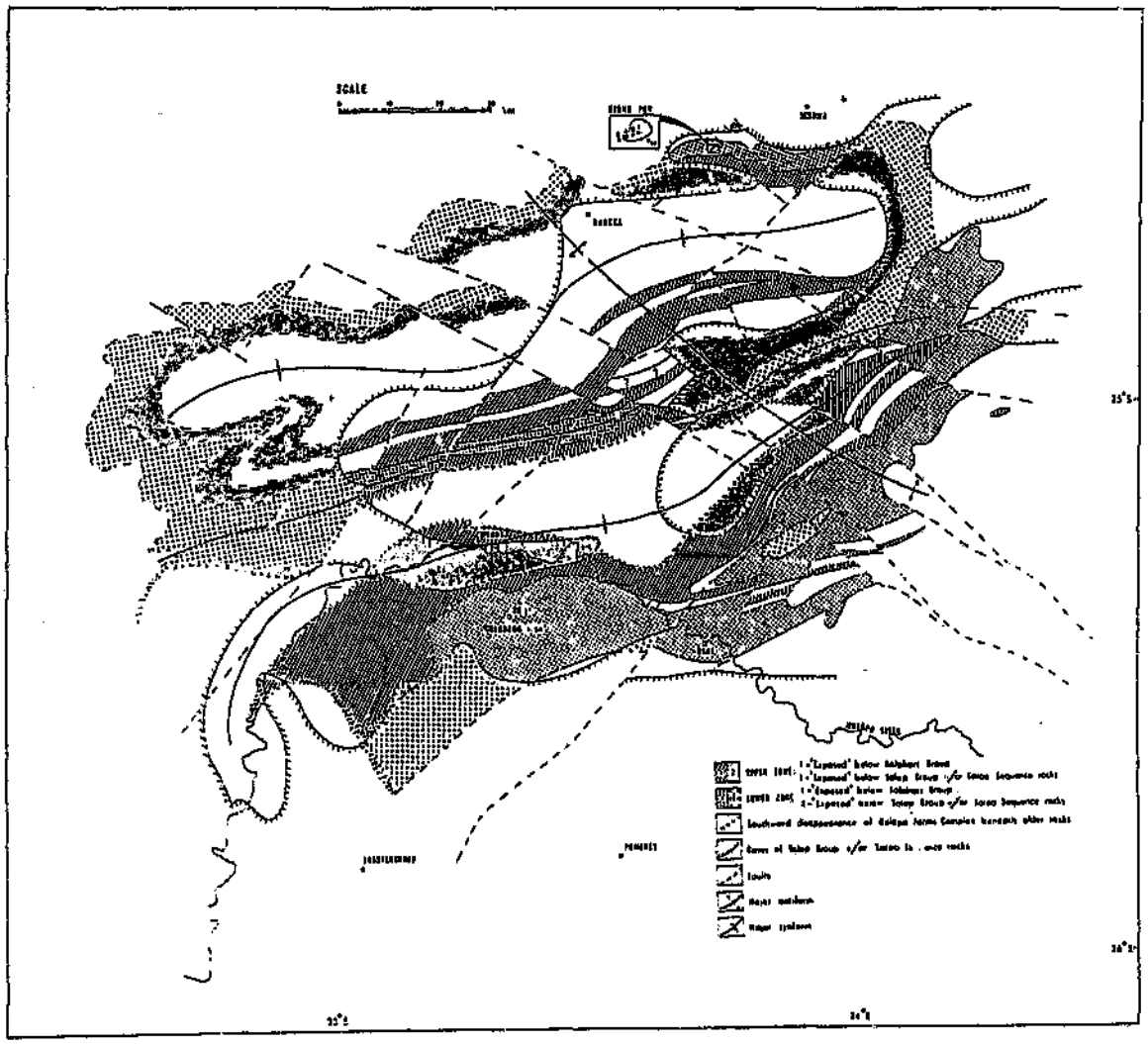


Figure 4.5 Geological structures of the Molopo Farms Complex area (after Gould *et al.*, 1987).

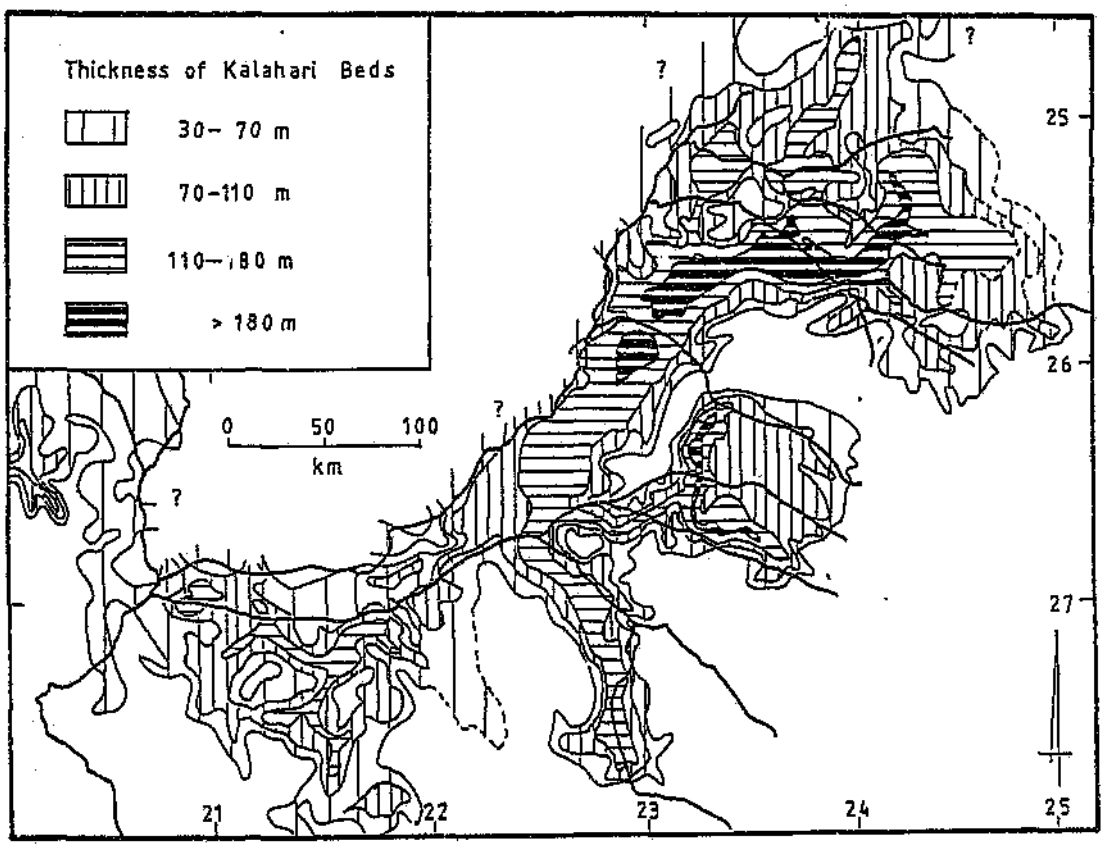


Figure 4.6 Thickness of the Kalahari Beds in the Molopo drainage area.

The initiation of the Kalahari Basin, which is estimated to have occurred at the end of the Cretaceous, would have disrupted this southerly drainage line probably near to the location of the present Griqualand-Transvaal axis of uplift (Fig. 4.7). The eastern edge of the basin may have been determined by an incipient form of the axis. The back-tilted headwater section of the proto-Molopo would initially have formed a large body of standing water that subsequently became filled with sediments. Upstream from the hingepoint reverse flow would have occurred over some distance. The proto-Molopo thus changed from an exoreic drainage system into an endoreic lacustrine system that was to exist for an extended part of the Tertiary. The size and shape of the sediment body of the Budin Clay Formation within the sub-basin provides evidence for the previous existence of the lacustrine phase. Later Neogene uplifts during the Miocene and Pliocene would have further accentuated the basining (Partridge and Maud, 1987) and back-tilting, and have provided further impetus for filling the basin.

The principle of 'ponding upon back-tilting' had earlier been mentioned by King (1951a: 243), in connection with some drainage developments in the Okavango region. The back-tilting of the headwaters of the Mzimvubu River, with respect to the Pliocene uplift along the Ciskei-Swaziland axis in the interior of KwaZulu-Natal, was discussed by Partridge and Maud (1987) as another example of drainage back-tilting. The Cedarville flats were cited as an example of aggradation in such a high-level basin. In a similar fashion aggradation occurred in the Molopo sub-basin.

4.5.1 Lacustrine phase

The occurrence of the extensive, elongate and substantially thick clay body in the Kalahari Group (see Figs. 4.8, 4.9, and 4.10 in combination) suggests that a lacustrine phase existed for a considerable time in the Molopo sub-basin. The nature of the clays is described below on the basis of clay mineral and soil analyses of samples from three locations which are compared with limited available data from Gould and Rathbone (1985), Gould *et al.* (1987), and Behr (1989).

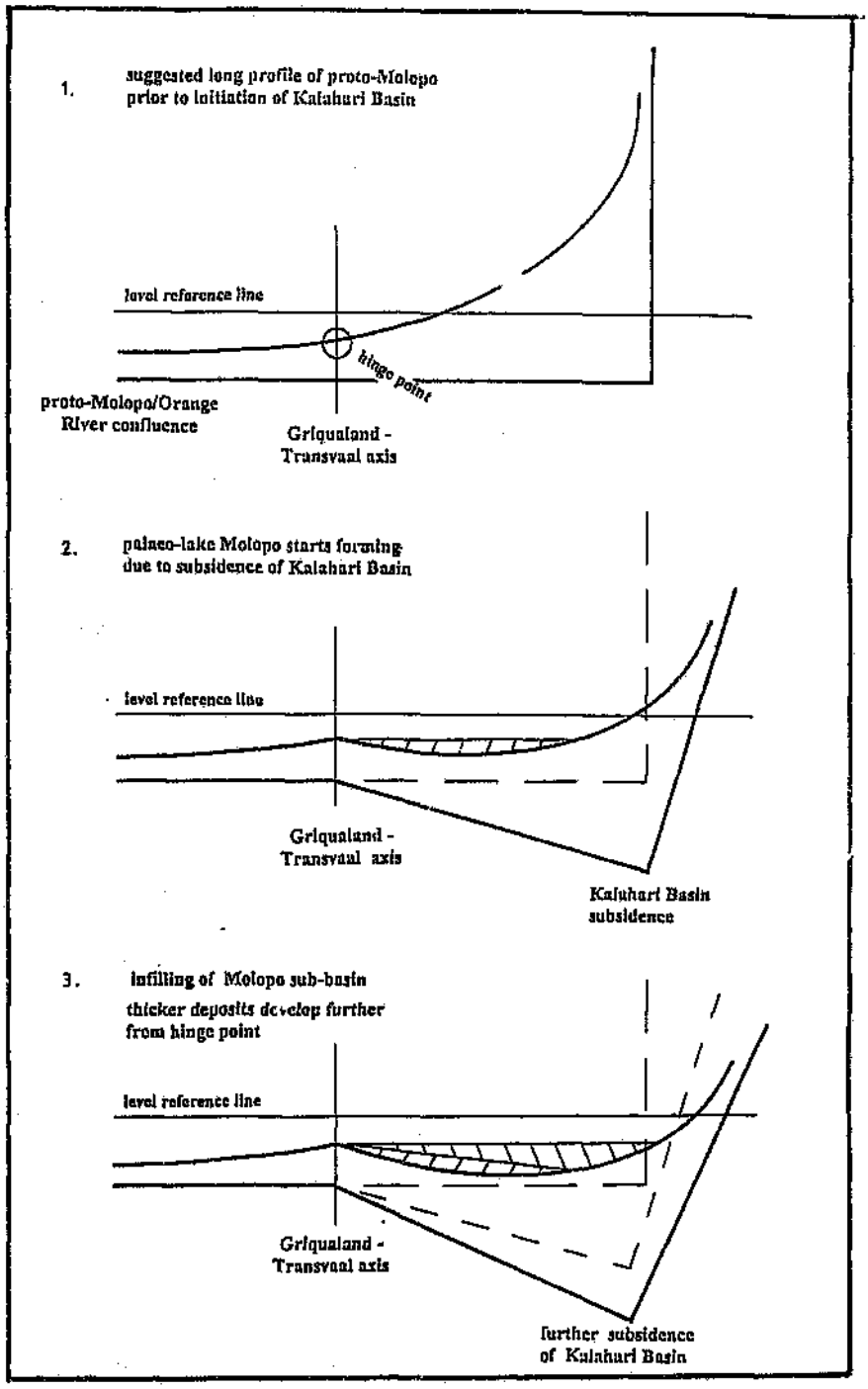


Figure 4.7 The back-tilting principle applied to the proto-Molopo.

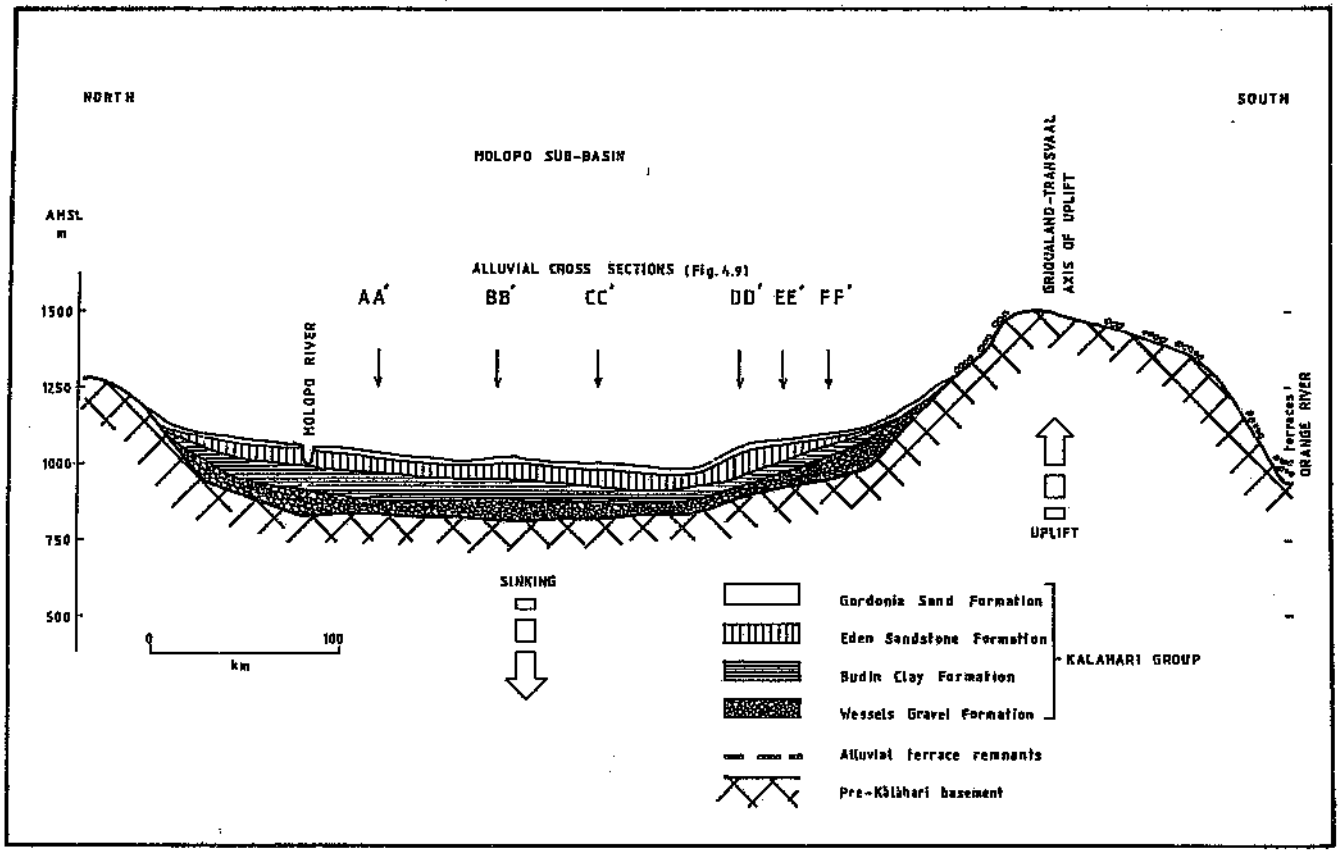


Figure 4.8 Longitudinal profile of the proto-Molopo after back-tilting and isostatic adjustments.

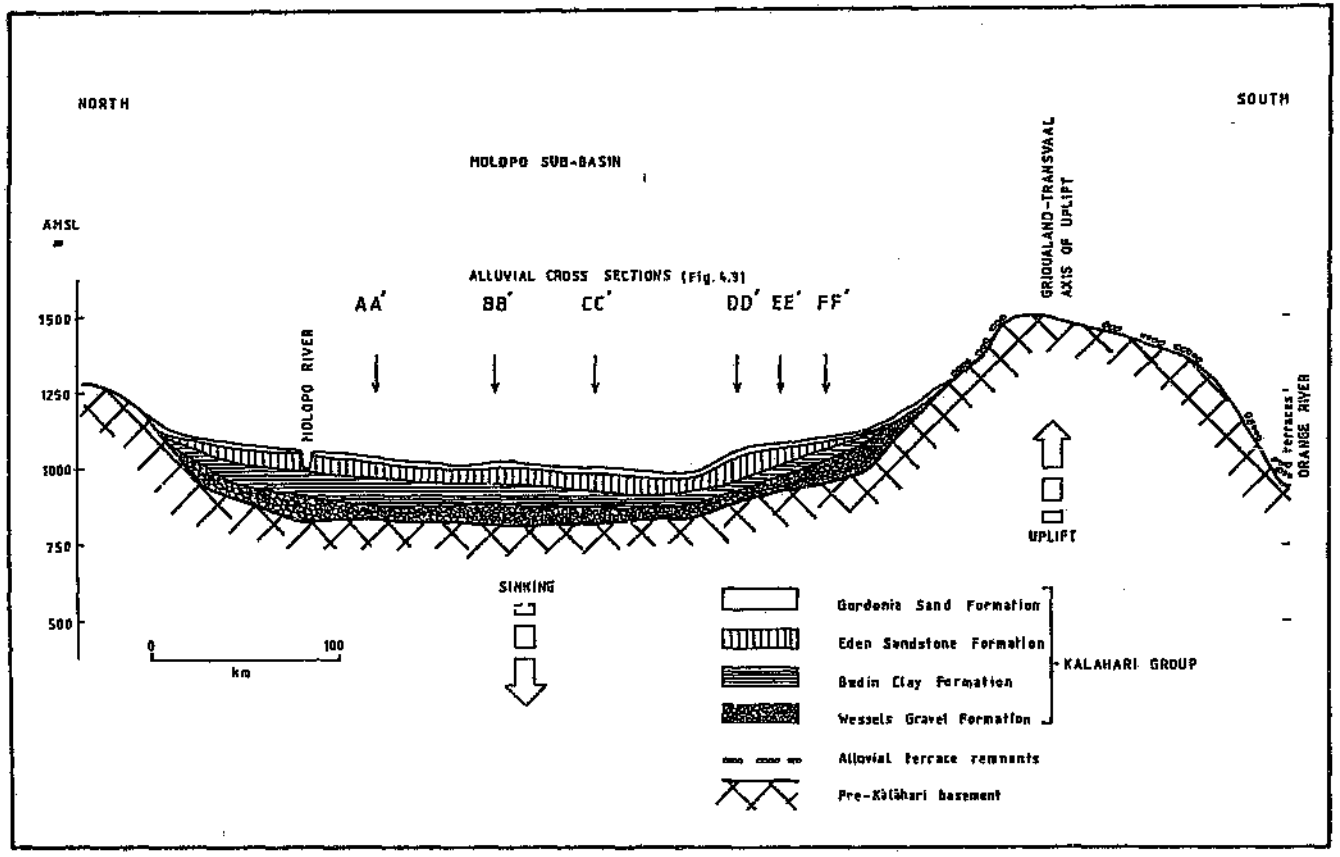


Figure 4.8 Longitudinal profile of the proto-Molopo after back-tilting and isostatic adjustments.

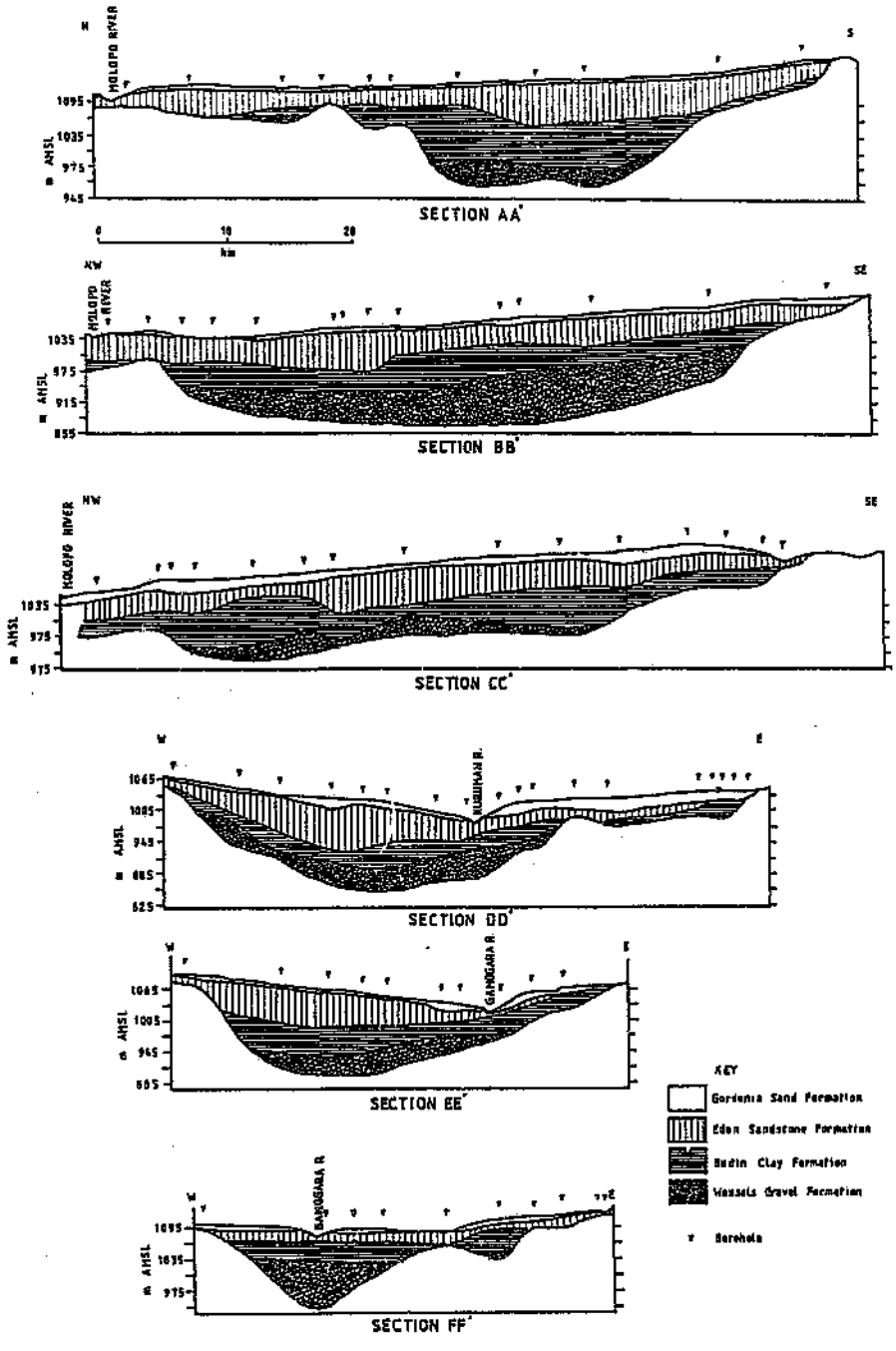


Figure 4.9 Cross sections of the sediment-filled proto-Molopo (modified after Smit, 1977).

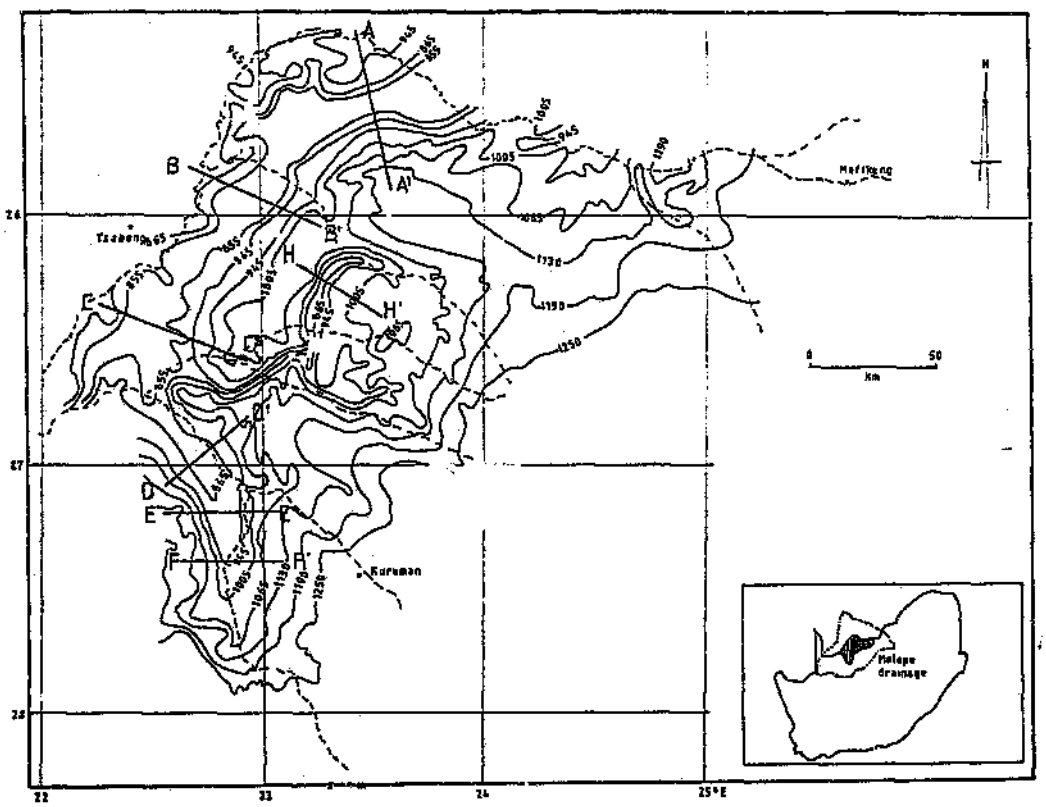


Figure 4.10 Location of alluvial cross sections (Fig. 4.9). After Smit (1977).

Two clay samples, 'Loporung' (G96/11300) and 'Logagane' (G96/11302), are from the north-easterly section of the Molopo sub-basin, and were collected in the Molopo River valley, only about 10 km apart, near the villages of Loporung and Logagane (about 90 km west of Mafikeng) respectively. In both instances the clay occurred as natural outcrops at the bottom of a terrace scarp. The clays were overlain by silcretes (3-5 m) and red Kalahari sands (0.5-1 m) successively. The third clay sample, 'Mamatwan' (G96/11301) is from the Mamatwan manganese mine, about 400 km to the south of the first two locations in the same sub-basin, and is also discussed in section 6.3.1.

The three samples vary considerable in the content of sand and silt, but generally have high clay contents (Table 4.2). The Loporung clay, although sampled far away from the Mamatwan clay has very similar ratios of smectites (montmorillonite) compared to the total clay (about 65%). The Logagane clay stands out by its high total clay percentage (83%) and high mica percentage (49%).

Table 4.2 Grain size and clay mineral analysis of red-brown Budin Clay Formation. The clays were analysed by North West Laboratory, North West Co-operative Ltd, Lichtenburg.

	Loporung Clay	Logagane Clay	Mamatwan Clay
	WT %	WT %	WT %
Coarse sand	4.2	0.4	1.8
Medium sand	4.0	0.4	1.0
Fine sand	20.8	5.2	4.2
Total sand	29	6	7
Silt	16	11	23
Smectite	36	34	47
Mica	10	49	17
Calcite	3	-	4
Quartz	5	-	-
Kaolinite	1	-	2
Total Clay	55	83	70

All three clays have an effervescent reaction when submerged in water, which is probably due to the presence of the extremely chemically active nature of the adsorbed sodium. Further soil analysis of the clays indicates that they all have a neutral to slightly alkaline pH value. The Loporung clay has a very high conductivity (3680 uS/cm) as well as high contents of sodium (12.16 mmol/l) and chloride (19.26 mmol/l) compared to the other two samples (Table 4.3). The Logagane clay has somewhat higher sodium (1.87 mmol/l) and bicarbonate (2.50 mmol/l) values.

Table 4.3 Soil analysis of Budin Clay Formation samples. The clays were analysed by North West Laboratory, North West Co-operative Ltd, Lichtenburg.

		Loporung Clay	Logagane Clay	Mamatwan Clay
Tests	Units			
pH H_2O		7.8	7.8	7.3
Conductivity	uS/cm	3680	217	235
Potassium	mmol/l	0.07	0.15	0.06
Sodium	mmol/l	12.16	1.87	0.94
Calcium	mmol/l	6.81	0.53	0 49
Magnesium	mmol/l	3.52	0.45	0.51
Chloride	mmol/l	19.26	0.84	0.56
Bicarbonate	mmol/l	[illegible]	2.50	1.30
Phosphorous	mmol/l	0.04	0.04	0.05

The combined data from the two tables suggests that these red-brown (Munsell code: 2.5YR 4/4; 3/4 moist), slightly alkaline clays with high smectite values are typical of playa clays as described for example by Matthews (1960), Droste (1961a, b) from present playa environments in the Mojave Desert (USA). Droste (1961a, b) found in the analyses of clay-mineral composition of desert saline sediments in southern California, that this was almost entirely controlled by the composition of surrounding source rocks, and that clay minerals were not significantly controlled by diagenetic processes in the continental

saline playa environments. Droste (1961a) further indicated that high montmorillonite content there was almost entirely due to surrounding volcanic terranes. A similar relationship was also suggested for the Mamatwan clay by Anthony (1996). The major element analyses of the Mamatwan clay and the Ongeluk lavas compare very favourably (Table 6.1). The presence of sodium and bicarbonates in the clay also points to a weakly alkaline environment of formation. Sodium bicarbonate occurs to various extents in many pans within the present arid to semi-arid environment of southern Africa. The substantially high contents of sodium and chloride in the Loporung clay, and high conductivity indicates locally more saline conditions in the environment of deposition than those of the other clays.

The results of the clay sample analyses resemble the interpretations of Gould *et al.* (1987) and Behr (1989). Gould and Rathbone (1985) and Gould *et al.* (1987) indicated that the red clay contained variable proportions of silt and sand. They suggested that in some places the clay is a local weathering product of argillaceous rocks, on which it rests directly, but more usually, especially in the southern section of the Molopo Farms area, it appears transported and resembles a palaeosol developed on lower pediment slopes. Clays from borehole data in the Molopo Farms area were interpreted by Behr (1989) as lake-floor sediments, belonging to the Budin Clay Formation. Clay mineral analysis indicated that the clay fraction was dominated by 'rock-flour' (Behr, pers. comm., 1996).

It is suggested that clays were deposited in saline playa environments which existed in the back-tilted basin of the proto-Molopo during a protracted period of the Tertiary. Possibly it resembled a string of interconnected playas which differed slightly in salinity and salts present, as influenced by local factors. The source area for the clay materials was the surrounding bedrock. A strong correlation between the Mamatwan clays and the Ongeluk Formation was indicated (Anthony, 1996), and the sandiness of the Loporung clays can be correlated with the neighbouring Archaean granites.

4.5.2 Causes of back-tilting

The lowest point of the back-tilted Molopo sub-basin now has an altitude of about 840 m AMSL, while the altitudes of the lowest points in the wind gaps on the Griqualand-Transvaal axis vary between 1510 and 1550 m AMSL. The back-tilting of the proto-Molopo was thus 670-710 m. For the proto-Morokweng River the total calculated amount of back-tilting would be somewhat less (about 600 m) as compared to the altitude of the diamondiferous gravels at Mahura Muthla (1450 m AMSL) near the position of Griqualand-Transvaal axis (see section 6.4.1). Back-tilting of the headwaters of the proto-Molopo could have occurred because of the combination of a number of potential mechanisms involved.

(i) The formation of the Cainozoic Kalahari Basin which has been related to a rejuvenation of older basining structures, as well as sediment-loading on the continental shelf, in response to rapid erosion in the coastal zone of the subcontinent after the break up of Gondwana (as discussed in section 2.1.1).

(ii) Neogene tectonics in the form of warping of the Griqualand-Transvaal axis and lowering of the Kalahari Basin have been related with the presence of the Great African Plume underneath the mantle of southern Africa (see section 2.1.1).

(iii) Isostatic adjustments related to sediment loading in the Molopo sub-basin. Isostatic rebound in response to general landscape denudation such as discussed by Bishop and Brown (1992) for the Lachlan Fold Belt in the Southeast Australian Highlands and the Murray Basin (Fig. 4.11) should also be considered between the Kalahari Basin and the Neogene interior axes of uplift. More specifically this mechanism should also be considered on a smaller scale between the Molopo sub-basin and the Griqualand-Transvaal axis of uplift. Unloading of the axis and loading of the sub-basin must have affected the total amount of basining of the Molopo sub-basin. The height difference between the lowest point of the basin and the highest point of the axis was thus not only determined by the amount of back-tilting of the proto-Molopo, but also by flexure through denudational isostatic rebound. Maximum

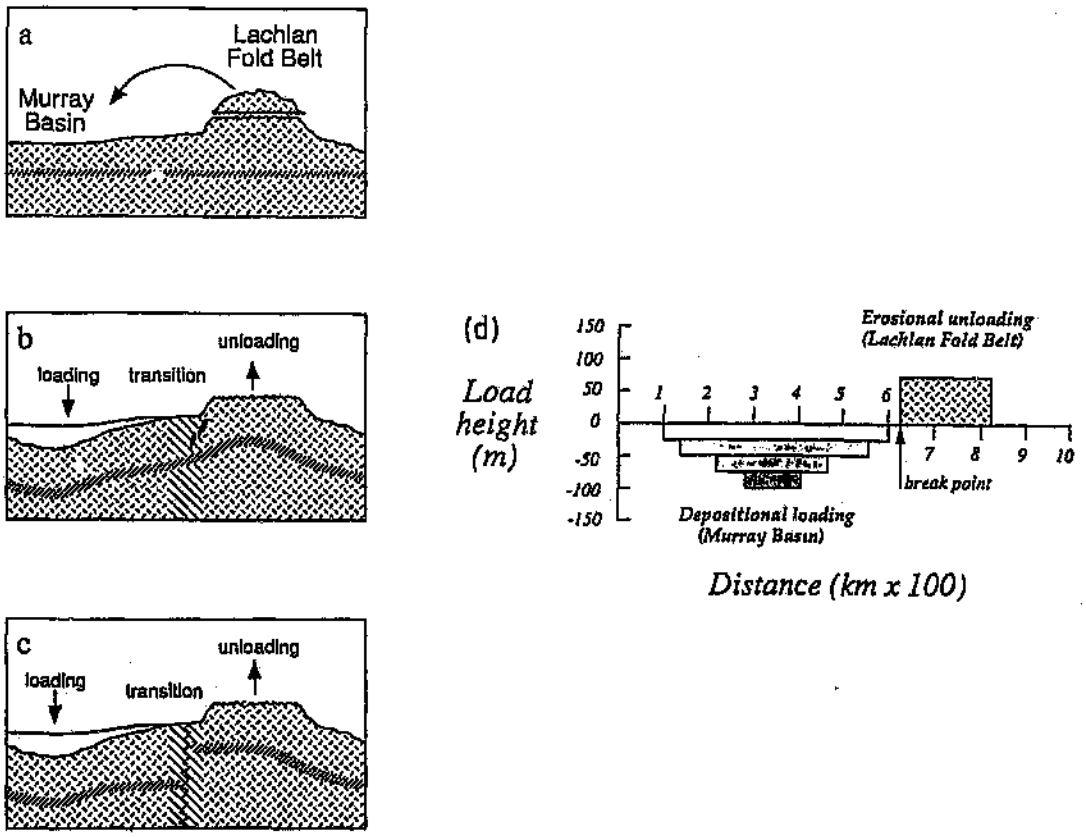


Figure 4.11 Diagrammatic representation of the principle of modelling denudational isostatic rebound for the Murray Basin (a, b, c) and distribution and magnitude (d) of loads (after Bishop and Brown, 1992).

thickness of the Kalahari Group sediments in the sub-basin exceeds 210 m in places (Fig. 4.9).

(iv) Possible lowering of the area underlain with the intruded rocks of the Molopo Farms Complex. The Molopo Farms Complex has been compared with the Bushveld Igneous Complex and shows many strong similarities (Gould *et al.*, 1987). The Bushveld complex subsided by up to 400 m during the Miocene (Partridge and Maud, 1987). Rocks of the Molopo Farms Complex underlie the thickest deposits of the sediments of the Kalahari Group in the lowest part of the Molopo sub-basin. It is speculated, that a similarly timed subsidence may have occurred in the Molopo Farms Complex area.

4.6 Implications for the Auob/Nossob Drainage Evolution

The presence of another sub-basin within the south-western section of the Kalahari Basin, underneath the central section of the present Auob/Nossob drainage was indicated by Thomas (1988). The pattern of the present Auob/Nossob/Kuruman/Molopo drainage system, superimposed on the map with the thickness variations of the Kalahari Group sediments within the Kalahari Basin, reveals some striking similarities between the Molopo/Kuruman drainage in the south-east and the Auob/Nossob drainage in the south-west. The evidence for a back-tilted palaeodrainage for the Molopo system has been discussed in this chapter. The present Molopo drainage lines cross the sediment-filled proto-Molopo at right angles in a number of places (Fig. 4. 6). A similar situation can be recognised in the Auob/Nossob system. The pattern of the isopachs of the Kalahari group sediments, similarly, strongly suggests a back-tilted, sediment-filled palaeodrainage system for the Auob/Nossob system. A proto-Auob/Nossob most probably back-tilted with reference to the southern Kalahari rim, at the initiation of the Kalahari Basin during the late Cretaceous. The drainage lines of the present Auob/Nossob system also cross the underlying Kalahari Group sediment body at near right angles (Fig. 4.12). In the case of the Molopo the drainage system has shifted, due to Tertiary warping of the Griqualand-Transvaal axis, in a westerly direction towards the NE-SW central axis of the Kalahari Basin. The present Auob/Nossob drainage system seems to have shifted, due to Tertiary

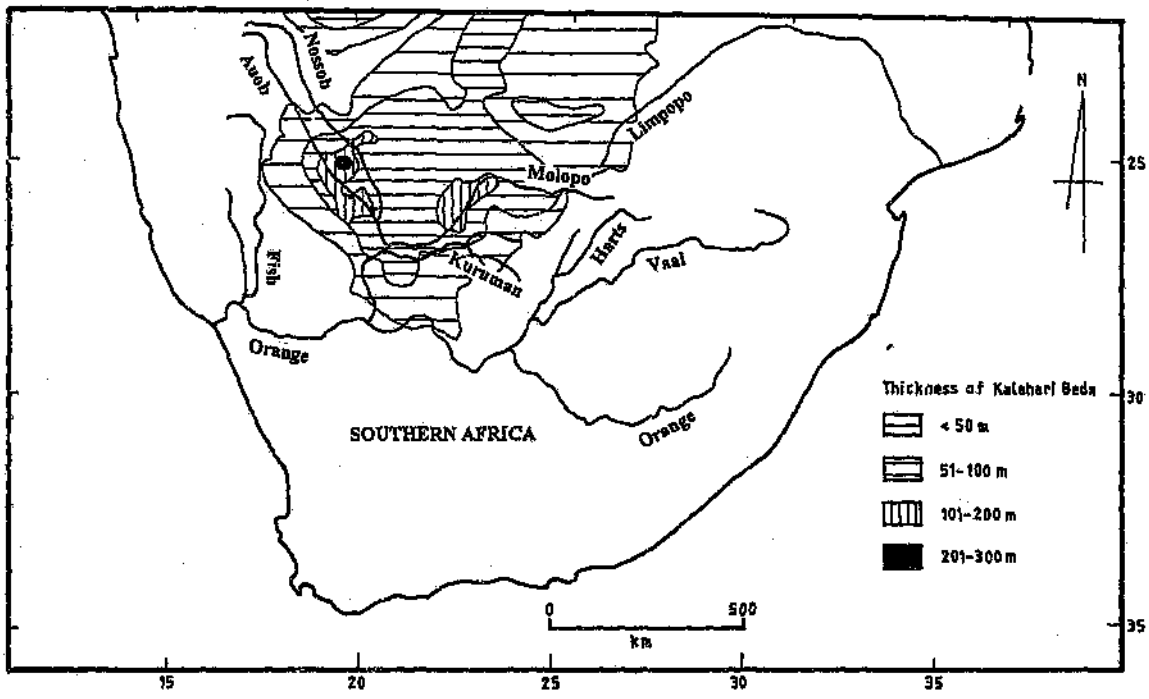


Figure 4.12 Southern Kalahari drainage and thickness of Kalahari Beds.

uplifts of the Great Escarpment in Namibia, in an easterly direction to the same central axis of the Kalahari Basin. It is suggested here, that both the Molopo/Kuruman and Auob/Nossob system were southward flowing drainage systems, prior to the formation of the Kalahari Basin. The proto-Molopo was a tributary to the upper proto-Orange (de Wit's, 1993: Karoo River), while the proto-Auob/Nossob was a tributary to the lower proto-Orange (de Wit's, 1993: Kalahari River). Lowering of the Kalahari Basin interrupted the proto-drainage lines. After filling of the Kalahari Basin both drainage systems shifted, during a late Tertiary humid climatic pulse, towards the central Kalahari axis while maintaining a north-south gradient, to find a common exit as re-established exoreic river systems across the southern rim of the Kalahari Basin.

* * * * *

It has been proposed in this chapter that the Kalahari Group body of sediments present in the Molopo drainage should be seen as a sub-basin of the larger Kalahari Basin with its own discrete stratigraphy, as determined by regional factors such as tectonics and climatic change. It is suggested that the Molopo sub-basin developed through the back-tilting of the headwaters of the proto-Molopo concurrent with the deepening Kalahari Basin. The sediment body within the sub-basin is therefore a key element in the reconstruction of the drainage evolution of the Molopo River. A similar evolution, in mirror fashion, has been suggested for the Auob/Nossob drainage. The next two chapters will provide further evidence for the evolution of the Molopo drainage from two areas situated on the eastern rim of the drainage basin. In chapter 5 the evidence from the evolution of the upper Molopo in the Mafikeng-Zeerust-Swartruggens-Lichtenburg area will be discussed, and in chapter 6 evidence concerning the proto-Molopo in Griqualand West will be presented.

CHAPTER 5 UPPER MOLOPO DRAINAGE AREA

5.1 Introduction

In this chapter indicators of the drainage evolution of the upper Molopo drainage will be presented from the Mafikeng-Zeerust-Swartruggens-Lichtenburg area. This area is situated in the north-eastern section of the Molopo drainage across the eastern rim of the Kalahari Basin, where there is evidence of interactions between the headwaters of the proto-Molopo and the proto-Harts Rivers. The upper Molopo drainage area generally has a thin and incomplete cover of Kalahari Sands, which in contrast with the Kalahari Basin, with its typical lack of exposures, allows for the easier identification of indicators of various phases of the drainage evolution. Two geomorphological features within this area have been repeatedly mentioned in the relevant literature, they are: the diamondiferous gravels and the Cainozoic Griqualand-Transvaal axis of uplift. Both play a major role in the unravelling of the history of the drainage evolution. The indicators presented here connect with the evidence presented on the Kalahari Group sediment body within the Molopo sub-basin in the previous chapter. The inset in Figure 5.1 shows the relationship between the upper Molopo drainage area and the Kalahari Group sediment body within the Molopo sub-basin.

5.2 Geology and Geomorphology of the Area

The upper Molopo drainage (Fig. 5.2) area can be subdivided into two geological regions. The western region is dominated by the sediments of the Kalahari Group, which become thinner towards the east. The exposed upper member of the Kalahari Group deposits is

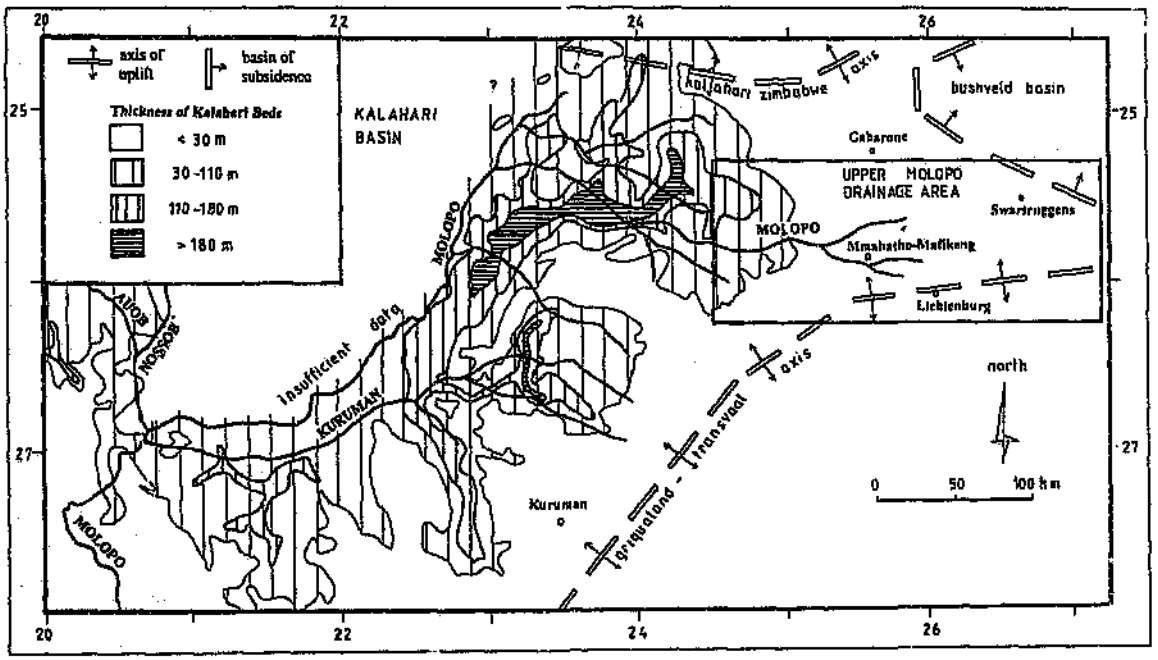


Figure 5.1 Proto-Molopo sedimentary sub-basin and tectonic structures (after Bootsman, 1997). Insert shows upper Molopo drainage area.

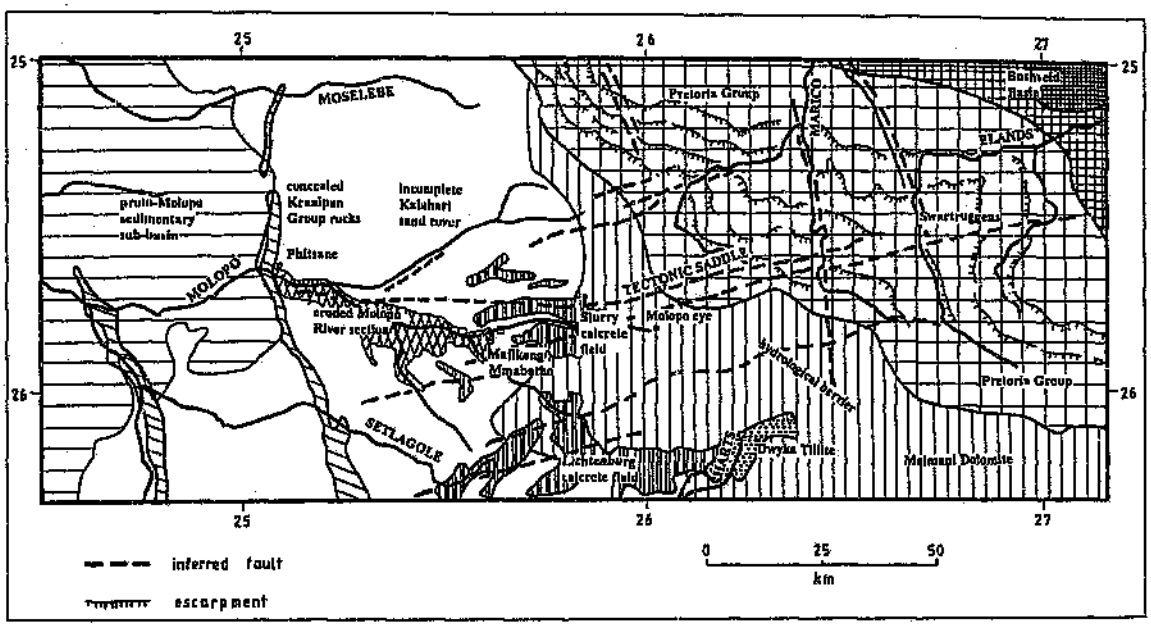


Figure 5.2 The upper Molopo drainage (after Bootsman, 1997).

referred to as the red Kalahari Sands or Gordonia Sand Formation (SACS, 1980). With the exception of some small outcrops of Archaean Kraaipan Group rocks and a stretch of the Molopo Valley on Archaean basement granites west of Mafikeng, the section has a continuous sand cover.

The area east of Mafikeng features belts of outcrops of successively younger rocks superimposed on Archaean basement granites towards the north-east. A zone of Ventersdorp Lava is followed by the Malmani Dolomite Subgroup, with the rest of the Pretoria Group and the Bushveld Basin rocks, including the Pilanesberg Complex, further towards the north-east. Near Lichtenburg there are some remnants of a suggested previously more continuous cover of Permo-Carboniferous Dwyka Tillites. There are two large occurrences of calcrete, one immediately to the east and another further towards the south-east of Mafikeng, which are referred to as the Slurry Calcrete Field and the Lichtenburg Calcrete Field respectively. Diamondiferous gravel patches (figure 5.3) further occur near Bakerville-Lichtenburg, south of Mafikeng (Setlopo, Makhubung, and Madibe), and Phitshane at the Botswana border. They are now all located at the highest local topographical positions as features of relief reversal, and appear to be aligned with the present course of the Molopo River. Three Cainozoic structural features have had a strong influence on the geomorphological evolution of the area. They are: the Griqualand-Transvaal and Kalahari-Zimbabwe axes of uplifts and the Bushveld Basin (Fig. 5.1).

The geomorphology of the western half of the area shows very limited relief and is dominated by the levelling-effect of the Kalahari Sands. The Molopo River with its shallow, wide open valley system is the major landscape feature. The eastern half is partially dominated by the flat appearance of the dolomite landscape, where the major features are shallow karst landforms and alluvial gravel ridges. The far north-easterly section is a dissected landscape of sub-parallel ridges and valleys caused by the variable resistance of the rocks of the Pretoria Group. The dolomite terrane has been interpreted as 'lowered African surface' and the far north-easterly section as 'other dissected areas'

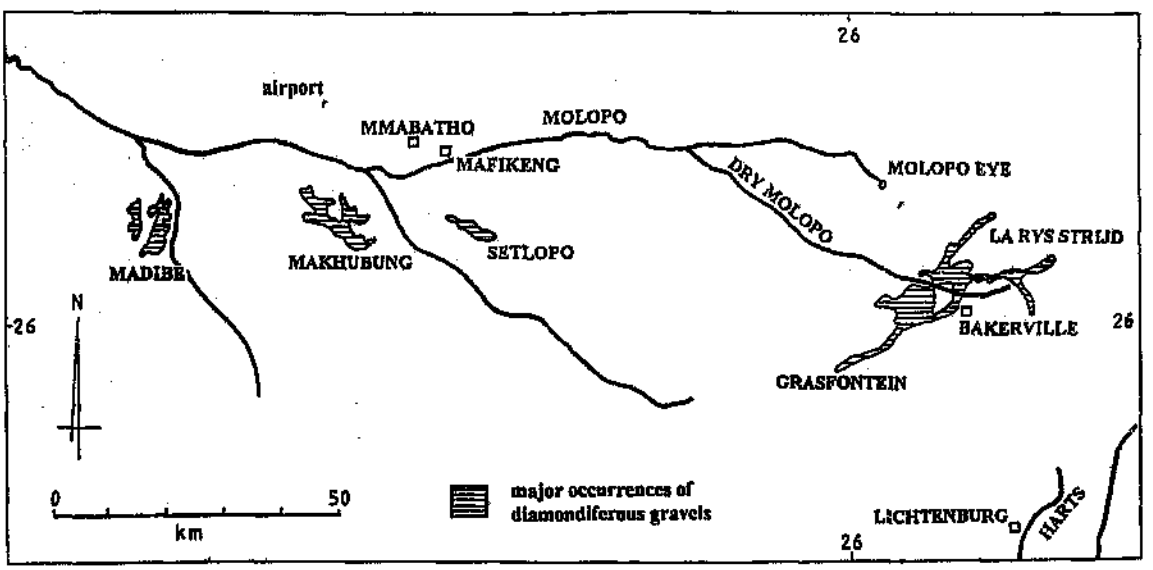


Figure 5.3 Major occurrences of diamondiferous gravels in the upper Molopo drainage area (after Bootsman, 1997).

(Partridge and Maud, 1987). The most westerly portion is dominated by the 'Cainozoic Kalahari sediments', while the area in-between the dolomite terrane and the area dominated by the red Kalahari sands, is labelled as 'undifferentiated Post-African surface'.

References to the upper Molopo drainage are generally connected with the literature on the diamondiferous gravels of the Mafikeng-Bakerville-Lichtenburg area, as well as the influences of intra-continental tectonic axes of uplift. As they are vital to the evolution of the upper Molopo drainage they are reviewed briefly below.

5.2.1 Review of the diamondiferous gravels as palaeodrainage indicators

At the beginning of this century du Toit (1907) noted that much of this area had been subjected to active river erosion, as indicated by the banks of coarse gravels between Mafikeng and Madibe. He interpreted these gravel patches as remnants of a former more extensive alluvial deposit, laid down upon a peneplain that extended from the Transvaal over Mafikeng towards the west. Wellington (1929), in his discussion of the Vaal-Limpopo watershed, suggested that the Malmani Dolomite area was previously covered by the Karoo beds, and that the rivers that were responsible for the diamondiferous gravel deposits most certainly originated on the Karoo deposits and flowed to the west into the valleys of the present Molopo River. He attributed the disappearance of the drainage on the dolomite surface to the process of 'underground piracy', a term suggested earlier by Wagner (1914). Some 'runs' (a regional name for stretches of alluvial gravels) of the diamondiferous gravels were later suggested to have been laid down by the former prolongation of the Molopo (du Toit, 1933). Subsequently du Toit (1951) discussed the influence of the dolomite on the diamondiferous gravel deposits, such as their accumulation in 'potholes' (a regional name for sinkholes). He indicated that they were deposited by southwards and south-westward flowing rivers from the area of the Pretoria Group rocks. He also suggested a connection between the Bakerville gravels and the gravels near Mafikeng and calculated a drainage slope of 1 in 240 between them.

The exceptionally flat dolomite plain with its numerous sinkholes and sinuous ridge-like deposits of river-laid sediments of palaeoriver courses which have undergone reversal of relief was described by Mayer (1973). He noted that the drainage pattern of the tributaries of the present upper Molopo River showed an almost orthogonal relationship with respect to the trend of the Harts River and suggested a genetic relationship. Mayer (1973) agreed with du Toit (1933) that the Barberspan-Gelukspan pan line, situated in a broad open valley between Mafikeng and DelaReyville, was once the former main course of the Harts River, which had been beheaded in its upper reaches by the Molopo River. Mayer (1973) concurred with du Toit (1951) that the gravel runs, interpreted as the fossilised course of an old Diepholte-Welverdiend-Grasfontein-La Rys Stryd River, represented the capture of the ancient Grasfontein River by the headward extension of a tributary of the Molopo. Mayer suggested that the drainage along the gravel run of Diepholte-Grasfontein ceased only at a very late stage in the drainage evolution, which was considered to be supported by the presence of the dry, though well-defined course of the present Dry Molopo River, which extends almost to the La Rys Stryd gravel run.

Stratten (1979) stated that the alluvial gravels along the Molopo River near Phitshane have two distinct layers. A palaeocurrent direction towards the south-east was determined for the lower layer and a palaeocurrent direction towards the south-west for the upper layer. On the basis of these observations it was determined that it was unlikely that the lower layer was laid down by the present Molopo River, but that the upper layer was fairly certainly laid down by the Molopo. For the gravels of the Mafikeng area Stratten (1979) did not give any palaeocurrent directions, but stated that in all respects they were similar to those near Phitshane along the Molopo River. He suggested a common source for the gravels around Mafikeng, Phitshane, and further south around Schweizer-Reinecke. This source was suggested to be situated somewhere in Botswana, 50 to 80 km west of Kanye. Stratten further concluded on the basis of the variation of the roundness of the Waterberg quartzite component that the diamondiferous gravels of the

Lichtenburg-Ventersdorp area were first transported by the Dwyka Transvaal ice sheet from the Transvaal in a southerly direction and subsequently reworked by fluvial forces.

de Wit (1981) emphasised that the sedimentation of the diamondiferous gravel runs of Bakerville, was greatly influenced by the karst topography of the underlying dolomite basement. The dendritic shape of the Grasfontein gravel run, the width of the major gravel runs and the size of some boulders indicated transport in large rivers. Potholes that developed in or alongside the river bed were gradually filled with sediment with a high concentration of diamonds in the basal layers. de Wit concurred with du Toit (1951), Mayer (1973), and Stratten (1979) that the gravels were initially glacially transported southward and deposited in the Dwyka Tillite and then reworked through fluvial action. Marshall (1990) contended that the gravel runs of the Lichtenburg area on the dolomite terrane, which were previously described as the courses of former drainage lines, were instead most likely to have been formed by gravel dumped into leached karst structures. Marshall also questioned Stratten's (1979) determination of the source areas for the diamonds on the basis of questionable palaeocurrent directions and pebble roundness. Marshall further indicated that the Mafikeng alluvial gravels may have been laid down as braided stream deposits during floods within an ancient course of the Molopo River. She suggested that the Mafikeng gravels had been derived from the reworking of the gravels in the Lichtenburg-Bakerville area (Bloomer, pers. comm., 1989, as cited by Marshall, 1990).

Mayer (1985), in his research on the heavy-mineral patterns in the upper reaches of the Harts River, concluded that the distribution pattern for andalusite heavy minerals reveals the existence of a portion of a palaeodispersal fan related to fluvial (proto-Harts) erosion of the andalusite-bearing hornfelses of the western Bushveld Complex. This pointed to a formerly further northward-extended proto-Harts drainage. The presence of andalusite-rich shales in the Setlopo alluvial gravels near Mafikeng have been cited as evidence, that the Molopo River system once extended much further eastwards (Brinn, 1991).

In a detailed sedimentological analysis of the Setlopo diamondiferous gravels (Brinn, 1991), three units were identified within the gravels. The lowest unit identified, was described as resembling a 'boulder clay' (unsorted clasts in a clay matrix) and stratified coarse silts and gravels, which are partly calcretised into a hardpan calcrete. The calcrete sometimes extends into the fissures of the underlying Ventersdorp andesite. Unit two was described as a widespread blanket of coarse, unsorted, heavily oxidised cobbles. Brinn indicated that oxidation is pervasive and extends several centimetres into the clasts. This description resembles McCarthy's (1983) term 'rubified gravels', the 'derived gravels' of Partridge and Brink (1967), and 'derived colluvial and eluvial gravels' of Marshall (1990). In the lower section of this unit silicified oolithic limestone cobbles are present. The Silverton Shale Formation was indicated as the source for the silicified oolithic limestone, which points to a former extension of the Molopo drainage towards the east, where the Silverton Shale Formation occurs. Distinct gully channels are scoured into the top of this unit. Unit three was described as consisting of aeolian sands, filling the channels and blanketing the topography.

Pollen analysis of clayey sediments overlying diamondiferous gravels, in some of the sinkholes near Lichtenburg has indicated that a late Cretaceous age can be ascribed to the sediments Smith, R. H. M. (pers. comm., 1994). This implies the presence of fluvial activity for concentrating the diamondiferous gravels in the area at some time during the Cretaceous. It is apparent from the review above that the diamondiferous gravels have invariably been linked with palaeodrainage lines since at least Cretaceous times, and that the Molopo has been involved in the redirection of the gravels at some stage. A mechanism for an increased influence of the Molopo River has so far not been clearly stated, and it is one of the objectives of this thesis to address this.

5.2.2 Review of the role of the intra-continental axes of uplift

du Toit (1933) was the first to discuss the effects of the Griqualand-Transvaal and Kalahari-Zimbabwe axes of uplift on the drainage system of central southern Africa. du

Toit suggested that the asymmetric Griqualand-Transvaal axis of upheaval should be seen as more due to sinking of the Kalahari Basin rather than to axial warping. He further described the Kalahari-Zimbabwe axis also as slightly asymmetrical with a steeper south-eastern slope. du Toit mentioned in addition the sinking of the Bushveld Basin and its effect on the loss of catchment of the Molopo River. The lowering of the water-table in the dolomite terrane of the Lichtenburg area resulted according to Mayer (1973) from the diastrophic uplift along the Griqualand-Transvaal axis. He suggested that the position of the Griqualand-Transvaal axis, as indicated by du Toit (1933) should, on the basis of geological and geomorphological indicators, be partly shifted eastward between Klerksdorp and Barkly West to a new position roughly parallel to the Vaal River.

de Wit (1981) stated that due to the uplift of the Griqualand-Transvaal axis the Welverdiend-Grasfontein-La Rys Stryd (Harts) tributary was redirected towards the north-west into the Molopo drainage. Partridge and Maud (1987) mentioned that the divide between the Limpopo and the Orange/Vaal drainage systems (including the proto-Harts River) must have occupied a line some 75-100 km north of its present position in the late Cretaceous times before any warping of the Griqualand-Transvaal axis. They further stated that initially, Miocene uplift along the Griqualand-Transvaal axis was confined chiefly to its western section (Northern Cape), and that its position partially differed from later rejuvenation in the Pliocene. They suggested that only as a result of Pliocene uplift the eastern part of the axis came into being for the first time. They mentioned in addition the effect of mid-Tertiary subsidence within the Bushveld Basin, of up to 400 m locally, upon the Molopo drainage. Rapid erosion of headwaters of the Limpopo River in the southern Bushveld resulted in river piracy and flow reversals.

Marshall (1990) suggested that the Griqualand-Transvaal axis be seen as the westerly continuation of the Rand Anticline. Marshall further stated that the axis may already have existed as a minor structurally controlled watershed or incipient upwarp since the Cretaceous, and that there are indications that Miocene uplift did also occur along the eastern section of the axis, contrary to the earlier opinions of du Toit (1933) and Partridge

and Maud (1987). Marshall reported that Pliocene uplift was concentrated along the same axis, but appears to have been of greater amplitude. However, Brinn (1991) postulated that rather than diastrophism along the Griqualand-Transvaal axis, the drying out of the climate during the Miocene caused the Malmani Dolomites to become a 'geographical barrier' to the extended upper Molopo drainage.

From the above review it appears that warping of the intra-continental axes played an important role in determining the palaeodrainage lines within the area. Both the diamondiferous gravels, as indicators of palaeodrainage lines, and the intra-continental axes of uplift, have been linked by the above authors with the Molopo River. It is the focus of this chapter, on the basis of a re-evaluation of those links, to present indicators to the development of the upper Molopo drainage towards an interpretative model of the larger scale development of the Molopo drainage (chapter 7).

5.3 Pre-Kalahari Drainage Indicators

The upper Molopo drainage (Fig. 5.2) area lies beyond the evidence of pre-Dwyka, Permo-Carboniferous drainage lines. There are no suboutcrops of Dwyka tillite nor exhumed pre-Dwyka valleys within the area. The earliest drainage indicators within the upper Molopo drainage area post-date the Dwyka glaciation, but pre-date the formation of the Cainozoic Kalahari Basin.

5.3.1 Proto-drainage lines and the Molopo sub-basin

Several previous authors have presented their views on the pre-Kalahari drainage in reconstructions for larger parts of the interior, that include the upper Molopo drainage area. The major drainage reconstructions that are relevant here, by Mayer (1973), Stratten (1979), Partridge and Maud (1987), and Marshall (1990), differ to some extent but show all agreement on a pre-Kalahari drainage system with dominant north-south trending segments, and favour a south-flowing proto-Harts drainage system across a major part of the present upper Molopo drainage area. The proto-Harts River dominated the easterly section of the area, while the proto-Molopo appears to have been more active in the more

westerly area. It is suggested, on the basis of the extent of the Kalahari Group sediment body in the Molopo sub-basin, that the position of a north-south trending ridge of resistant banded ironstone rocks of the Kraaipan Group near Phitshane (Fig. 5.2) indicates the divide between the ancient proto-Molopo in the west and proto-Harts in the east.

5.4 Extended Upper Molopo Drainage

The shape of the sub-basin indicates that the pre-Kalahari proto-Molopo did not extend further east than Phitshane (Fig. 5.2), but it is suggested here that the present eastwards extension of the upper Molopo River up to the Molopo Eye, and most probably at some earlier time much further east (up to 200 km), was generated by the downwarp of the Kalahari Basin in relation to the incipient intra-continental axes of uplift. It is postulated that a topographical low developed, which started to extend from the deepening Kalahari Basin towards the east, between incipient forms of the Kalahari-Zimbabwe and Griqualand-Transvaal axes of uplift (Fig. 5.1).

5.4.1 Development of the tectonic 'trough'

It has been indicated on the basis of drainage developments of the palaeo-Bloemhofspruit, that the Griqualand-Transvaal axis may have already existed during the development of a drainage network on an incipient upwarp or low structural divide since late Cretaceous times (Marshall, 1990: 150,153). Marshall thus suggested the presence of an incipient form of the axis prior to suggested Miocene and Pliocene warpings of the axis. However, evidence of an inferred southward-flowing river (proto-Morokweng River, discussed in section 7.3.2) of Upper Cretaceous age (80-75 Ma) sets a maximum age for this axis (Partridge, pers. comm., 1996). If the Kalahari-Zimbabwe axis is of the same age as the Griqualand-Transvaal axis, then this would create the possibility for a tectonic low or 'trough' to have developed simultaneously with the Kalahari Basin.

Fault systems would have probably guided the general position of this eastward expansion of the palaeo-Molopo. The term palaeo-Molopo is used here to describe the Molopo drainage from the initiation of the Kalahari Basin until its incision into its present

position. The 1 : 1000 000 geological map shows a distinct geological lineament pattern with a WSW-ENE trend between Mafikeng and Swartruggens (Fig. 5.2), which lies more or less in the centre of this suggested tectonic 'trough'. The implications of an enlargement of the Molopo catchment into an area of higher precipitation than the rest of the drainage area have to be considered. This could mean increased competence without the need of a climatic change towards more humid conditions. The association of long-term alluvial aggradation and increased catchment area associated with drainage extension has been mentioned by Maizels (1990). Changes between the sedimentation character that occur between the various Kalahari Group formations need perhaps also to be considered in relation to such catchment changes for the different sub-basins.

5.4.2 Reworking and redirection of the diamondiferous gravels

The diamondiferous gravel areas of importance are the occurrences of the Bakerville-Lichtenburg area, the Mafikeng area, and the Phitshane area at the Botswana border. The Phitshane and Mafikeng gravel runs form gravel terraces at 40-60 m above present river level. All these gravel occurrences, being former drainage ways, now occupy the highest parts of the landscape and thus represent examples of inverted relief. The source of the diamonds still appears to be a point of debate (compare: Stratten, 1979; de Wit, 1981; and Marshall, 1990), although there seems to be some agreement on the provenance of the gravels. The consensus of the researchers cited above is that the gravels were first transported from the central Transvaal either by ice or water, and that subsequently they were reworked and partially transported to the west by the palaeo-Molopo drainage. Stratten (1979) and Marshall (1990) both pointed to the similarities between the gravel runs of the Bakerville-Lichtenburg, Mafikeng, and Phitshane. A topographic low, developed in a tectonic trough, as suggested above, could have created drainage conditions where the palaeo-Molopo River extended itself eastwards, and thus pirated the upper drainage network of the proto-Harts, and subsequently reworked and redirected the transport of the diamondiferous gravels within the area. The alignment of the gravel runs of Phitshane and Mafikeng (Setlopo, Makhubung, and Madibe) along the present Molopo River, and the nearness of the Bakerville-Lichtenburg gravel runs to the present Dry

Molopo River (Fig. 5.3), can further be seen as circumstantial evidence that these gravels were transported and reworked by the palaeo-Molopo River. Chromite pebbles, which originated from the Bushveld Igneous Complex, have been found in the Setlopo gravels (Behr, S. H., pers. comm., 1996). The argument of the provenance of the andalusite and the silicified oolithic limestones found in the same alluvial gravels (Brinn, 1991) further supports this.

The age of the diamondiferous gravels has not been firmly established, but recently a substantially older age has been indicated than before. The legend of the 1: 250 000 geological map for the area (unit Qa on the sheets 2524 Mafikeng and 2526 West Rand) suggest a Quaternary age, but Smith, R. M. H. (pers. comm., 1996) has indicated, on the basis of pollen analysis, for some gravel a late Cretaceous age. Smith indicated that in some potholes in the Lichtenburg-Bakerville area diamondiferous gravels are covered with clays that contain late Cretaceous pollen. The Makhubung gravel run south of Mafikeng is underlain by at least 30 m of deeply weathered Ventersdorp Lava. An estimation of the timespan needed for the development of such a deep weathering profile might give a further indication of the age. It is further suggested that the gravel runs be considered as palaeo-environmental indicators. Raised channel systems have been used with some success as indicators of palaeo-hydrologic change (Maizels, 1990) in other arid areas. The period during which the extension of the upper Molopo drainage formed and existed, is estimated as has having started from the late Cretaceous onwards during the concurrent development of the tectonic trough with the deepening Kalahari Basin, and as having ended with the occurrence of Neogene tectonics in the area. The extended upper Molopo may thus have existed for much of the early to middle Tertiary.

5.5 Reduced Upper Molopo Drainage

The period of an extended upper Molopo drainage, which had developed at the expense of the headwaters of the proto-Harts, was ended because of at least two new developments: the lowering of the Bushveld Basin and the establishment of the Malmani hydrological barrier. Evidence fro. an area near Mafikeng indicates that calcrete

tongues extending from calcretes on the adjacent dolomite terrane represent sediment-filled palaeovalleys, where groundwater calcretes have formed preferentially. These sediment-filled palaeovalleys formed as a consequence of conditions that favoured the shortening of the palaeo-Molopo.

5.5.1 Development of the tectonic 'saddle'

It has been suggested by Partridge and Maud (1987) that the lowering of the Bushveld Basin during the mid-Tertiary (end of early Miocene ~18 Ma) greatly contributed to a 75-100 km southward shift of the divide between the Limpopo drainage system and the Orange/Vaal system. Rapid erosion of the headwaters of the Limpopo drainage caused in some streams a reversal of direction of flow and river piracy. Bishop (1995) and Ollier (1995) have discussed such examples as indicators of tectonics on landscape evolution in Australia. LANDSAT imagery of the area under discussion shows a large valley-shape near Swartruggens, as evidenced by the rock outcrops of the near-horizontal Pretoria Group formations, opening up towards the south-west. This suggests a remnant of a valley formed by a river flowing in a south-westerly direction. The present tributaries of the Elands River (Fig 5.2), that occupy this valley now flow in a north-westerly direction. The lowering of the Bushveld Basin would thus have resulted in a sharp reduction in the flow of the formerly expanded upper Molopo, and would have transformed the tectonic trough into a tectonic 'saddle'. This tectonic saddle would have the Kalahari-Zimbabwe axis of uplift to the north, the Griqualand-Transvaal axis of uplift to the south, the Kalahari Basin downwarp to the west of it, and the lowered Bushveld Basin in the north-east.

5.5.2 Malmani Dolomites as a hydrological barrier

The near-horizontal Malmani Dolomites were most probably covered with sediments of Dwyka Tillites, when the diamondiferous gravels were initially transported (Wellington, 1929; de Wit, 1981). Exposure of the dolomites from beneath the Karoo cover and lowering of the groundwater level, would have created an effective barrier (Fig. 5.2) for surface drainage from north-easterly directions. This barrier would have sharply reduced

flow from the north-east to the upper Molopo drainage, and caused the cessation of flow across the dolomites. It is suggested that a sequence of springs on the Malmani Dolomite, east of the Molopo Eye (e.g. Wonderfontein, Kalkfontein, Grootfontein, and Bronkhorstfontein), are evidence of a lowering groundwater level, and associated shortening of the course of the palaeo-Molopo. Evidence of insular fish populations in isolated karst sinkholes (Twentyman Jones, 1993) seems to support the idea of the emergence of the dolomites from beneath a Dwyka Tillite cover.

5.5.3 Calcrete tongues of the Slurry Calcrete Field

The Slurry and Lichtenburg calcrete fields are both situated on the edge of the dolomite terrane and appear to thin out into 'tongues' over the adjacent Ventersdorp Lava. Borehole data from a calcrete tongue north of Mafikeng (Martinelli and Associates, 1981) revealed that this calcrete tongue represents the sediment in-fill of a palaeodrainage channel and valley, of which the top section has been calcretised. The deepest borehole indicated a 36 m thick sediment body, of which the upper 9 m had been calcretised. The base of the palaeochannel is about 60 m below the level of the nearest gravel run, which has been associated with an the previously extended upper palaeo-Molopo. This incision could be explained as a reaction to a tectonic (Miocene and/or Pliocene?) uplift. The dimensions of the palaeovalley are similar, if not larger, than the present Molopo River valley west of Mafikeng. The nature of the sedimentary filling of the palaeovalley suggests, that the entire valley appears to have been drowned in its own sediment.

Nature of the calcrete crust

The $CaCO_3$ necessary for the development of the 9 m thick calcrete, developed in this sediment, could have originated in the nearby Malmani Dolomites as well as from the adjacent Ventersdorp Lava. Groundwater flow through the sediment would have encouraged the development of a 'phreatic' calcrete. These calcretised palaeodrainage lines closely resemble descriptions by Mann and Horwitz (1979), Arakel and McConchie (1982), Arakel (1986, 1991), and Arakel *et al.* (1989) of palaeodrainage lines in the central groundwater discharge zone of Australia. Some evidence of mounding of the

centreline and the development of twin lateral drainages along the edges of the calcrete tongues, as described by Mann and Horwitz (1979), can be seen in the example north of Mafikeng. Most of the calcrete in the palaeodrainage systems of inland Australia was identified by Arakel and McConchie (1982) as 'phreatic' calcrete, where carbonate precipitation was facilitated by the lateral movement of groundwater. It is suggested here that similar conditions existed in some of the palaeodrainage lines of the upper Molopo and that their description can be used as a model to explain the nature of the calcrete tongues.

Sediment-filled palaeovalley at Ramatlhabama

A similar sediment-filled palaeovalley with a calcrete crust is exposed in a small quarry north-east of the Ramatlhabama border post, at the international border with Botswana (Plate 5.1), and was described by Brinn (1991). This sediment-filled palaeovalley has a north-south trend and is situated at the present headwater of the Ramatlhabama tributary. The sedimentary infill (Fig. 5.4) is up to 4 m thick, thinning laterally to 50 cm over a total width of 200 m to either side of the palaeovalley (Brinn, 1991). There is a 2 m deep, 50 m wide composite channel at the base of the palaeovalley. The channel is filled with grey-green clayey silt, and has trough cross-bedding, which decreases upward in scale and visibility. Coarser lag deposits occur at the base and consists mainly of angular granite rubble with fewer siltstone and mudstone clasts (Brinn, 1991). On top of the filled channel is a 50 cm thick flood deposit of unsorted pebble clasts. Above the flood deposit are 3 metres of stratified silts and sands with distinct gravel lenses. The top of this layer is calcretised up to a depth of 2.5 m. Cut into the hardpan calcrete there are 20 to 30 cm wide gullies which are filled with unsorted gravel, and the whole sedimentary fill is covered with about 50 cm of red Kalahari sand.

Brinn (1991) interpreted the sequence as initially due to temperate pluvial conditions, and attributed the rest of the infill and calcretisation as due to increasing aridity. The erosive gullies in the top of the calcrete were related to a sudden increase in run-off, which might be associated with widespread river rejuvenation in the area. It is suggested that, although

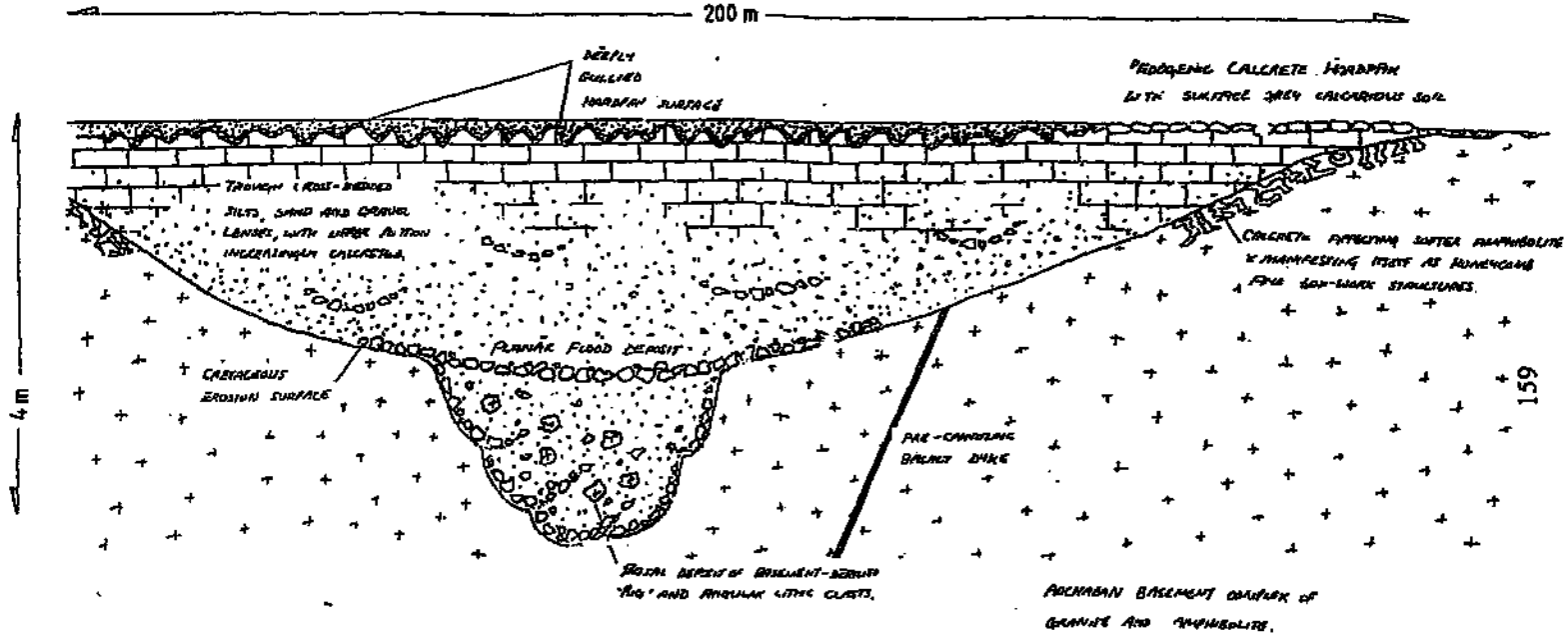


Figure 5.4 Schematic cross section of the Cainozoic deposits of a palaeovalley near Ramathlabama (after Brinn, 1991).

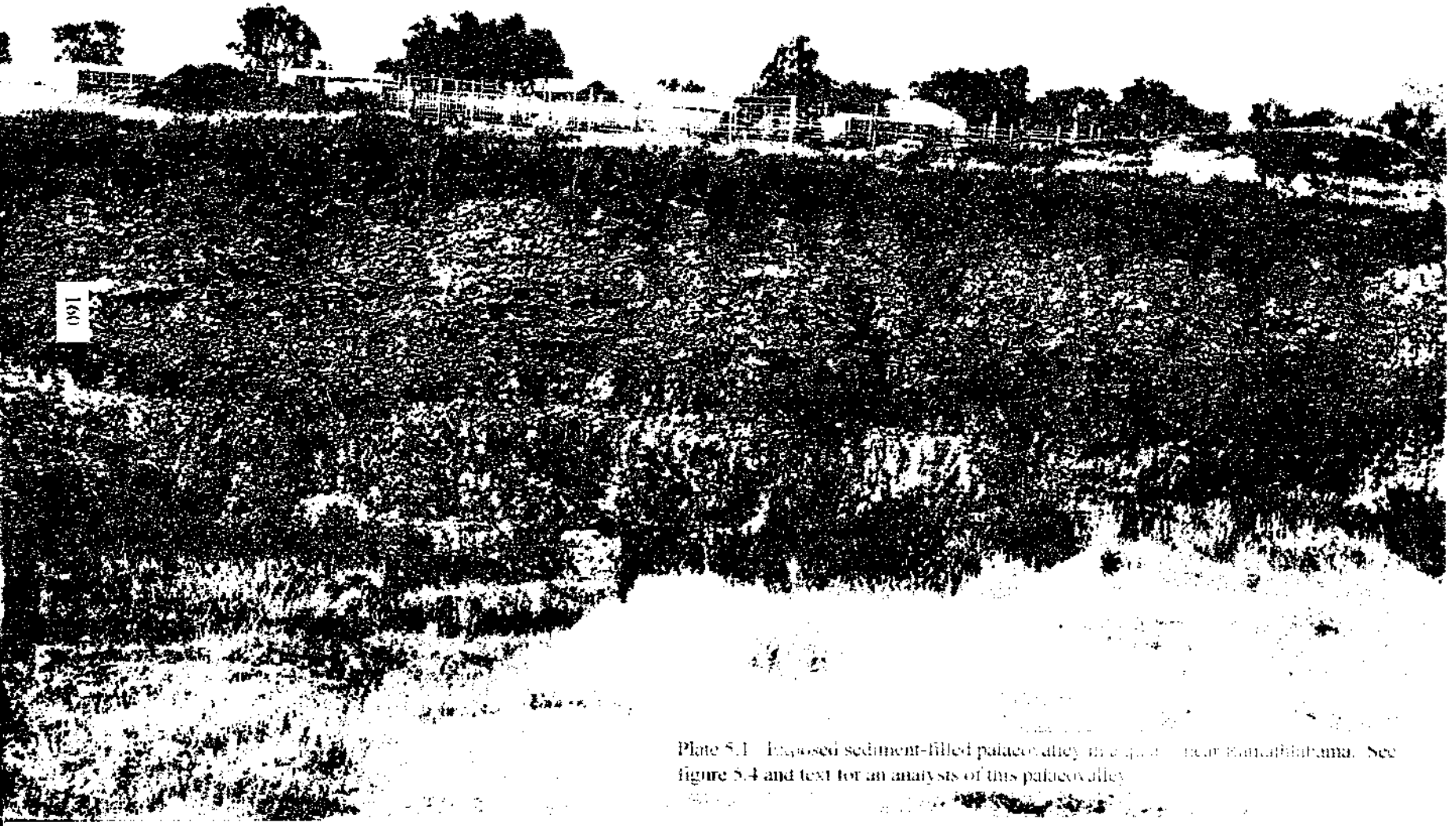


Plate 5.1 Exposed sediment-filled palaeovalley in a quarry near Kamathilahama. See figure 5.4 and text for an analysis of this palaeovalley

this palaeovalley is of much smaller dimensions than the calcretised, sediment-filled palaeo-Molopo Valley north of Mafikeng, which is only known from a few boreholes, the Ramatlhabama palaeovalley can be used as a model, to represent the sediment-filled palaeovalley of the palaeo-Molopo.

Implications of the nature of the calcrete tongue

Or · this calcrete tongue north of Mafikeng was identified as a calcretised palaeodrainage line of the upper Molopo, further scrutiny of the area around Mafikeng revealed more remnants of such sediment-filled drainage lines, seemingly 'frozen' (preserved from erosion) by thick calcrete duricrusts. Remnants of a palaeodrainage network can be identified around Mafikeng (Fig. 5.5) from the 1:250 000 geological map (unit T-Qc on map 2524 Mafikeng). The revealed drainage network indicates the calcretised, filled-in palaeochannels as they existed during the late Tertiary prior to the major rejuvenation phase, which is tentatively placed in the Pliocene. The sediments in the palaeochannel beneath the calcrete form a good aquifer, as indicated by relatively shallow groundwater depths (Knight *et al.*, 1993). This might also be the explanation for the local myth of the 'underground river' around Mafikeng.

The calcrete extending from the palaeodrainage lines into the calcrete crusts, directly on top of the Malmani Dolomite, is most probably of the *in situ* type as described by Blank and Tynes (1965). The contact between the calcrete and the dolomite can be observed in the quarries of the Slurry cement works (Michaluck and Moen, 1991). It is suggested here that this emerging picture from the Slurry Calcrete Field can also be tested at the Lichtenburg Calcrete Field.

5.6 Rejuvenation Evidence

The preserved sediment-filled palaeodrainage valley north of Mafikeng, as described above, gives evidence of at least one major phase where the drainage lines of the palaeo-Molopo could not cope with the sediment in transport, and seemed to have 'drowned' in its own sediment. A flushed-out section of the palaeo-Molopo River east of Mafikeng,

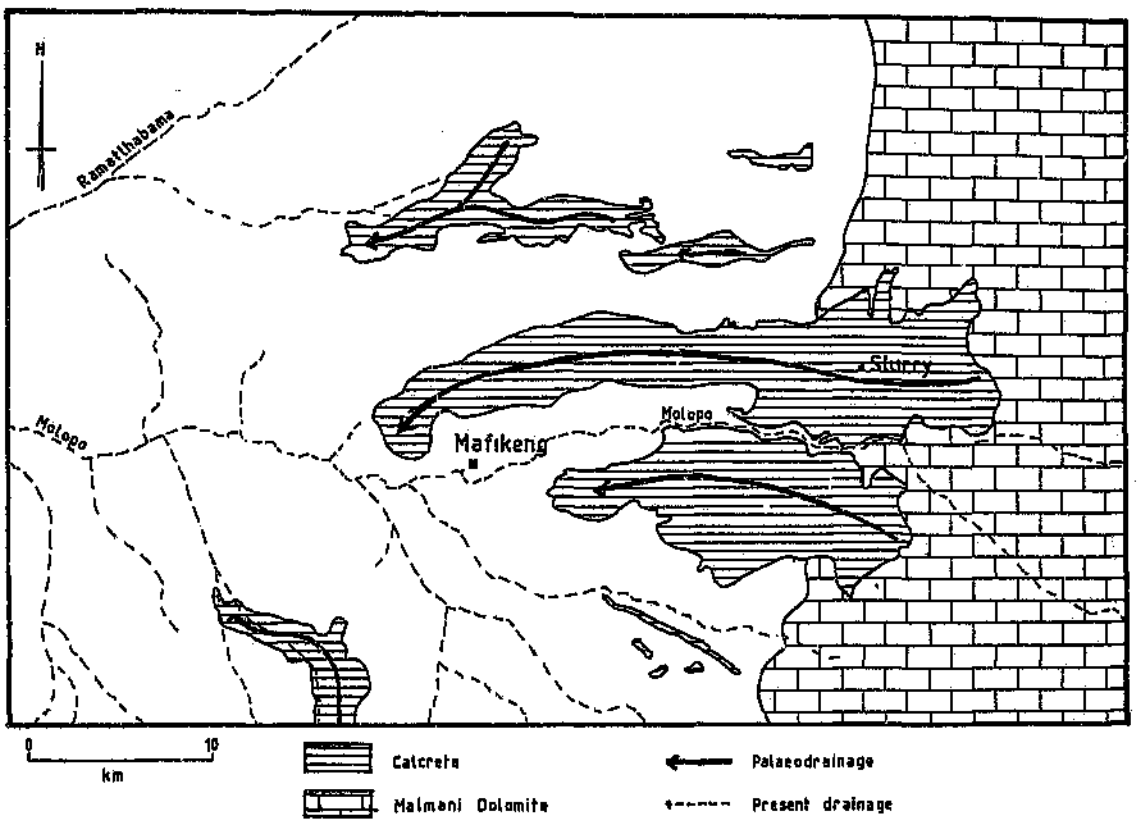


Figure 5.5 Calcrete tongues and palaeodrainage lines around Mafikeng.

the occurrence of sediment-filled hanging valleys (section 5.6.2), and its incision into the thick sediments of the Kalahari Group in the Molopo sub-basin west of Phitshane, give evidence of a major rejuvenation phase subsequent to the shortening of the upper Molopo and the sediment drowning of the upper palaeo-Molopo.

5.6.1 Flushed-out Molopo River section

The sediment-filled section of the upper palaeo-Molopo north of Mafikeng and Mmabatho (from late 1997 the capital of the former Bophuthatswana has been renamed and joined with Mafikeng) abruptly joins the course of the present Molopo River just west of Mafikeng. The sediment-filled section is considered to be the old course of the upper palaeo-Molopo. The part of the present Molopo upstream of Mafikeng lies some distance to the south of the sediment-filled palaeo-Molopo. The part of the modern Molopo downstream of Mafikeng, however, follows closely the course of the former palaeo-Molopo. A stretch of the present Molopo River of about 60 km between Mafikeng and Phitshane bears no evidence of a previous calcretised sedimentary infill, while the section even further downstream, beyond Phitshane, has been incised to about 20-30 m into the sediments of the Kalahari Group. It is suggested that a rejuvenation phase of climatic origin affected the drainage system sometime after the emergence of the hydrological barrier. The result of the rejuvenation was the flushing out of any calcretised sediments from the upper palaeo-Molopo immediately downstream of Mafikeng, as well the incision into the sediments of the Molopo sub-basin further downstream. The rejuvenation was of such magnitude that the Molopo drainage was reinstated as an exoreic system, that now connected with the Nossob River to the Orange River (Partridge and Maud, 1987). The shape of the eroded section of the Molopo River west of Mafikeng (Fig. 5.2) seems to indicate, that the erosion that re-exposed the basement bedrock, progressed from the south-east of the upper Molopo drainage area, through tributaries that were less affected by the hydrological barrier. The relatively large dimensions of these tributaries seem to support this.

The idea of a geographical (hydrological) barrier was suggested by Brinn (1991) as having been caused by climatically induced lowering of the general groundwater level. It is, however, difficult to imagine that a rejuvenation pulse, powerful enough to flush-out a stretch of 60 km of the palaeo-Molopo channel west of Mafikeng, and to cause an incision of about 20-30 m by the Molopo River into the sediments of the Kalahari Group, did not affect the sediment-filled former north-easterly section. It is therefore suggested that the hydrological barrier was created by tectonic forces, which left the inactivated calcretised palaeovalley in-fills unaffected by the subsequent climatic episode of more humid conditions.

5.6.2 Hanging valleys

A number of sediment-filled right-bank minor tributaries occur along the Molopo River immediately west of Mafikeng. Borrow pits for road works and excavations for a water pipeline from the new Modimola dam to storage tanks at Signal Hill, just outside Mafikeng, revealed a number of now minor tributary valleys, which had much larger dimensions before, but had subsequently been filled with sediments which have been calcretised to various degrees. It seems that those sediment-filled tributaries became choked with sediments at one stage. They were, however, not reactivated during the rejuvenation phase, due to a lack of runoff and now resemble minor hanging valleys. The lack of runoff might well be connected with the emerged hydrological barrier.

5.7 Quaternary Climatic Change

A number of river features testify to climatic changes which have occurred subsequent to the rejuvenation phase that resulted in the incision of the Molopo River. Remnants of rock-cut lower terraces occur immediately west of Mafikeng, and the nature of the present Molopo River is that of an underfit river.

5.7.1 Lower terraces

Two lower terraces, respectively at 6 m and 14 m above the level of the floodplain of the Molopo River west of Mafikeng were reported by Goudie (1973). These are remnants of

rock-cut terraces that carry thin covers of calcretised gravels and red Kalahari sands. These terraces are best preserved immediately west of Mafikeng but become less easy to identify towards the Botswana border. Visual macroscopic comparison of terrace gravels shows a generally decreasing sediment size from higher to lower terraces. It is suggested that a further analysis of gravel dimensions of these terraces and the higher (40 60 m) terrace gravels can be used to give a rough indication of the loss of competence during the evolution of the Molopo River. Similar lower terraces of other southern African river systems have generally been considered to be a response to climatic fluctuations during the Pleistocene.

5.7.2 Underfit river

The present Molopo River has the general characteristic of a dry river. Except for some limited episodic flows in response to severe weather systems, and very restricted spring flow in a few headwaters, the Molopo drainage at present must be considered to be an inactive drainage system. No permanent integrated flow is known during living memory. Heine (e.g. 1982) suggested that the evidence of the last perennial flows was dated between 17 000 and 15000 BP and possibly at 12 000 BP, but this evidence is questioned by other authors (section 1.1.1).

A meander-in-meander floodplain morphology west of Mafikeng, which persists for about 350 km further downstream the Molopo River into the Molopo sub-basin, as well as the presence of a several metre thick sand-clay channel back-fill, suggest a history of decreasing flow, before the Molopo drainage reached its present dry character. Large valley meanders occur for several hundred km downstream from the upper Molopo drainage area. Their wave length and amplitude are as large as four km in places, and many examples of meander cut-offs occur in the same area. Dury (1964, 1965) considered some theoretical implications of the relationships between the amount of discharge and wavelength of meanders, which could lead to an estimate of discharge of the Molopo River during the major rejuvenation phase at the end of the Tertiary. Although Dury obtained good agreement between regression predictions and modern

measurements for many rivers, several authors have, however, criticised the application of Dury's theoretical calculations. Bowler (1978), for example, noted the limited reliability of discharges calculated by regressions based on bedwith, meander length and channel cross-sectional area. He concluded that in the case of the Murray/Goulburn drainage in Australia the regression predictions could not be used with confidence, to provide anything but trends of change in past discharge regimes. Bowler further suggested that changes in the calibre and quantity of channel sediment were involved, besides variations in magnitude of flow at the bankfull stage alone. Dury's (1965) regression relationship suggests a discharge at bankfull stage of about 240,000 cubic feet (6800 m^3) per second for a meander wavelength of four kilometres. Such a strong flow of water would indicate that the Molopo River during the major rejuvenation phase would have been a most powerful river, that certainly would have been competent to cut the gorge at Khuis, as well as the gorge near the lower Molopo/Orange River confluence at Riemvasmaak. With all the uncertainties pertaining to Dury's regression values, care should be taken in the interpretation of them, but the above figure can at least indicate the trend of change from the rejuvenation phase to the present.

It appears that the competence of the Molopo drainage decreased from the major rejuvenation phase at the end of the Tertiary, and that episodic periods of perennial flow occurred during the Pleistocene, as witnessed by the lower, rock-cut terraces, in response to fluctuating climatic changes. There is, however, no certainty about the timing of this scenario, as absolute dating of events from the Pleistocene is either absent or at best uncertain.

5.8 Further Palaeoenvironmental Indicators

The lack of absolute dates for palaeoenvironmental events in the Molopo sub-basin and the Kalahari Basin at large, has been stated in this thesis. Recent work on stalagmites in the dolomite caves near Lobatse, at the north-western edge of the upper Molopo drainage area, are an exception to this. Furthermore, analysis of a pseudo-karst grike network on

an interfluvial section of the upper Molopo drainage just west of Mafikeng, may provide further scope for palaeoenvironmental reconstructions.

5.8.1 Lobatse Caves

The existence of two dolomite caves near Lobatse (Fig. 5.6) was reported by King (1951b), and additional data have been reported by Cooke (1975), Shaw and Cooke (1986), and Thomas and Shaw (1991). Most recently stalagmites have been analysed by Holmgren (1995). Thomas and Shaw (1991) indicated that both caves are connected to the surface by collapsed sink holes. Cave II has two levels and a maximum depth of about 38 m, and has extensive formations of flowstones and sinters. Holmgren (1995) reported on stable carbon and oxygen isotope analyses of several speleothems from the Lobatse II Cave, and indicated episodes with increased inputs of winter rainfall at 50 000 to 47 000, 27 000, 25 000 and 23 000 BP (towards the Last Glacial Maximum), which according to them coincided with low lake levels in the middle Kalahari. They suggested on the basis of their results, that an enhancement of the westerlies and an equatorward displacement of the subtropical convergence, led to an increased influence of the winter rainfall belt in the southern Kalahari, but to drier conditions in the middle and northern Kalahari.

5.8.2 Pseudo-karst grikes

Air photo analysis of the Molopo valley and interfluvial areas on Archaean basement granites just to the west of Mafikeng revealed a pattern of grikes (joints widened by solution). A network of grikes developed in the ancient granites is covered by a levelling layer of red Kalahari sand. The covered grikeland slopes very gently towards the Molopo River, and after heavy downpours the grikes provide a temporary network to channel the subsurface moisture. The darker tones of the sediment-filled grikes show up on air photos at the end of the rainy season. The network was exposed in a shallow sand quarry next to the University of North West. The nature of the sandy-infill of the grikes suggests that the grikes were deepening simultaneously with the thickening of the sand cover (McGregor, D., pers. comm., 1997). It seems that the joints provided inherent planes of

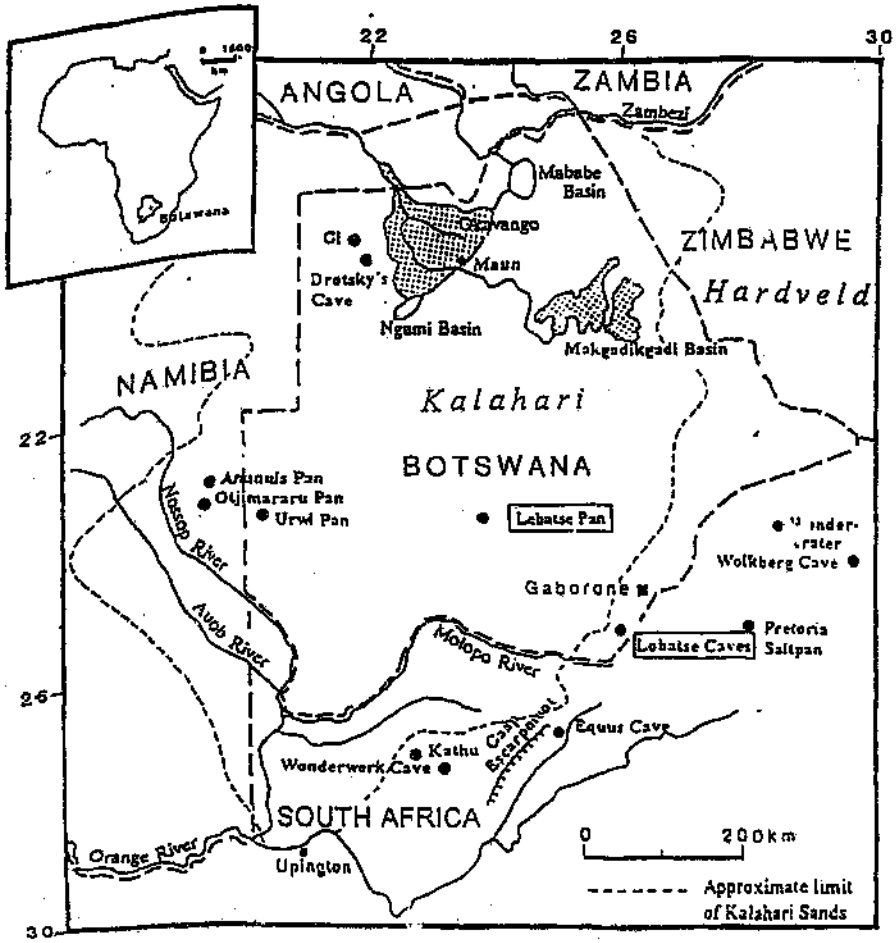


Figure 5.6 Map of central southern Africa showing the Lobatse Caves and the Lebatse Pan, and other sites of earlier palaeoclimatic studies (after Holmgren, 1995).

weakness which would channelise water flow within the rock, which in turn would promote their own growth and widening, as was suggested by Wray (1997) for grikes developed in quartz sandstones. It is suggested that the grikes are not in phase with the present climatic conditions with very low groundwater tables, but represent previously more moist conditions. The age of the grikes is very difficult to assess but, as a working hypothesis, it is suggested that deepening of the grikes progressed especially during the wetter and probably warmer episodes of the Pleistocene, after the period of major expansion of the Kalahari Sands, which is estimated to have occurred during the Pliocene.

* * * * *

In this chapter indicators toward the evolution of the upper Molopo drainage have been presented. It has been indicated that before the development of the Kalahari Basin the area was dominated by the north-south trending proto-Harts and proto-Molopo drainage lines. It was not until the development of a topographical low between the intra-continental axes of uplift concurrently with the deepening Kalahari Basin that the palaeo-Molopo extended itself by about 200 km in an easterly direction. Subsequent shrinking of the upper Molopo drainage occurred in response to the lowering of the Bushveld Basin and the development of a dolomite hydrological barrier. A major rejuvenation phase occurred prior to the Quaternary period of climatic variations. This area of the upper Molopo which is essentially situated outside the Kalahari Basin has thus provided important clues toward the evolution of the Molopo drainage. In the next chapter evidence will be provided concerning the existence of the proto-Molopo in Griqualand West as well as the consequences of the lowering of the Kalahari Basin on this ancient drainage line.

CHAPTER 6 THE GRIQUALAND WEST AREA

6.1 Introduction

In this chapter indicators to the Molopo drainage evolution from an area that lies across the present Griqualand-Transvaal axis of uplift in Griqualand West are presented. The significance of the warping of the Griqualand-Transvaal axis for the drainage systems of the interior was previously indicated by du Toit (1933). Earlier du Toit (1910) had mentioned the great antiquity and the complexity of the drainage history of the present river system of the Griqualand West area, as well as the importance of old drainage systems in the area. It is suggested that this area holds important clues to the explanation of the drainage evolution of the early proto-Molopo River. The discussion focuses successively on the nature of sediments of the Kalahari Group in the Mamatwan mine just north of the axis of uplift, the diamondiferous sediments near Mahura Muthla, the patches of river-terrace/alluvial gravels between Sishen and Griquatown, and wind gaps and gap-crossing gravels near Danielskuil, all situated upon and near the axis itself, as well as terrace sediments near the Sand River/Orange River confluence to the south of the axis. In addition, the significance of the presence of the Kathu Pan doline and Wonderwerk cave within the area will be discussed as palaeoenvironmental evidence, in relation to late Pleistocene and Holocene changes in the Molopo drainage.

The westerly section of the Griqualand West region comprises the area situated approximately between the towns of Hotazel, Postmasburg, Griquatown, and Reivilo (Fig. 6.1). It includes the south-western part of the Ghaap Plateau, the range of the combined Kuruman and Asbestos Hills, and the lower area between the Asbestos Hills and Langeberge around Postmasburg. This area is of particular interest because it lies between the proto-Molopo drainage and the Trans-Tswana drainage near Griquatown. It is suggested that the proto-Molopo, most probably as part of the Trans-Tswana drainage, flowed across the present Griqualand-Transvaal axis before it was activated towards the end of the Cretaceous, and this southward-flowing major drainage line was interrupted.

6.2 Geology and Geomorphology of the Area

The geology of the area is dominated by the rocks of the Campbell Group, which constitute the Ghaap Plateau and which are also exposed in a smaller area towards the west of the Kuruman Hills. The Asbestos-Kuruman Hills are formed by the resistant banded ironstone rocks of the Asbestos Hills Formation. They form a prominent relief feature in a shallow syncline of the stromatolitic carbonate rocks of the Campbell Group. The setting of the Ghaap Plateau is in many respect very similar to that of the plateau of the Malmani Dolomites between Mafikeng and Lichtenburg in the area of the upper Molopo drainage. The diamondiferous gravels on the farm Welgura Muthla on the Ghaap Plateau are comparable to the diamondiferous gravels between Mafikeng and Lichtenburg. The Ghaap Plateau is the oldest (Archaean) preserved giant carbonate platform in the world (Altermann and Siegfried, 1997). Towards the west the rocks of the Campbell Group are successively overlain by the younger lavas of the Ongeluk Formation and the quartzites of the Matsap Formation which form the Langeberge hills. In the north these rocks dip gently towards the centre of the Kalahari Basin, and are increasingly covered with the sediments of the Kalahari Group.

The Ghaap Plateau has been identified by Partridge (1997b) as a lowered remnant of the African surface. The ranges of the Langeberge and Asbestos-Kuruman Hills were

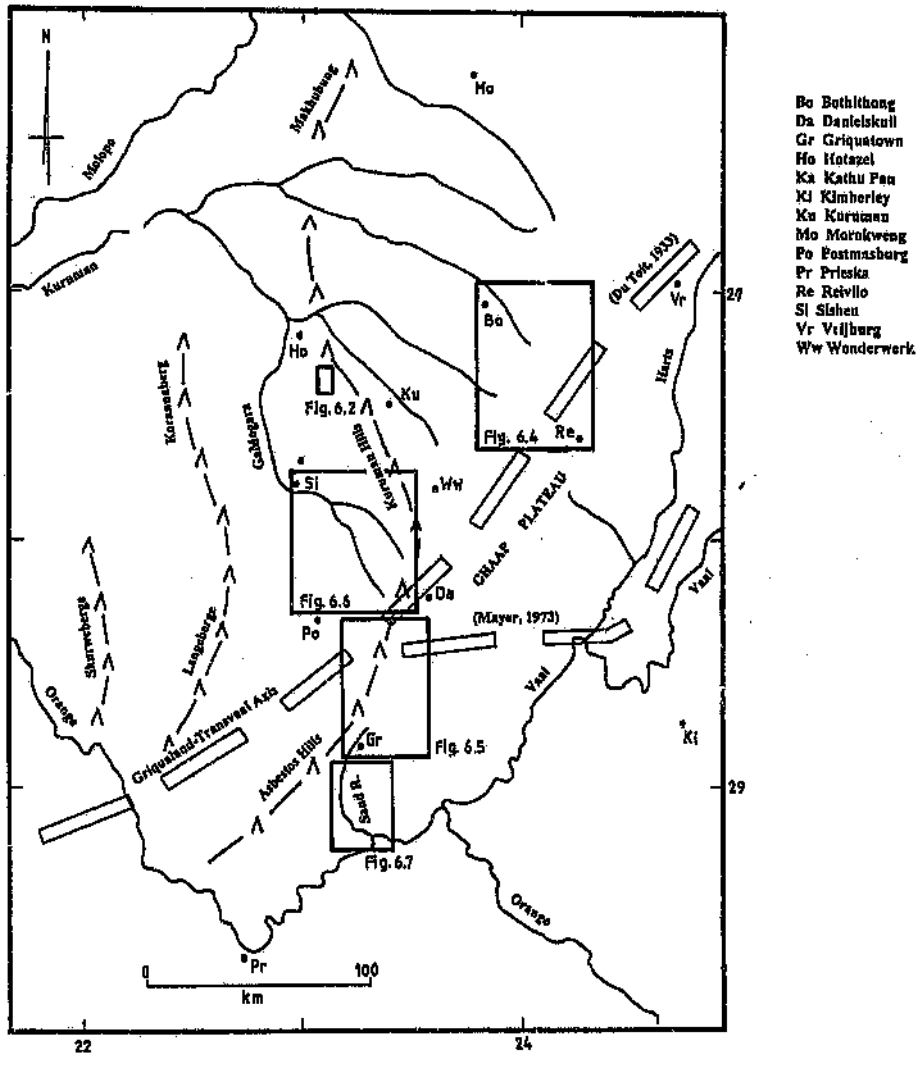


Figure 6.1 Griqualand West study area.

identified as belonging to the principal elevated areas above the African surface. The highest point of the entire Molopo drainage at 1855 m, the Gakarosa Hill, occurs in the Kuruman Hills just south of the Wonderwerk cave. The African surface and elevated areas are surrounded by the terrain of the Post-African I surface, with the flat landscape dominated by the sediments of the Kalahari Group in the north-west and the dissected areas of the Vaal-Harts and Orange Valleys in the east and south.

6.3 Kalahari Group Sediments in the Mamatwan Mine

It has already been mentioned (section 4.2) that the study of the geology and geomorphology of the Kalahari Basin has been seriously hampered by the general lack of exposure of the stratigraphy of the sediments of the Kalahari Group. The Mamatwan manganese mine of Samancor, about halfway between Kuruman and Hotazel, however, provides an unique opportunity to observe a section of the stratigraphy of the Kalahari group sediments within the Molopo sub-basin, which is otherwise mainly known from borehole descriptions. The Mamatwan mine is situated in a topographically low, flat area in between the Kuruman Hills in the east and the Korannaberg in the west. The mine is located on the eastern margin of the Dwyka Tillite and Kalahari Group sediment-filled 'Hotazel Valley'. The Mamatwan mine is more specifically positioned at the southern extremity of the Kalahari Manganese Field (Samancor, undated), which is associated with a larger structural basin further to the south, known as the Dimoten Syncline. This manganese field extends for about 40 km along a north-south axis and between 5 to 15 km in an east-west direction, and plunges gently with the Kalahari Basin towards the north-west. Manganese bodies occur within the Kalahari Manganese Field as sub-horizontal stratabound horizons situated within the banded ironstones of the Voelwater Formation of the Olifantshoek Group. The manganese ore body within the Mamatwan mine is overlain by up to 15 m of banded ironstone and 30 to 50 m of Kalahari Group sands, calcretes, gravels and clay (Samancor, undated; Anthony, 1996). The Mamatwan mine (Plate 6.1) is a conventional open cast mine, where the overburden is removed to uncover the manganese ore body (Fig. 6.2).

6.3.1 Kalahari Group stratigraphy in the Mamatwan mine

The sediments of the Kalahari Group of the northern section of the Mamatwan mine have been reviewed briefly by Williams (1993). Williams described three benches of the Kalahari Group overburden: a sand bench, a top lime bench, and a mid-lime bench (Fig. 6.3). A lower lime bench, which he did not indicate for the north wall, occurs in the south-western corner of the mine, but has broken up through the development of rotationally slumped blocks. The unstable red-brown clay beneath the lower lime bench caused its slumping. The lime benches are made up of carbonate cemented (calcretised) sediments. The clay beds underlying the lower lime bench, in the south-western area of the mine have been described cursorily and analysed by Anthony (1996). The description of the material of the sand bench resembles that of the Gordonia Sand Formation, the lime benches are similar to those of the Eden Sandstone Formation, while the basal clays are similar to that of the Budin Clay Formation of the Kalahari Group, as defined by the SACS (1980).

In his description of the overburden stratigraphy Williams (1993) firstly described:

(x) A 4 m thick sand bench, as consisting of loosely consolidated, medium grained quartz sand with an outer orange-red coloured staining (Plate 6.2). The carbonate cementation of the red sand increases with depth and causes a paler grey colour.

Williams further subdivided the top-lime bench into 4 geological units:

(ix) An upper 2 m thick homogeneous featureless cream coloured calcrete horizon. Excavations in 1997 exposed grey-green cemented sediments at this level in the northern section of the mine and reddish cemented sediments in the south-easterly section.

(viii) A 0.5 m thick brown-orange palaeosol or iron pan with prominent essentially horizontal laminations and contortions due to expansion and contraction.

(vii) A mottled calcareous sandstone unit, 1.5 m thick. The matrix is a fine to medium grained pale-orange sandstone with a carbonate cement. The matrix is mottled by



Plate 6.1 Overview of the Mamatwan manganese mine. The Korannaberg is visible in the distance. Note the flatness of the Kalahari 'sediplain'. More than 30 metre of the Kalahari Group overburden is exposed here on top of the manganese. the mobile clay 'tongues' on top of the manganese have caused the collapse of a lower lime bench.



Plate 6.2 Red Gordonia Sand Formation exposed in the Mamatwan mine. OSL samples taken from the base of the sands overlying calcrete were dated at 53 000 BP.

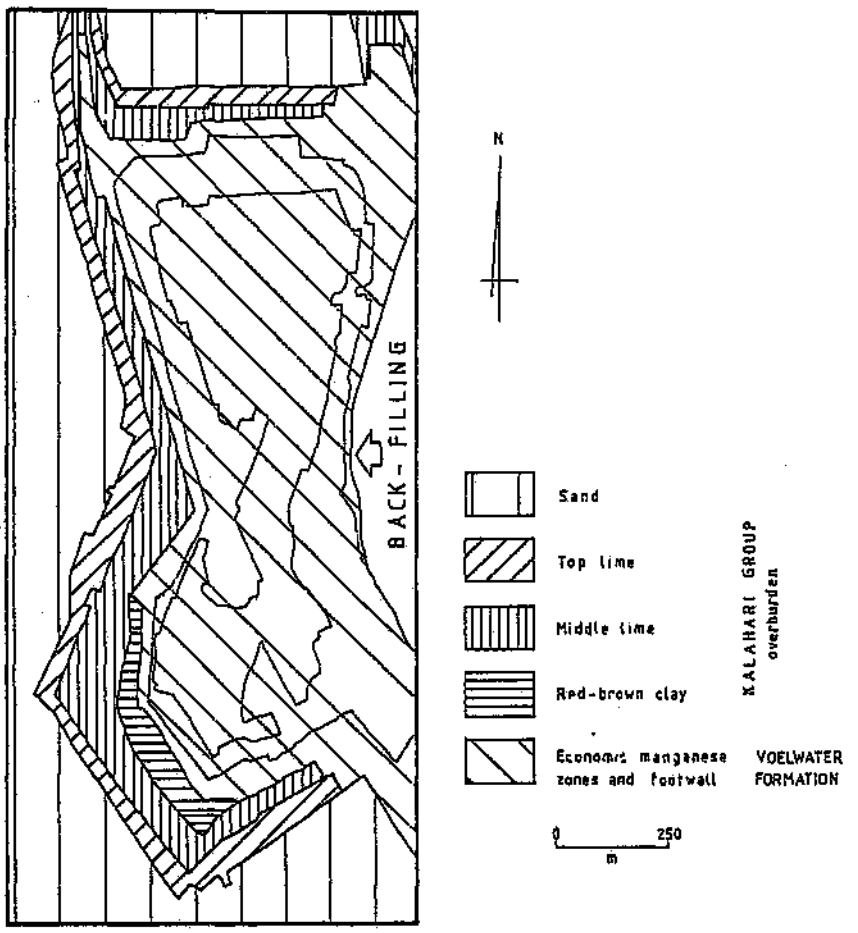


Figure 6.2 Simplified face plan of the Mamatwan manganese mine (after Samancor, undated).

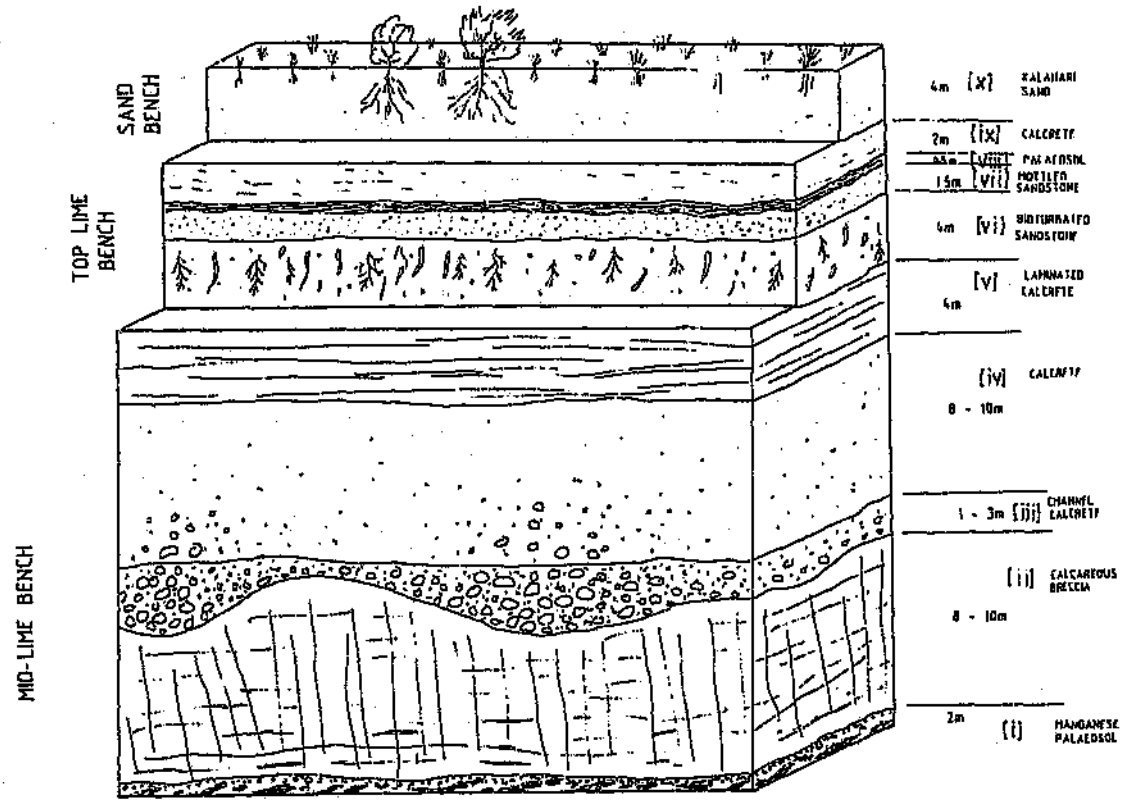


Figure 6.3 Overburden at the north wall of the Mamatwan mine (after Williams, 1993).

red-orange stained "clasts" or patches of 0.5 to 2.0 cm in size and contain no carbonate cement.

(vi) A 5 m thick orange coloured fine grained sandstone with calcareous cement. This unit is bioturbated by rootlets which have been replaced by pale pink-white calcrete.

The mid-lime bench was subdivided by Williams (1993) into another 5 geological units.

(v) A 4 m thick cream-pink coloured calcrete with numerous narrow horizontal laminations of crystalline calcrete within a calc-arenite matrix.

(iv) An 8-9 m thick cream coloured calc-arenite with fine to medium sized quartz grains and occasional small rounded clasts of banded ironstone. It contains sub-horizontal laminations and lenses of cream to colourless crystalline calcrete. Thin lenses of white crystalline calcrete occur occasionally.

(iii) A unit varying in thickness from 0.5 to 2.5 m with a sharp erosional base with a channel lag of medium to large rounded clasts of banded ironstone and manganese (Plate 6.3). Williams (1993) suggested that although the lag is now matrix-supported due to calcretisation it was once a clast-supported lag. It was interpreted as a palaeo-channel deposit, with the majority of the clasts concentrated in the thicker parts of the unit or the channels *senso stricto*. The thinner parts of the unit represent over bank deposits. This unit grades up to unit (v).

(ii) A pale pink coloured calcrete breccia which ranges from 8 to 10 m in thickness. The major constituent of this unit is a pale pink coloured calcareous material of a clay-silt grain size which has been altered to a very soft, low density rock. The matrix of the unit is brecciated by a stockwork of sub-horizontal and sub-vertical white calcite veins. Small rounded clasts of manganese and banded ironstone are present in the lower 2 m of the unit. The clasts exhibit concentric, onion ring alteration and are also of a significantly lower density. The very low density, and very fine grained nature of the material is due to a high sepiolite clay mineral content Williams (1993). It was further suggested that the alteration event was

probably associated with palaeo fluid movements far removed from the current low water table in the area. In the 1997 exposures the calcrete breccia in the south-easterly section of the mine was silcretised and extremely hard to break into very sharp angular pieces upon hammering.

(i) A 2 m thick unit directly overlying the manganese in the northern section of the mine which is interpreted as a palaeo-weathering horizon or palaeosol modified by later calcrete growth. Large and small rounded and angular clasts of manganese are enclosed by pale pink calcrete veins and calc-arenite. The clasts increase in size and angularity downwards. This unit is not present in the south-western section of the mine, where instead a red clay unit is found in between unit (ii) and the underlying manganese ore body.

Red-brown Mamatwan clay body

During the mining of the manganese ore body a clay body was encountered in the south-west section of the mine (Anthony, 1996). In that section a 'reddish-brown' (Munsell code: 2.5YR 4/4; 3/4 moist) clay body lies directly above the manganese ore body and is the lowermost unit of the Kalahari Group. The problematic nature of the clay resulted in the temporary cessation of mining operations in that area of the mine. The problem being that the clay migrates readily towards the area of least overburden pressure (which is the mined area of the pit) at a rate of almost 70 mm per day (Plate 6.4). Borehole data within the potential mining area has revealed that the thickness of the clay unit varies between 0.5 and 27 m. The clay body generally thickens towards the west. The structure of the dry hard Mamatwan clay changes dramatically when exposed to rains and becomes soft and powdery when dry.

Anthony further suggested, on the basis of the morphology of the clay body, that it had been deposited in a basin-type structure, supposedly a lake, which was fed by two rivers. The river system apparently flowed in a south-westerly direction into the lake. Anthony (1996) indicated on the basis of comparisons of geochemical analyses and the flow directions of the palaeo-rivers that the clays had been derived from the weathered



Plate 6.3 Gullies in the Eden Sandstone Formation at the north wall in the Mamatwan mine. The sediment in the gully was originally clast-supported but due to heavy calcretisation has become matrix-supported.



Plate 6.4 A tongue of the Budin Clay Formation at the south-west wall in the Mamatwan mine. The red montmorillonitic clays are extremely mobile and a hazard to mining operations. See Plate 6.1 for the stratigraphic position of the clays.

materials of the nearby lavas of the Ongeluk Formation. The major element analyses of the Mamatwan clay and the Ongeluk Lavas are very similar (Table 6.1).

Table 6.1 Comparison of analyses of Mamatwan clay and Ongeluk Formation. Source: Anthony (1996).

	Mamatwan clay	Ongeluk Formation
	(WT%).	(WT%)
SiO_2	62.53	55.80
TiO_2	0.92	1.00
Al_2O_3	14.51	14.46
Fe_2O_3	9.71	12.43
MnO	0.94	0.18
MgO	5.16	5.67
CaO	5.33	6.80
Na_2O	0.04	2.42
K_2O	0.85	1.26
P_2O_5	0.10	0.15

Although Anthony (1996) suggested that the existence of the clay body was restricted to the area of the Mamatwan mine only, and does not occur elsewhere in the Kalahari Manganese Field, it has been indicated by Smit (1977) that extensive occurrence of similar clays can be found in the deeper parts of the larger Molopo sedimentary sub-basin (section 4.5.1). It seems most likely, that the clay body thickens further west of the Mamatwan Mine, to connect with the extensive clay body described by Smit.

6.3.2 Interpretation of the Kalahari Group stratigraphy in the Mamatwan mine

The lowest geological unit of the Kalahari Group stratigraphy exposed in the south-western area of the Mamatwan mine, the red-brown montmorillonitic clay was most probably deposited in a lacustrine environment, which existed north of the Griqualand-Transvaal axis uplift. This lacustrine environment may have existed over a length of about 400 km in the back-tilted section of the proto-Molopo. The red-brown colour

indicates an oxidising environment, where the sediments would not be submerged at all times. Such conditions exist in some contemporary playa environments (Blatt *et al.*, 1972).

The lowest geological unit in the northern area of the mine as described by Williams (1993), the palaeo-weathering or palaeosol (i), could be the lateral variant of the clay body at the times of sedimentation in the lacustrine environment. The next unit, the pale pink calcrete breccia overlying the palaeosol and playa sediments would indicate a coarsening upward in the sediments, where coarser-grained distal alluvial fan environments encroached upon the playa clays. They were subsequently heavily calcretised and silcretised under conditions with fluctuating groundwater tables, which were generally much higher than at present.

Units (ii) through (v) generally present a fining upward sequence after an erosional phase, that had started with the deposition of the channel lag in the palaeo-channels. The cause for this erosional phase could have been tectonic and/or climatic. The sediments are heavily calcretised.

The units of the top lime bench could indicate the existence of pan environments as currently found in the southern Kalahari, and the red sand of the sand bench represent the aeolian reworking of the earlier deposited sand. The red colour of the sand is due to rubification. A single OSL date at the base of the red sand indicates an age of 53 000 BP (dated by: Sheffield Centre for International Drylands Research), and would represent the latest period of reworking in this eastern marginal area of the southern Kalahari. This date is much older than those of the generally drier south-western section of the Kalahari, where aeolian reworking occurred more recently (Thomas, D. S. G., pers. comm., 1997).

The present flat area between the parallel ridges of the Kuruman Hills and the Korannaberg (65 km apart) is an indicator of previous environments of deposition. Skirts of rubble and scattered scree (Moen, 1979a, b), in the form of coalesced alluvial fans and

the presence of the extensive buried clay body in between the ridges suggests a former, more extensive playa environment with a range of alluvial fan environments surrounding the playas.

In summary the sediments of the Kalahari group exposed in the Mamatwan manganese mine indicate evidence of an initial lacustrine environment north of the Griqualand-Transvaal axis of uplift. With further infilling of the Molopo sedimentary sub-basin, coarser-grained material was deposited, in which at least one major erosional phase can be identified. The upper section would indicate distal alluvial fan environments with evidence of strong calcretisation and aeolian reworking at the top. The coarsening upward sequence should be seen as the encroachment of the coarser, marginal facies on top of the playa clays, during the filling-in of the lacustrine environment.

6.4 Alluvial Sediments near and across the Griqualand-Transvaal Axis of Uplift

Alluvial gravels can be found at various places across the present Griqualand-Transvaal axis of uplift, as indicators of fluvial activity in an area where there is no such activity or very limited fluvial activity today. Diamondiferous alluvial gravels are found for example near Mahura Muthla on the Ghaap Plateau, patches of river-terrace gravels occur along a stretch between Sishen and Griquatown, as well as 'gap-crossing' gravels in a wind gap, across the Asbestos Hills near Danielskuil.

6.4.1 Mahura Muthla diamondiferous gravels

du Toit (1910) mentioned the presence of *in situ* water-worn gravels between Kuruman and Vryburg on the Ghaap Plateau at an altitude of 4 750 feet (~ 1450 m). His description of the location of these gravels agrees with the position of the gravels at Mahura Muthla. Wagner (1914) provided a few years later a brief discussion of the alluvial deposits of the Ghaap Plateau, and more specifically on the Mahura Muthla farm about 65 km east of Kuruman (Fig. 6.4). He described them as a 'lime-cemented diamond-bearing gravel' as well as a 'calcareous conglomerate'. Wagner further described the gravel, as occupying an ancient river channel with a NW-SE trend,

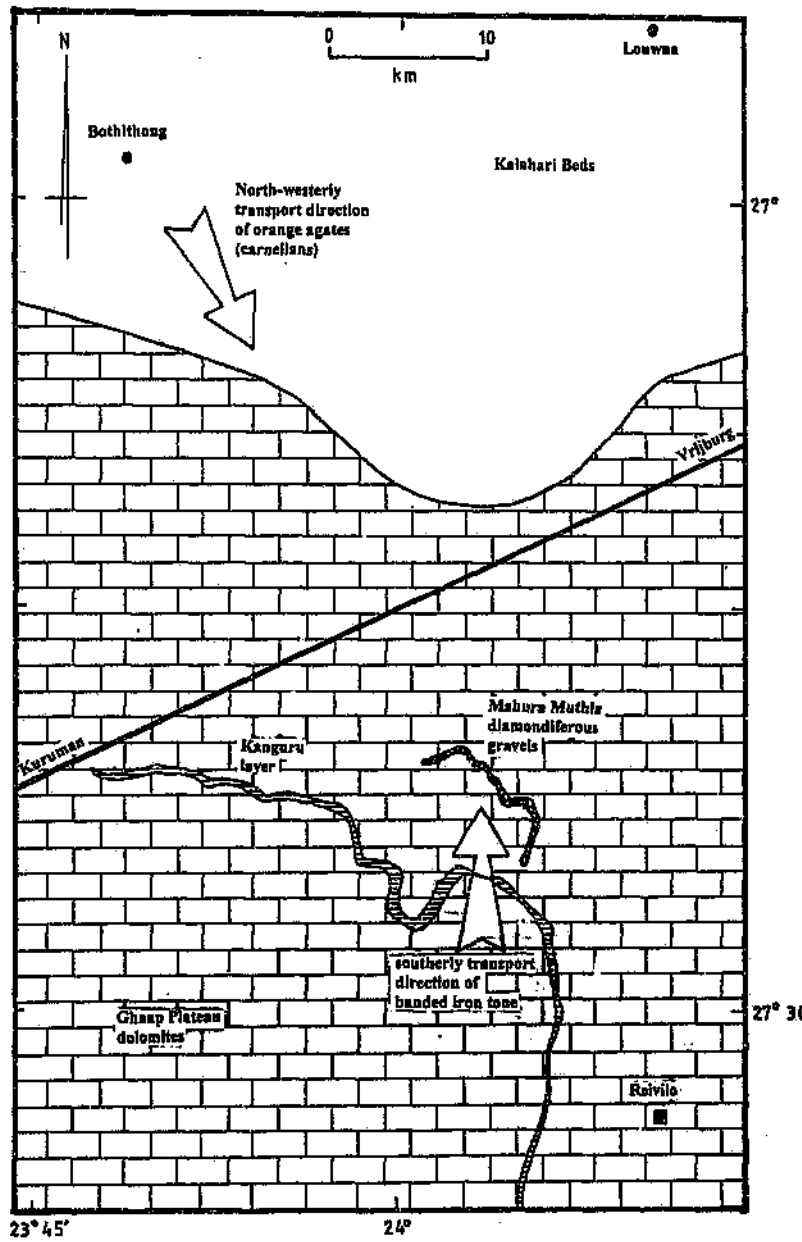


Figure 6.4 Mahura Muthla diamondiferous gravels on the Ghaap Plateau.

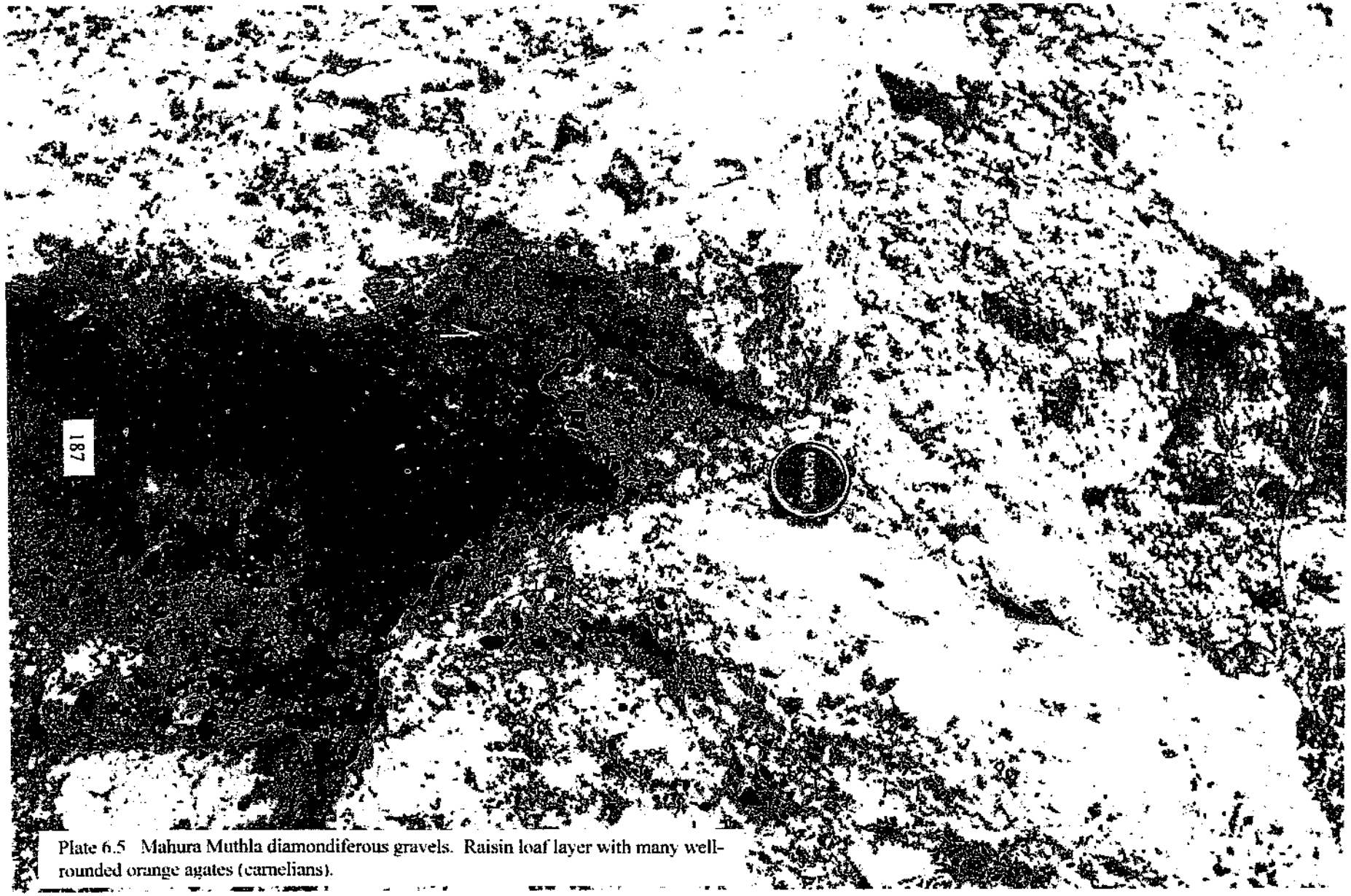


Plate 6.5 Mahura Muthla diamondiferous gravels. Raisin loaf layer with many well-rounded orange agates (carnelians).

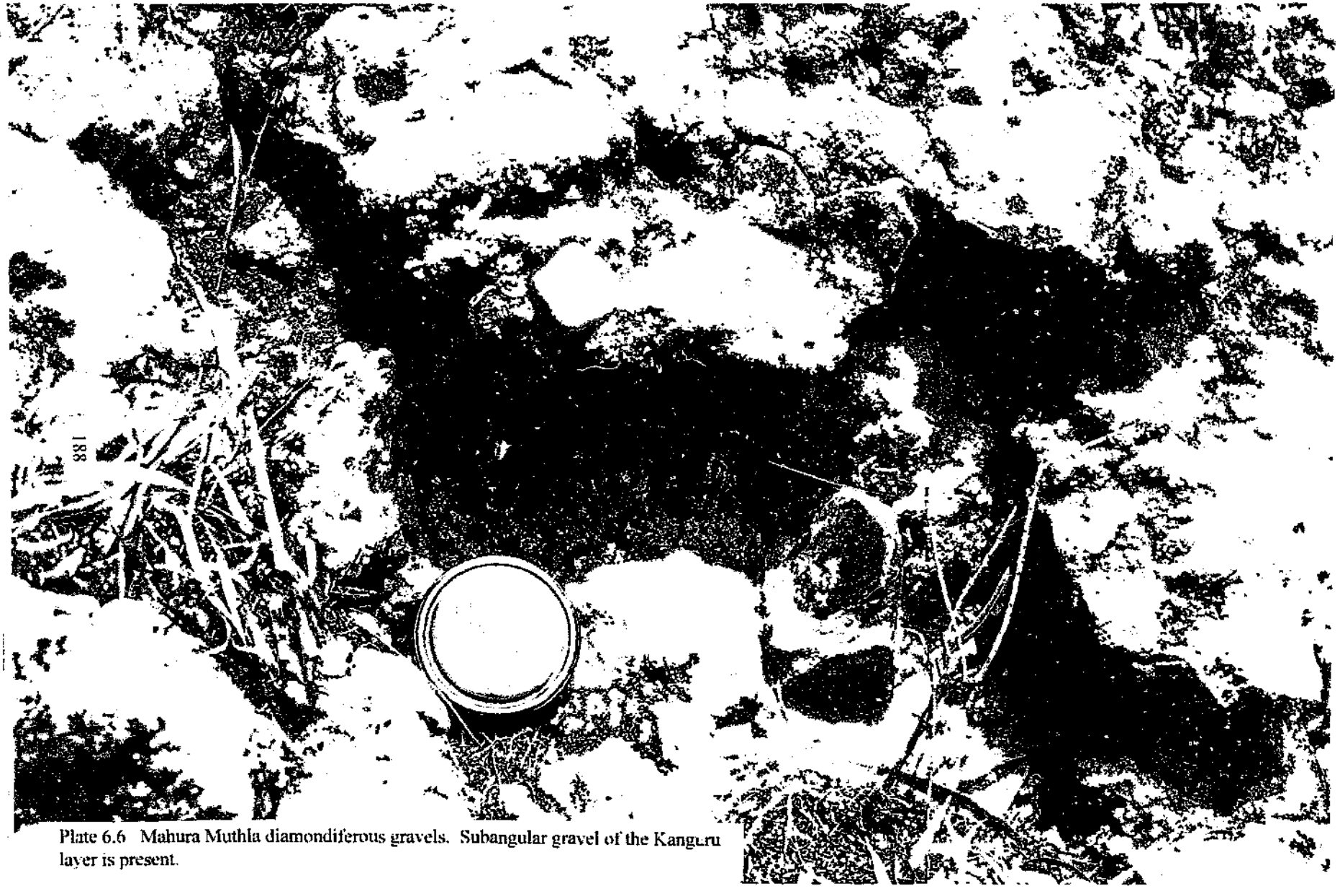


Plate 6.6 Mahura Muthla diamondiferous gravels. Subangular gravel of the Kangu.ru layer is present.

developed into the Ghaap Plateau dolomite bedrock. According to him the depth of the channel in places exceeds 9 m, with the diamonds confined to a so-called 'raisin-loaf' layer about 3.5 m thick, in the lower part of the deposit. Besides the diamonds the conglomerate consists of carnelian, agate, chert, jasper, and silicified wood pebbles. Wagner suggested that the diamonds derived from a neighbouring kimberlite pipe or dyke. Although around 1911 the diamond production had been quite sizeable, it diminished quite rapidly, as the exploration of the conglomerate was hampered by the fact that it had to be blasted with dynamite and then broken with hammers, before the washing and sieving processes. Similar deposits of lime-cemented gravels occur on the neighbouring Graspan farm. Schutte (1994) provided a similar description of the diamondiferous gravels of the Ghaap Plateau. The distribution of the gravels on the Mahura Muthla farms displays a conspicuous sinuous pattern (Fig. 6. 4).

The Mahura Muthla Formation was described by Partridge *et al.* (in prep.) as a series of channel fills over a distance of 4.5 km. The channel width averages 150 m, and the channel is filled with strongly calcretised alluvium, with alternating layers of calcified pebbles ('raisin loaf') and calcretised cobble conglomerate. The raisin loaf (Plate 6.5) has a high concentration of well-rounded orange agates (carnelian), which are derived from the Ongeluk lava. The nearest occurrence of the Ongeluk Lava occurs underneath a cover of the sediments of the Kalahari Group about 95-145 km to the west-north-west. The cobble conglomerate (Plate 6.6) contains some occasional logs of silicified wood, as well as subangular to subrounded jasper cobbles of the Kanguru layer, which occur immediately to the south of the Mahura Muthla Formation. One of the logs has been analysed by Dr. M. S. Zavada of the University of South-Western Louisiana to be a *Protoatherospermoxylon rennei,* which is which is an angiosperm, correlated with the Upper Cretaceous The Mahura Muthla Formation is estimated to be of a Santonian-Campanian (~80-75 Ma) age (Partridge *et al.*, in prep.). The provenance of the carnelians indicates a southward flow of the river system which is supported by evidence from sedimentary structures. The southward flow component as indicated by the presence of subangular cobbles from the Kanguru layer, is interpreted as evidence of

flow-reversal after initiation of the Kalahari Basin in relation with an incipient form of the Griqualand-Transvaal axis (Fig. 6.4)

The occurrence of the diamondiferous gravels across the Griqualand-Transvaal axis of uplift is very significant, as it indicates the former presence of drainage during the Upper Cretaceous across the present axis, which in this place presently constitutes the divide between the Molopo drainage and the Harts drainage to the east of it. The evidence of reverse flow further points to the influence of the incipient axis.

6.4.2 River-terrace/alluvial gravels between Sishen and Griquatown

The gravels between Sishen and Griquatown are generally rubified, red-stained gravels with an average size of about 10 cm and some occasionally larger well rounded cobbles of about 30 cm in diameter. On the various geological maps of the area they have been referred to as river terrace gravels (Visser, D. J. L., 1958; Moen, 1977a, b) and alluvial gravel (Visser, no date). du Toit (1910:354) expressed the importance of these gravels, because of their *in situ* nature, and the provenance of water-worn brown jasper pebbles in the gravels. du Toit indicated that these pebbles could only have derived from the "Griqua Town Hills" (a name that was used for a section of the range of hills between the southerly Asbestos Hills and the more northerly Kuruman Hills, near Danielskuil) and the Doornberg Hills south of the Orange River. He further mentioned that the durable brown jasper pebbles are not found in the Dwyka Tillite in the area under consideration. The jasper pebbles have their source area in the banded ironstones of the Griqua Town Hills. Jaspillite is associated with the occurrence of banded ironstone. The term banded ironstone is used in South Africa, while the synonym banded jaspilite is used in some other countries (e.g. Australia: Visser, W. A., 1980). The occurrence of the gravels is aligned with the presently insignificant upper section of the southward flowing episodic Sand River, which joins the Orange River south of Griquatown. du Toit (1910: 354) concluded that ".....they are so numerous that there can hardly be any doubt that they are

the relics of gravels deposited by streams arising in the north and north-west, for it is noteworthy that the pebbles become gradually smaller in size to the south-east.....".

An analysis of a randomly-picked sample of about 60 pebbles and cobbles of surficial gravel from the farm Rooipan (Fig. 6.5), about 15 km north of Griquatown, contained predominantly rounded and sub-rounded components of both the Kuruman and Danielskuil members of the Asbestos Hills Formation, subrounded red jasper from the Ongeluk Formation, and angular black chert from the Ghaap Plateau Formation. All these components have their provenance in the north and north-west and indicate deposition by a southward flowing drainage. The gravel is deeply red stained and contains some heavily weathered components that were extremely difficult to analyse.

Similarly situated gravels are also present north of the wind gap in a stretch between the Griqua Town Hills and Sishen, along the present course of the ephemeral GaMogara River (Fig. 6.6). The northward-flowing GaMogara is a major tributary of the Kuruman River. Access to these gravels is very much restricted because of the presence of a military area which is a shooting range for heavy artillery. A surficial sample of these gravels on the Demaneng farm, about 15 km south of Kathu, was analysed and contained predominantly rounded to subrounded components of the Kuruman and Danielskuil members of the Asbestos Hills Formation, as well as some heavily weathered cobbles of the Ongeluk Formation, and some very angular chert breccia pebbles of the top manganese marker of the Ghaap Plateau Formation which crop out in the adjacent hills. The gravels are similarly red-stained and also contained similarly heavily weathered components as the gravels from the farm Rooipan. A general drainage direction from the south-east was inferred (Beukes, N. J., pers. comm., 1998) from the surficial gravel, which is indicative of a reverse flow after the initiation of the back-tilting of the proto-Molopo. This confirms the findings in respect of the gravels of the Mahura Muthla Formation.

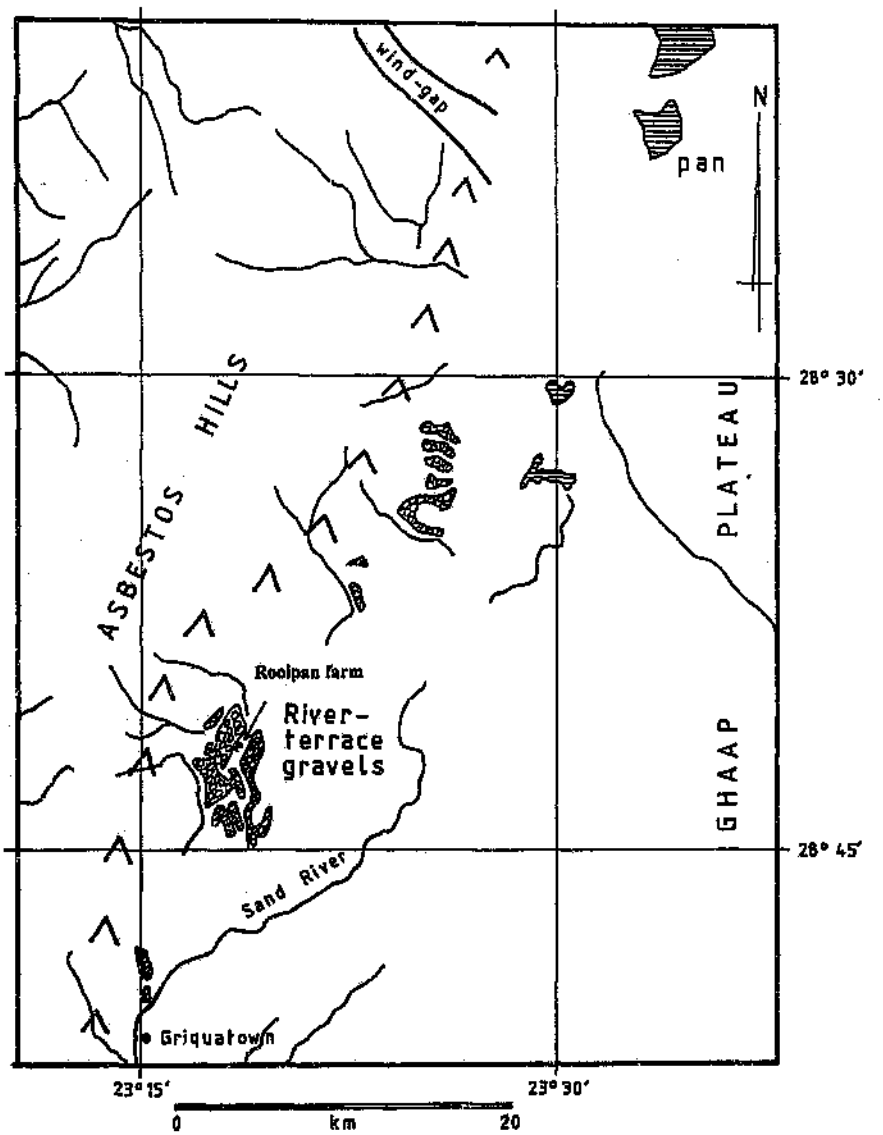


Figure 6.5 River-terrace gravels south-east of the Asbestos Hills.

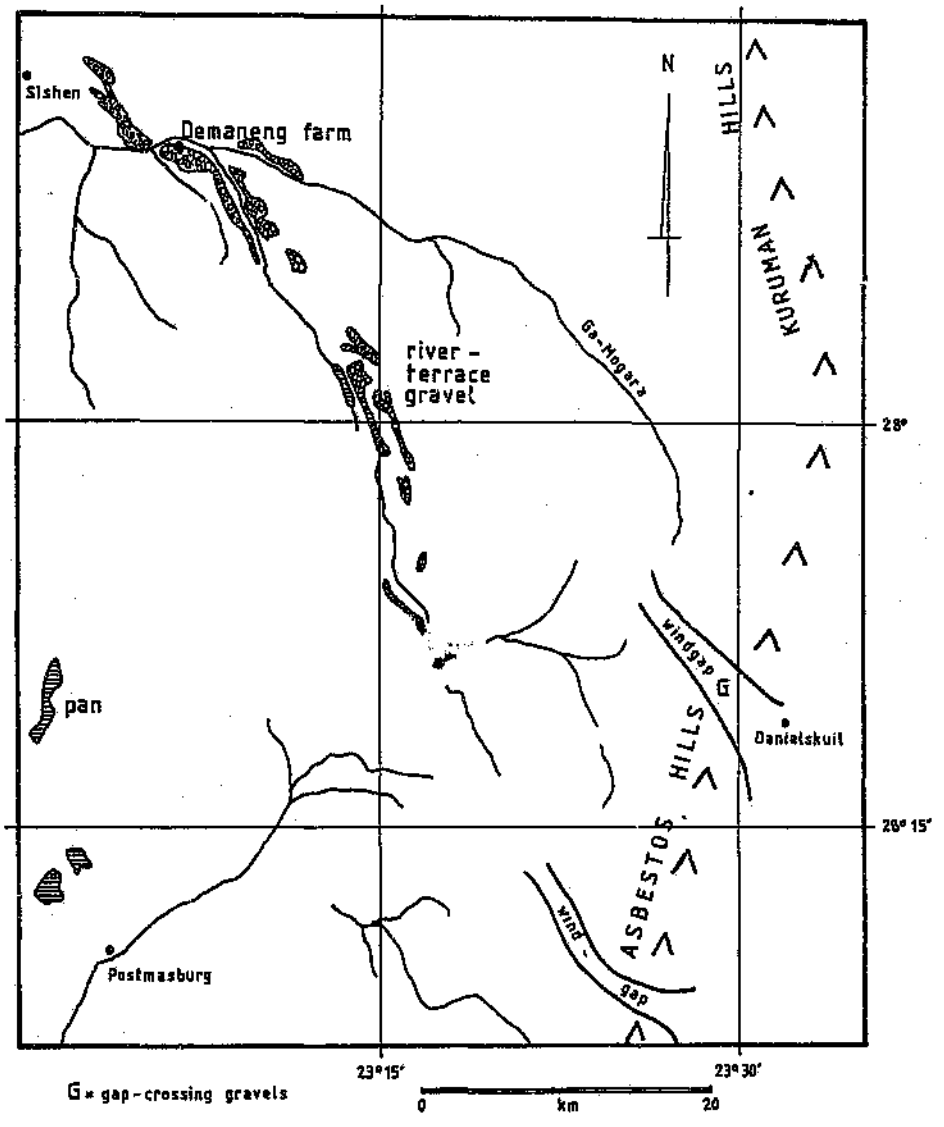


Figure 6.6 River-terrace gravels north-west of the Asbestos Hills.

The spatial pattern of the stretches of terrace gravels, both north and south of the Griqua Town Hills, provides further evidence of the previous existence of a drainage line across the Griqualand-Transvaal axis of uplift.

6.4.3 Wind gaps and gap-crossing gravels in the Asbestos Hills near Danielskuil

Two sizeable wind gaps are present in the Asbestos Hills west of Danielskuil. They are positioned between the gravel deposits on either side of the Asbestos Hills. The northerly gap is referred to as the 'Danielskuil' wind gap and the southerly one the 'Lime Acres' wind gap. The Danielskuil wind gap (Plate 6.7) has a NW-SE trend and a length of about 18 km with a width of about 1 km. The floor of the U-shaped valley is fairly flat and slopes away in both directions from the highest point in the centre of the gap at about 1550 m AMSL. Some borrow pits about 5 km upstream from the exit near Danielskuil, next to the gravel road through the gap, reveal rounded platy gravel, as well as larger size well-rounded cobbles, underneath a thin (10-20 cm) cover of red Kalahari Sand (Plate 6.8). Analysis of a sample of the gravel near the farm Beadles Post revealed mainly gravel from the Danielskuil member of The Asbestos Hill Formation, and a single red jasper cobble from the Ongeluk Formation. No pebbles of the Kuruman member were present, although the gravel bar is situated on a major outcrop of that rock type. The provenance of the gravel clearly points to a former flow direction from the north-west, from the area to the west of the hills.

The Lime Acres wind gap has a NNW-SSE trend and is situated immediately between the alluvial gravel deposits described above. This gap has a length of about 23 km and a width of about 1 km. Its U-shaped valley floor is extremely flat, which explains why the gap is traversed by the railway line from Postmasburg to Kimberley through the Asbestos Hills. This wind gap did not have any borrow pits in which gravel could be examined, and thus did not provide direct evidence of river flow through the gap. This wind gap is however situated directly between the gravel deposits to the north and south of it, and it is anticipated that alluvial gravel can be found under a Kalahari Sand cover. Both wind gaps may at some time have been used by the proto-Molopo drainage.

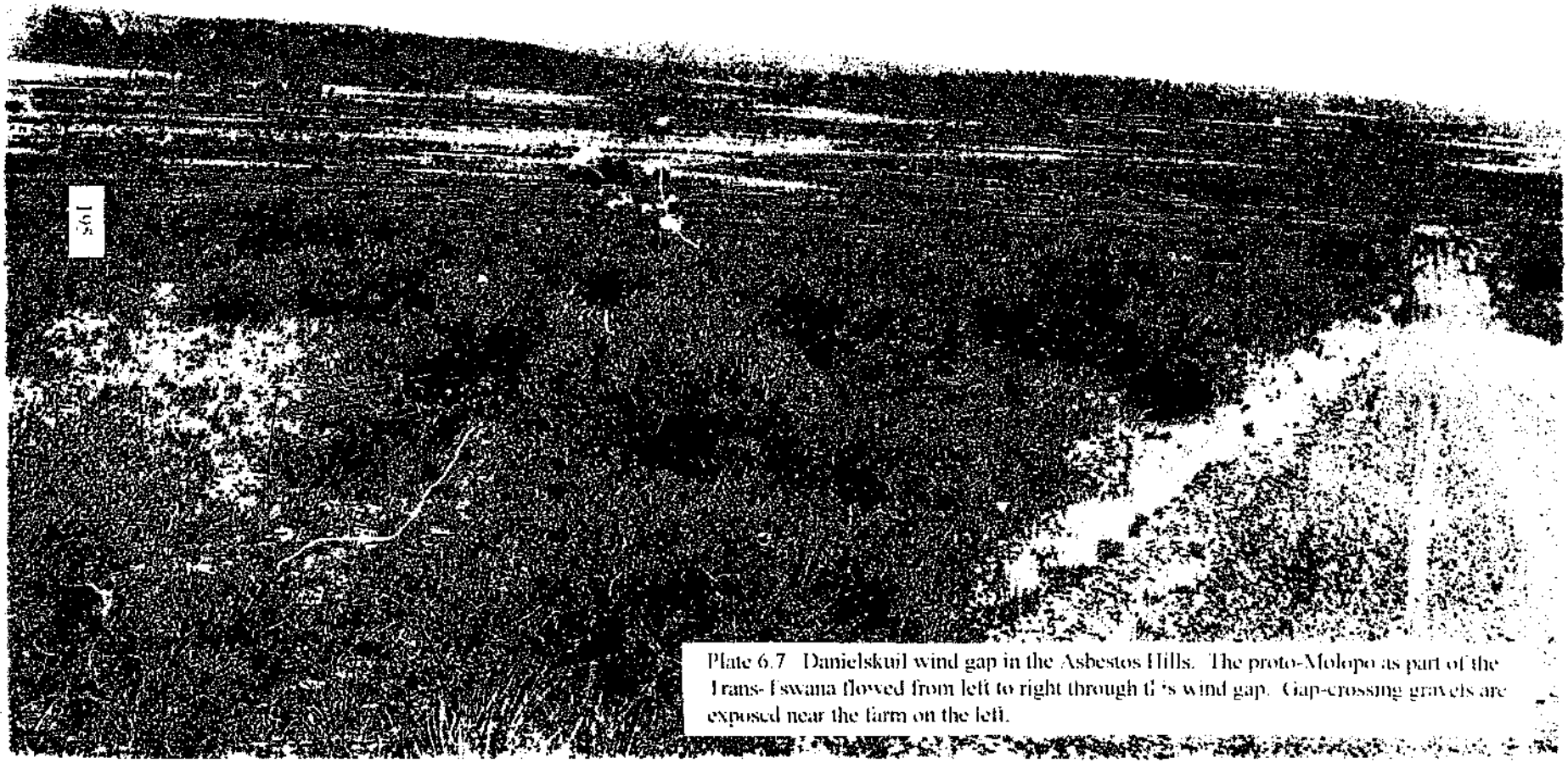


Plate 6.7 Danielskuil wind gap in the Asbestos Hills. The proto-Molopo as part of the Trans-Eswana flowed from left to right through it's wind gap. Gap-crossing gravels are exposed near the farm on the left.

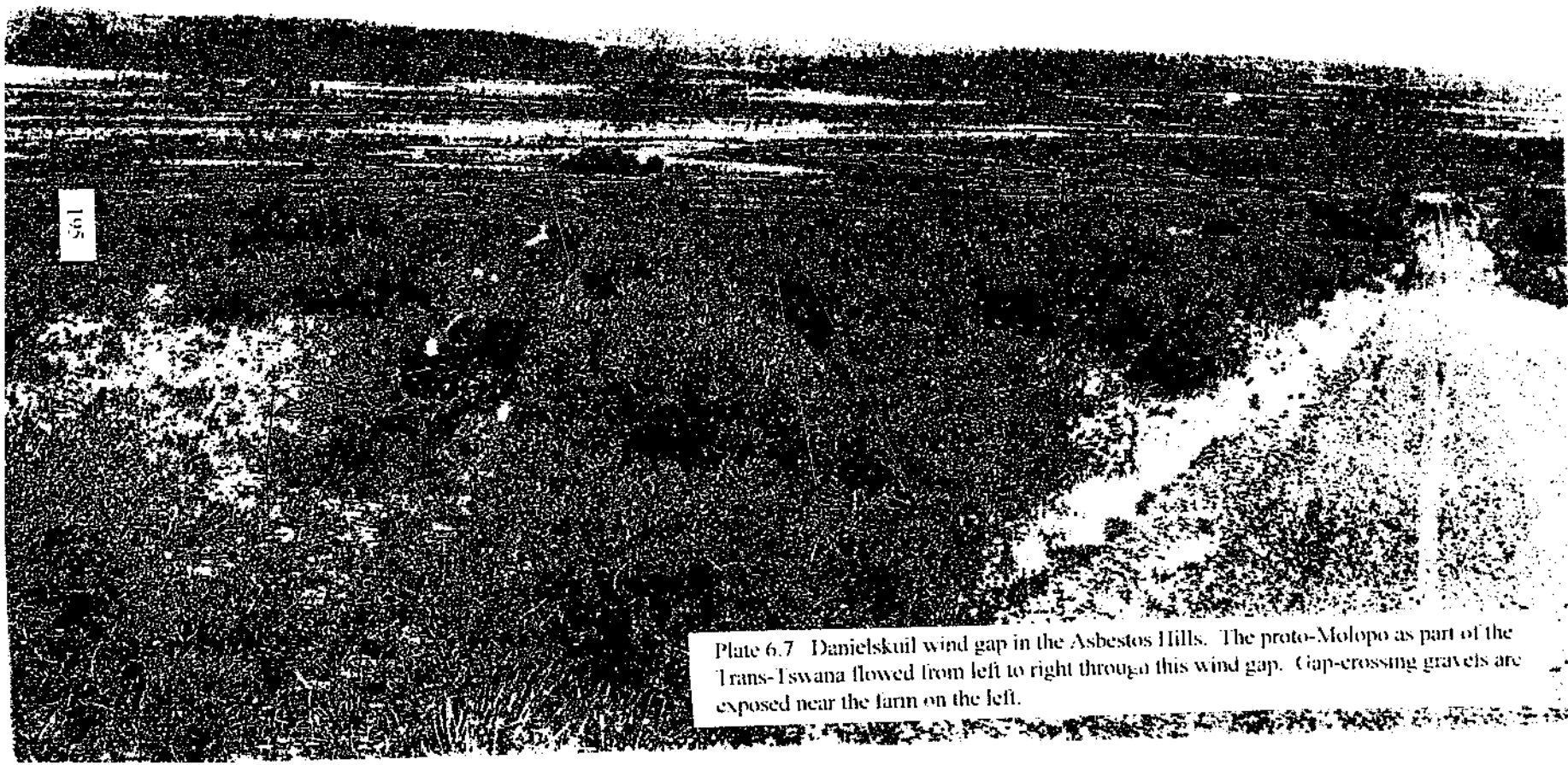


Plate 6.7 Danielskuil wind gap in the Asbestos Hills. The proto-Molopo as part of the Trans-Tswana flowed from left to right through this wind gap. Gap-crossing gravels are exposed near the farm on the left.



Plate 6.8 Gap-crossing gravels in the Danielskuil wind gap. The gravels are covered with a veneer of red Kalahari Sands. Palaeoflow was from right to left.



Wind gaps are, according to Bishop (1995), one of the most common types of morphological evidence for drainage realignment. Wind gaps alone are not sufficient evidence for drainage diversions, but need to be supported by additional geomorphological or geological evidence. As many types of evidence as possible should be used in reconstructing drainage rearrangements. Bishop (1995) cautioned the use of gap-crossing gravels as evidence for drainage realignments, as several Australian examples of gap-crossing gravels can be accounted for satisfactorily by weathering of local bedrock. The combination of the gravel deposits on either side of the Asbestos Hills in combination with the windgaps and the gap-crossing gravels in the Danielskuil gap, provide additional evidence for a previous course of the proto-Molopo across the present hills.

6.5 Sand River/Orange River Confluence Terrace Gravels

Evidence for the former existence of a major southerly flowing river in this area has been presented by McCarthy (1983). An extended review and interpretation of his findings follows because of the importance of the work to the drainage evolution of the Molopo. McCarthy's findings are based on clast assemblage analysis of alluvial gravels on high level Orange River terraces along the Sand River tributary (Plate 6.9), some 30 km downstream from the Vaal/Orange River confluence (Fig. 6.7).

6. 5. 1 Nature of the terrace gravels

The gravel on the Sand River terraces is generally poorly sorted and clasts typically measure about 25 cm along the major axis. Boulders up to 50 cm in diameter are not uncommon. The gravels are most strongly calcretised nearest the dolomitic rocks of the Ghaap Plateau. Above the calcretised gravels is often a zone of rubefied or "derived" gravel, which almost entirely consists of siliceous clasts. Some of the rubefied gravel has been recemented. Clast assemblages were determined through a sampling method. The most common clasts were green Ventersdorp lavas, quartzite, banded iron-formation (including riebeckite and jaspilite, and chert. McCarthy considered the banded iron-formation to be of paramount importance to his study, as these were derived from outside

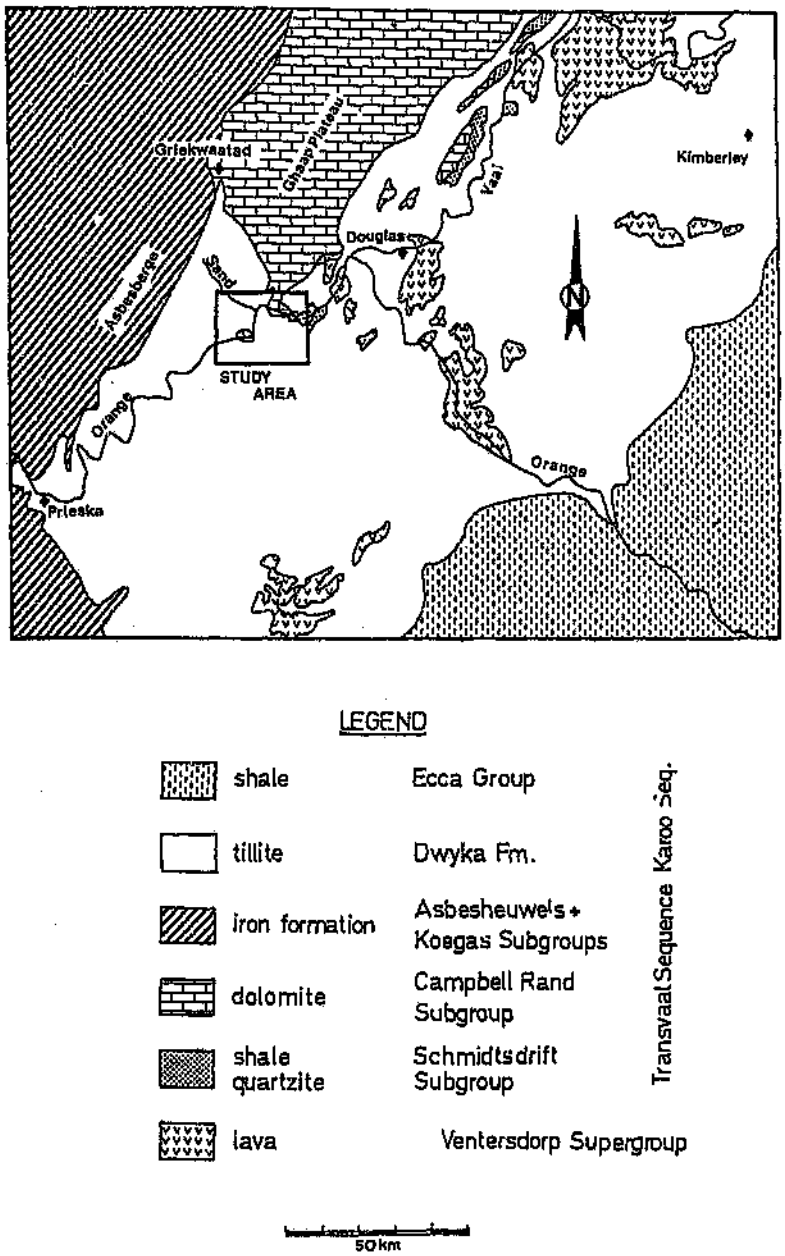


Figure 6.7 Regional geology and location of the Sand River/Orange River confluence (after McCarthy, 1983).



Plate 6.9 Caleretised gravels of the Sand River terraces. Cobble and boulder-size sediment is visible on top and inside the caleretised terrace.

the study area (Fig. 6.8). The results of the gravel clast analysis show, that to the east of the Sand River confluence the banded iron-formation is an insignificant constituent of the assemblages. The negligible presence of the banded iron-formation clasts in the terraces to the east of the Sand River, has according to Cooke (1946, as cited by McCarthy, 1983) been transported via old drainage systems through the western Transvaal. To the west of the Sand River, however, clasts of the banded iron-formation become a major component in all terrace levels, and boulders up to 30 cm diameter along the major axis are not uncommon. Furthermore, Lebombo Group lava, which is an important marker of an Orange River source, is entirely absent from the terraces of the Sand River, which supports a different provenance for the gravels of the Sand River terraces. McCarthy argued that the banded iron formation does not occur within the Transvaal Sequence rocks, neither in the Dwyka Tillite Formation in the study area, and thus, that the banded iron-formation clasts must have been fluvially introduced from outside of the area investigated. The ultimate source within the catchment of the Sand River terraces is most probably the iron-formation which forms the Asbestos Hills to the north west. With reference to the depositional environment of the gravels McCarthy (1983) inferred, that the bulk of the coarse gravels represented gravel bar accumulations formed during periodic flood events. He suggested that because of the extreme coarseness of the bulk of the gravels formation would have occurred in abnormally rapid flowing water. It was further suggested that the terraces recorded catastrophic events in the history of the river, with bed loads deposited at localities remote from the normal thalweg. McCarthy also noted the destructive effects of calcretisation on the clasts assemblages. Non-siliceous clast are increasingly eliminated through the process of calcretisation.

In an attempt to trace the input of the banded iron-formation clasts back in time McCarthy (1983) indicated, that there has clearly been a consistent input of large amounts of banded iron-formation from the earliest times, as they appear in the highest deposits at 1006 m and 1052 m, up to and including the time of formation of the 975 m deposits. Downstream of the Sand River all Orange River-associated terraces are dominated by

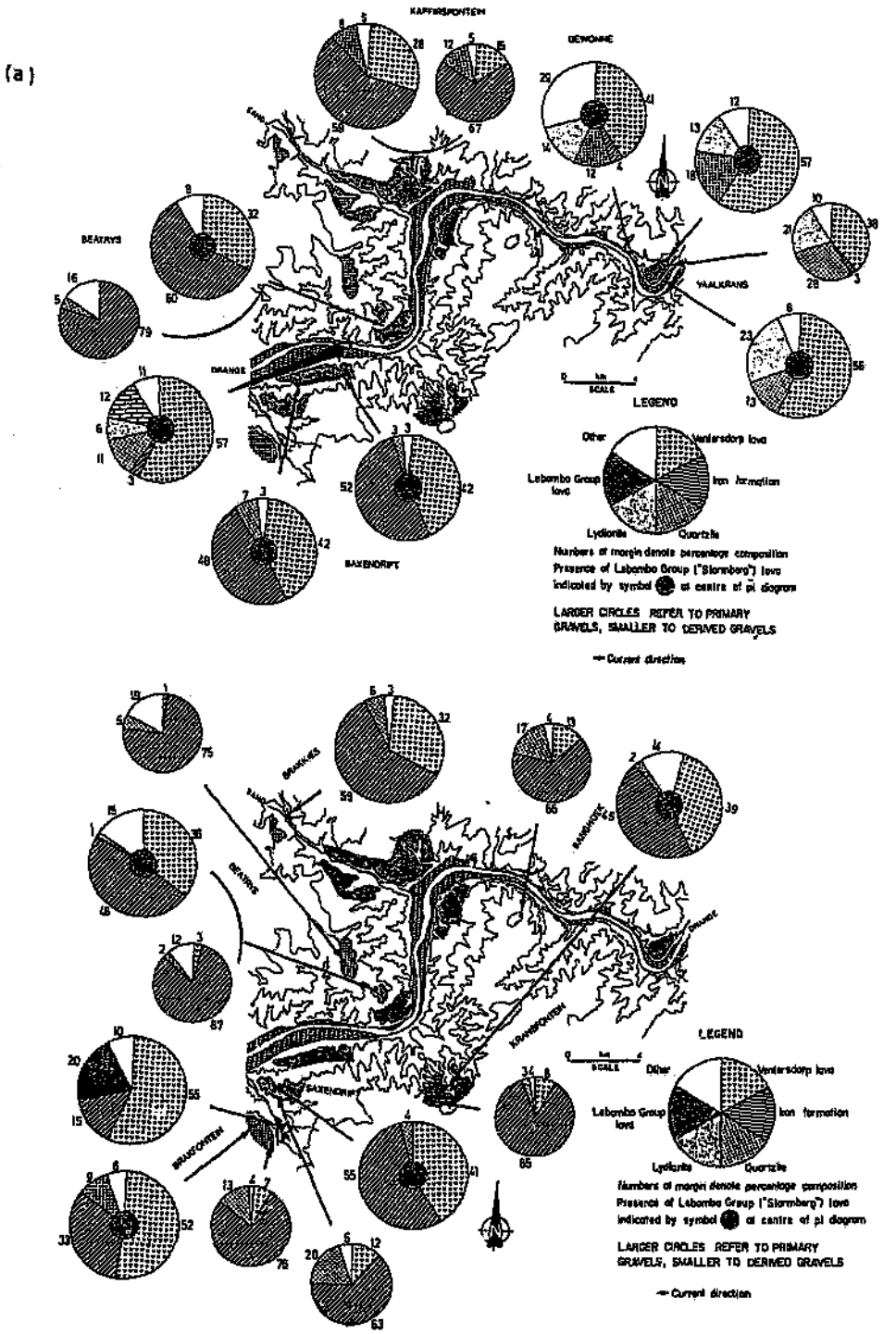


Figure 6.8 Composition of gravels on (a) the younger and (b) the older terraces in the Sand River/Orange River confluence area (after McCarthy, 1983).

banded iron-formation clasts. Upstream of the Sand River banded iron-formation clasts were at no stage a major component of the either the Vaal or the Orange River sediment input.

6.5.2 The Trans-Tswana River

On the basis of the contribution of the banded iron-formation to the mixed gravels McCarthy calculated, that the palaeo-river entering from the north consistently carried about four times as much sediment as the Orange River in an equivalent size range, and that this must relate to the relative size of the rivers. Because of the uncertainties included in these calculations McCarthy concluded cautiously, that the northerly river was at least of a substantial size, certainly equivalent of the Orange River, and probably much larger. It existed for a considerable period of time as reflected in the terrace gravels from 1067 m down to 975 m. The limited erosional incision of the 975 m terrace suggests that the demise of the northerly river was extremely rapid after the deposition of the 975 m terrace gravels. The distribution of the Riverton Formation suggests that the northerly river had definitely ceased by that time (~ later middle Pleistocene: Helgren, 1979). The catchment of the northerly river must have been very large, and according to McCarthy, assuming that semi-arid conditions had already commenced in the Kalahari region, could have encompassed much of south-central Africa. He proposed the name Trans-Tswana for this river. It appears that because of the geographical position of the proto-Molopo River it would have been a major part of this Trans-Tswana River.

The idea of a central southern African river had been discussed much earlier by Schwarz (1919). He suggested the existence of a 1800 km long proto-Orange (Fig. 6.9) which had the Loangwe (now Luangwa) River as its origin in the highlands of Tanzania. The Loangwe River flows south-west to join the Zambezi near Zomba (now Zumbo). From there the proto-Orange flowed upstream the Zambezi channel to Wankie (now Hwange) and then through the Gwai River and Gwai Gap to the Sua Pan. From the Sua Pan it then flowed to Andersson's Vlei and the Molopo River near Makopong, or possibly to the Nossob River. It finally reached the sea via the lower Molopo and the Orange River.

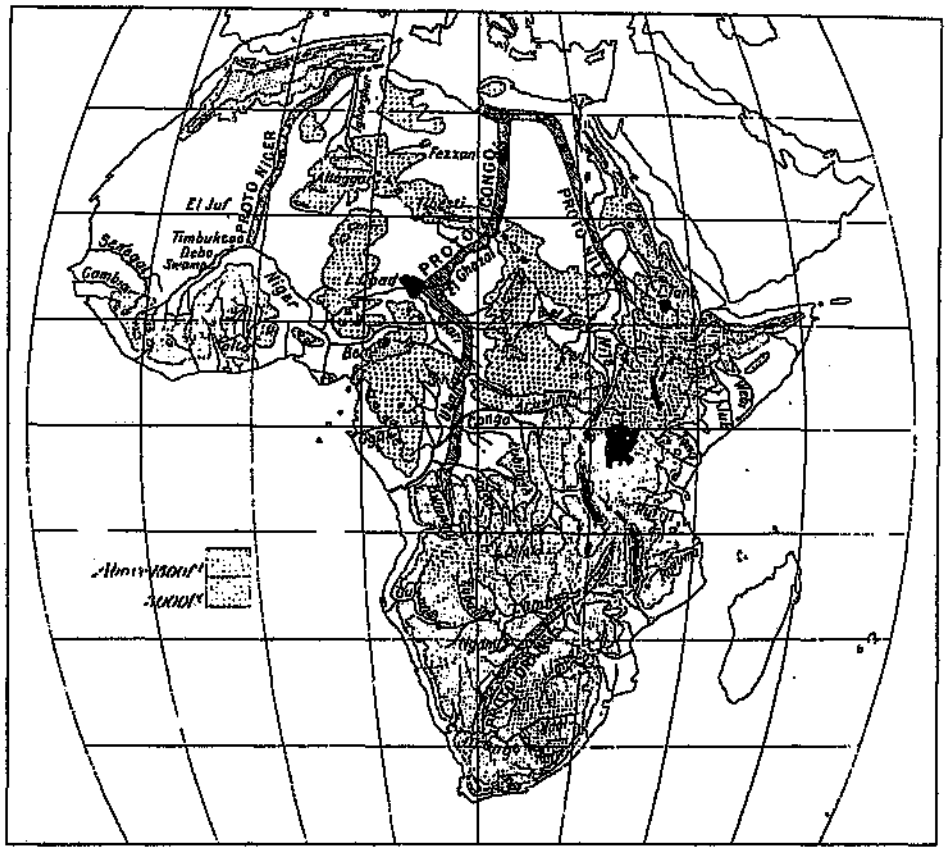


Figure 6.9 The proto-Orange as suggested by Schwarz (1919).

Schwarz (1919) further supported his ideas by citing Livingstone (1858), who had made an indirect reference to an ancient river, which once flowed from the Linyanti (Chobe) Basin to the Orange. The southern sector of the proto-Orange formed the basis for Schwarz's Kalahari irrigation scheme (Schwarz, 1920, 1923) through reviving ancient Kalahari lakes (Barnard, 1992). Although Schwarz's work has not always been taken very seriously (Barnard, 1992), it is interesting to see that the Trans-Tswana River (Fig. 1.7) as suggested by McCarthy (1983) resembles to some extent the course of Schwarz's (1919) proto-Orange as well as its suggested relative size. McCarthy suggested that the volume of the Trans-Tswana could have been four times that of the upper Orange, making it the main stream. McCarthy's (1983) anticipated course of the Trans-Tswana differed from that of Schwarz's (1919) proto-Orange upstream of the Sua Pan. McCarthy suggested the Trans-Tswana to have flowed from the north and to have drained via the present Okavango Delta. Schwarz suggested the proto-Orange to have come from the north-east via the Gwai River to the Sua Pan. Both suggested courses are highly hypothetical. The sediment-buried valley of the proto-Molopo is, however, inclined towards the Limpopo, and an informed guess about the previous components of the Trans-Tswana would be, that it is more likely that it used the headwaters of the proto-Limpopo, which followed the present western border of Botswana. A southward flow and drainage reversal in the upper Limpopo has been suggested earlier (Wellington, 1955: 361).

6.5.3 *Uplift along the Griqualand-Transvaal axis and drainage diversion*

Since the Griqualand-Transvaal axis of uplift lies across the inferred course of the ancient Trans-Tswana River, McCarthy (1983) ascribed the demise of the Trans-Tswana to the effect of warping of the axis. du Toit (1933) indicated that the position of the axis of uplift constitutes the divide from which the 'primitive' drainages were 'thrown off' in opposite directions and that they are generally not far off from the present watersheds. This seems applicable to the proto-Molopo/Trans-Tswana. Since the introduction of the axis by du Toit (1933) there has been some debate about the nature and timing of the activity of the axis. du Toit (1933) had initially suggested that northerly sinking was the

main process along the axis, but Mayer (1973), Partridge and Maud (1987) and Marshall (1990) considered this rather to be an axis of uparching at various times during the Cainozoic. du Toit (1933) suggested that the diastrophism associated with the Griqualand-Transvaal axis began in the middle of the Tertiary, possibly extending into the Pleistocene.

Mayer (1973) argued, that if the axis was an isostatic adjustment of the basement in the form of a comparatively narrow ridge-like uplift, it would have been inevitable that there would also have been tilting of the landsurface on the southern side of the axis comparable with the northern tilt suggested by du Toit (1933). Mayer (1973) suggested a more southerly position for the Griqualand-Transvaal axis, roughly parallel to the Vaal River between Klerksdorp and Barkly West, and a late Tertiary to early Pleistocene age for the main phase of diastrophism. Partridge and Maud (1987) indicated that uplift along the Griqualand-Transvaal axis in Miocene times (~ 18 Ma), was confined chiefly to the western section of the axis in the northern and north-western Cape only (outside the Molopo drainage area). They further indicated that the eastern part of the axis came into being for the first time only during the late Pliocene (~2.5 Ma), but was located a considerable distance to the south of its Miocene counterpart.

Evidence of an incipient structure of the Griqualand-Transvaal axis, in the form of a minor watershed between the Vaal and the lower Harts River, sometime before the end of the African landscape cycle was indicated by Marshall (1990). Contrary to Partridge and Maud (1987), she believed that active uplift along the eastern portion of the Griqualand-Transvaal axis did occur in the Miocene, and that an incipient form of the axis existed even prior to that. She further indicated that the uplift in the Pliocene was concentrated along virtually the same axis but appears to have been of greater amplitude. A slight alteration of the position of the uplift axis, however, resulted locally in the inversion of topography (Marshall, 1990).

McCarthy (1983) speculated that in case of rapid warping of the axis under semi-arid climatic conditions the mature, low gradient Trans-Tswana drainage system would dam up, and form an internal drainage basin. The damming of the Trans-Tswana could have occurred as early as the beginning of the basining of the Kalahari, when the proto-Molopo became back-tilted in relation to the incipient form of the Griqualand-Transvaal axis. With permanent closure a gradual infilling of the drainage basin would have occurred. McCarthy noted in this respect that the nature of some of the lower formations of the Kalahari Group are compatible with deposition in a lacustrine environment. This is supported by the findings of Smit (1977), Behr (1989), and the description of the Mamatwan clay body and overlying Kalahari Group sediments in this study.

It is suggested that the influence of the incipient form of the Griqualand-Transvaal axis and/or subsequent warping of the Griqualand-Transvaal axis of uplift caused a diversion of the proto-Molopo, which probably formed the lower part of the Trans-Tswana as described by McCarthy (1983). The drainage diversion resulted in the development of a lacustrine basin as the headward portion of the proto-Molopo was back-tilted with the lowering of the Kalahari Basin. It seems that the initiation of the lowering of the Kalahari Basin started already at the end of the Cretaceous. This would indicate the start of gentle back-tilting of the proto-Molopo. Additional warping of the axis of uplift would have exacerbated the back-tilting. Warping of the axis would have happened during the Miocene (Marshall, 1990) and Pliocene (Partridge and Maud, 1987). The combined lowering of the Kalahari Basin together with the warping of the axis, would have resulted in a very sizeable total amount of back-tilting. Simultaneous warping of the Kalahari-Zimbabwe axis would have further affected the Trans-Tswana drainage which supposedly extended much further north.

6.6 Kathu Pan Doline and Wonderwerk Cave

Two significant palaeoenvironmental sites are situated within the Griqualand West area (Fig. 6.10). Kathu Pan was described by Thomas and Shaw (1991) as an ephemeral

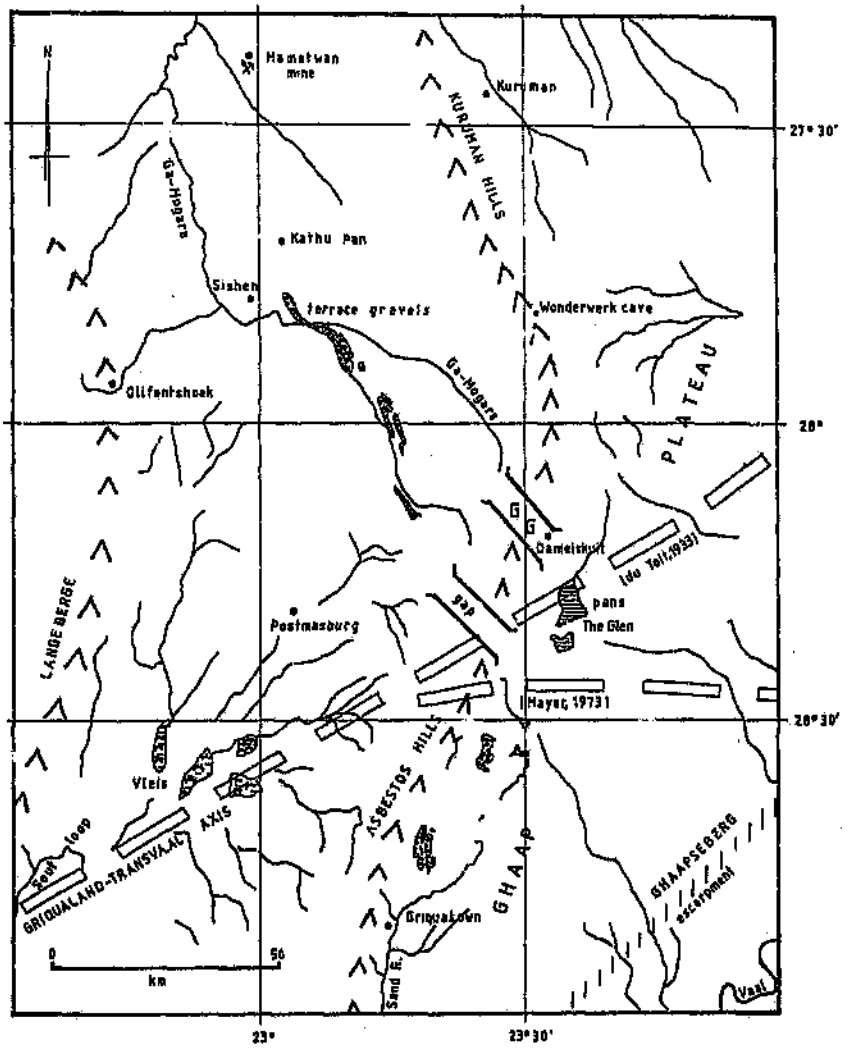


Figure 6.10 Palaeodrainage indicators in the Griqualand West study area.

wetland area underlain by basal Kalahari Group sediments, that are capped by 40-50 m of karstic calcrete. The Kathu Pan doline developed during the Tertiary in calcrete, which itself post-dated the deposition of basal Kalahari beds. The doline must have developed under a very low water table and thus dry climatic conditions. A sediment sequence subsequently developed within the doline under varying climatic conditions. Although Butzer (1984) described the wetland sediment sequence as 'possibly the best palaeoenvironmental sequence from the Kalahari Basin', Thomas and Shaw (1991) cautioned that the data considerably lack resolution and that the broad time periods for the individual sediment units make correlation events from this area with other areas of the southern Kalahari uncertain. They considered the sequence to be 'some supporting evidence for fluctuating climates from within the southern Kalahari'.

The Wonderwerk cave is situated about halfway between Kuruman and Danielskuil and has developed as a solution cavity into the stratified dolomitic limestone base of a low conical foothill on the eastern flank of the Kuruman hills. The horizontal cave is 139 m deep and was exposed by hillside erosion. The entrance of the cave has a sequence of four metres of wind-blown dust with a variable admixture of roof slabs. The sequence is supposed to cover a period of 800 000 yrs (KDC-MM, undated). Beaumont *et al.* (1984) and Butzer (1984) presented a palaeoenvironmental framework for the late Pleistocene and the Holocene section of the sequence. They envisaged several fluctuations between more moist and cold climatic conditions with frost-spalling in the cave and conditions similar to today.

Both sites indicate fluctuating climatic conditions during the latter part of the Pleistocene and the Holocene. Beaumont *et al.* (1984) in a summary indicated that a shift occurred from a cool dry Pleistocene to a more moist and warm Holocene optimum which lasted from 8 000 to 5 000 BP. Cooler, wetter and more variable conditions were indicated from 4500 BP onwards. Such climatic fluctuations would have resulted in variations in the flow in the upper sections and possibly further down the network of the Molopo drainage, especially from the rivers which originate outside the sand covered area of the Kalahari

Basin. The only indication of perennial flow in the lower Molopo River during the late Quaternary, on the basis of sediments and carbon-dated molluscs, is suggested to have occurred during 17 000 - 15 000 BP, and possibly around 12 000 (Heine, 1978a; 1979b; 1982). Heine indicated no further perennial discharge during the Holocene, and described the nature of the drainage as episodic.

6.7 Griqualand West Palaeodrainage Indicators

The Griqualand West area holds many clues to the reconstruction of the palaeodrainage of the region. Especially the gravels close to or on the Griqualand-Transvaal axis of uplift point to previously more extensive drainage systems. The alignment of the terrace gravels between Sishen and Griquatown together with the wind gaps near Danielskuil and the drainage evidence at the Sand-Orange River confluence provide substantial indicators for a proto-Molopo that flowed through the Hotazel valley and connected with the proto-Orange south of Griquatown. The back-tilted upper drainage of the Soutloop south-east of Postmasburg (Fig. 6.10), the large pans (e.g. The Glen) near Danielskuil, the presence of the Mahura Muthla diamondiferous gravels, and the interrupted proto-Molopo drainage line, all provide evidence for the position and warping of the Griqualand-Transvaal axis.

* * * * *

This chapter has presented indicators to the proto-Molopo palaeo-drainage in the Griqualand West area. All points of evidence presented are related to the Griqualand Transvaal axis of uplift. The sedimentary sequence exposed in the Mamatwan mine presents the infilling of a structural basin developed north of the axis of uplift. Gravels near and on the axis (Mahura Muthla) indicate river flow across the axis prior to its existence. The former course of the proto-Molopo (and possibly the Trans-Tswana) can be traced from the basin sediments to the north of the axis via remnant patches of terrace gravels and wind gaps with gap-crossing gravels on the axis to the palaeo-drainage evidence at the Sand River/Orange River confluence to the south of the axis. There is

thus ample evidence that the proto-Molopo, possibly as part of the Trans-Tswana, entered the Orange River south of the present Vaal/Orange confluence, but much further upstream from the present entrance point, just downstream of the Augrabies Falls. Chapter 7 will review the evidence of the drainage indicators presented in Part 2 as a time sequence from Permo-Carboniferous times to the present.

PART 3 DRAINAGE RECONSTRUCTION AND CONCLUSIONS

CHAPTER 7 MOLOPO DRAINAGE PHASES

7.1 Introduction

Based on the evidence presented in the preceding chapters a synthesis will be proposed in this chapter. The shape of the Kalahari Group sediment body within the Molopo sub-basin, in conjunction with evidence of remnant surficial alluvial sediments and additional morphological evidence at the north-easterly and south-easterly sections (upper Molopo and Griqualand West) of the drainage area, form the basis of an explanatory model which attempts a reconstruction of phases in the evolution of the Molopo drainage. These phases should be seen as time frames in a more continuous development model, and the indicated ages can be used only as an approximation of the actual age, since the ages are based on very limited, and often unreliable data, that so far have appeared in the literature on the geology and geomorphology of the Kalahari. The following drainage phases can be identified: a Permo-Carboniferous drainage phase, a mid to late Cretaceous phase, a mid-Tertiary drainage phase, a Plio-Pleistocene drainage phase and a Pleistocene-Holocene drainage phase. It appears that the evolution of the Molopo drainage can be traced back over a considerable period of time (~ 300 Ma) and that during that time the evolution of the drainage pattern has been influenced by tectonics, associated with large-scale intra-continental structures (e.g. the Kalahari Basin and Griqualand-Transvaal axis of uplift), as well as climatic changes. In addition an attempt is made to separate the effects of tectonics and climatic changes.

7.2 Permo-Carboniferous Drainage Indicators

A reasonably well-developed picture of the drainage lines before the Permo-Carboniferous glaciation, as well as from the various stages during the maximum glaciation and in the final stages of the deglaciation in this area has emerged, based largely on the work of Visser (e.g. 1981; 1983a, b; 1987; 1989). Drainage information emerges from knowledge about the outcrops and suboutcrops of the Permo-Carboniferous Dwyka Tillite within the drainage area as well as from the larger picture of the palaeogeography of southern Africa during that period.

7.2.1 Preglacial landscape

The Permo-Carboniferous Dwyka glaciation is estimated to have started around 300 Ma ago and lasted for about 20-30 Ma (section 3.1). Visser (1987) stated that the Kalahari (*Karoo*) Basin represented a glacially modified valley system with valley-extensions into the higl land areas. The suboutcrops of the Dwyka Formation in the Molopo drainage area are interpreted as a glacial sedimentary infilling of a pre-Dwyka, late Carboniferous valley system. This very ancient (more than 300 Ma) valley system incised rocks of the Cargonian Highlands. The largest valley is the NNE-SSW Terra Firma valley which forms a valley system together with the SSE-NNW Hotazel Valley and the N-S South Valley tributaries (Fig. 7.1). The Terra Firma and Hotazel Valleys were both cut mainly into the relatively soft lavas of the Ongeluk Formation, while the South Valley was cut into the less resistant rocks within the Matsap Formation. The three valleys shared a joint exit via the Khuis Outlet and drained into a north-westerly direction to join the central WSW-directed Kalahari (*Karoo*) Basin drainage towards an incipient South Atlantic sea arm. Additional drainage lines from the area of the present lower Molopo seem also connected with the valley system downstream of the Khuis Outlet. Since the fluvial valley system was modified by glacial action and more than 270 Ma of surface lowering, the dimensions of the initial, pre-glacial fluvial valley system can now only be estimated. Total surface lowering of the Cargonian Highlands could be as much as 4 km. The major valley, which is situated just east of the position of the present Molopo River, has a length of about 150 km, a width of about 30 km and a depth in places of more than 900

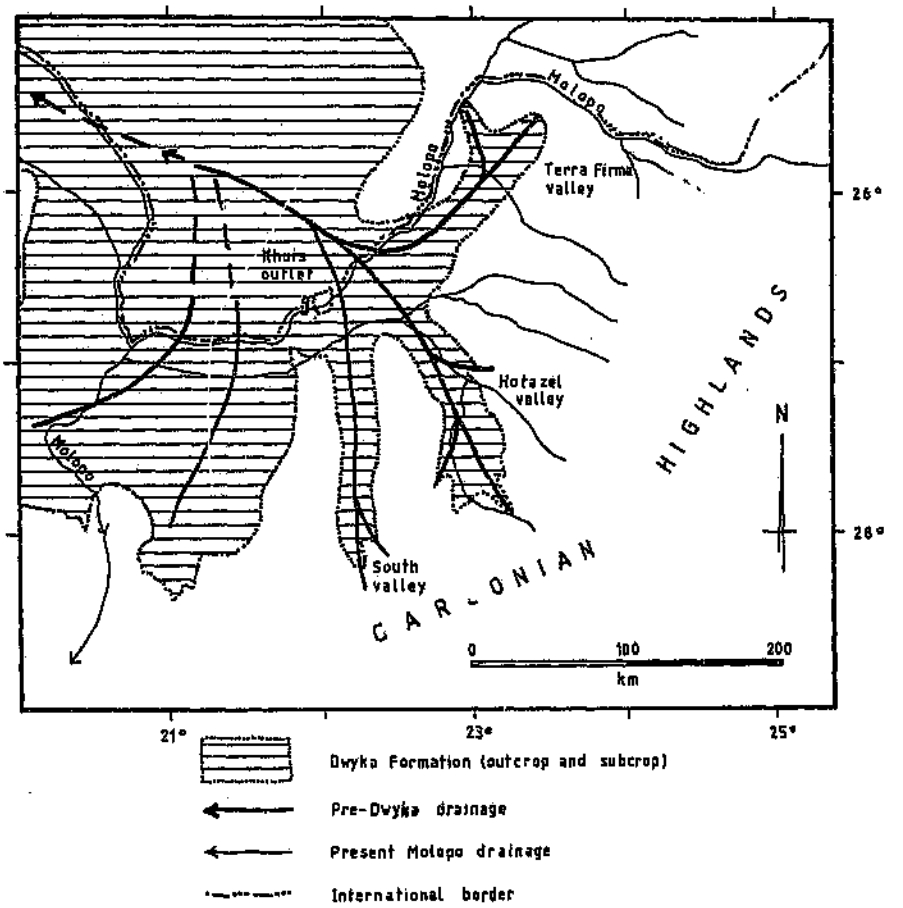


Figure 7.1 Pre-Dwyka palaeodrainage in the Molopo drainage area.

m. The great depth has been attributed to glacial overdeepening (Visser, 1987) and more recently Reimold (pers. comm., 1997) suggested that it might be related to the Morokweng impact structure. The dimensions of the slightly smaller, Hotazel tributary valley have already been described above (section 3.3.2), while little is known about the South Valley. Only Visser (1987) indicated a glacially modified valley in the location of the South Valley, and it is not shown on later geological maps. The pre-Dwyka Molopo drainage was most likely a rather deeply incised river system which originated near the axis of the Cargonian Highlands as an eastern tributary of a larger Kalahari (*Karoo*) drainage system. The valley system was subsequently remoulded by the glaciers and ice fields of the Permo-Carboniferous glaciation and a subsequent very long period of erosion.

7.2.2 Modifications during maximum glaciation and deglaciation

It has been noted that the thickness of the ice cap during the maximum glaciation was estimated at 3000 m. The effect of isostatic rebound after deglaciation on post-Dwyka drainage lines has been difficult to assess, but in certain places it seems to have been largely cancelled out by regional depression. The Permo-Carboniferous glaciation is suggested to have come to an end during the Sakmarian stage, and most of the ice must have disappeared around 270 Ma ago (Visser, 1987). Conditions during the Tses glaciation stage could have been very similar to the present conditions in parts of Alaska, where the glaciers are in direct contact with the ocean and have fjords where the sea made incursion into the land and ice field. Subsequent to the end of the Tses glaciation the Ecca Sea transgressed further inland (Fig. 3.12) and shallow marine conditions occurred over much of the Kalahari (*Karoo*) Basin (including the Molopo sub-basin) and the main Karoo Basin.

7.2.3 A geological hiatus

The end of the Permo-Carboniferous glaciation and the initiation of the sedimentation in the Cainozoic Kalahari Basin (70-80 Ma) are separated by a geochronologic hiatus of around 190-200 Ma. Africa became separated from the other parts of Gondwana between

142 and 127 Ma. With most of Africa standing at about 2000 - 2500 m AMSL (Partridge and Maud, 1987), this marked the start of a period of rapid erosion (the African erosion cycle), which was aided by the humid, tropical climates of the Cretaceous and resulted in deep weathering profiles (Partridge, 1997a).

About 75 Ma before the beginning of the formation of the Cainozoic Kalahari Basin the eastern section of the Molopo drainage was affected by a very large meteorite impact near Morokweng, which resulted in a impact structure with a diameter of about 70 km, and possibly as large as 300-340 km (Corner *et al.*, 1997). The Morokweng impact structure has been dated at 145±2 Ma and coincides with the lower boundary of the Cretaceous. The enormous impact strongly influenced the regional drainage conditions, as well as the distribution of the Karoo sediments, over a large area surrounding the impact centre (Reimold, *et al.*, in press). The general drainage direction for the region before the Morokweng meteorite impact was towards the north-west, while drainage evidence from after th ··teorite impact, points to a southerly drainage direction (Bootsman and Reimold, .. 8). In addition, the post-impact drainage pattern shows distinct annular components. Patches of Dwyka tillite of up to 300 m thick, about 35 km to the east of the impact centre, could form part of an inner ring structure (see Fig. 3.9, 7.2, and 7.3 in combination).

Several kimberlite intrusions were emplaced within and around the Molopo drainage during the Upper Cretaceous (e.g. Kimberley Pool 95.4 Ma, Finch 94.1 Ma, and Orapa 93.1 Ma: Cahen *et al.*, 1984). Except for evidence of some deposits of the Prince Albert Formation (Khuis and Middelputs Formations in Botswana), in the south-western region of the Molopo drainage, there is no further evidence of sedimentation over tha period within the drainage area, but certainly substantial erosion certainly must have occurred during that period. It is generally assumed that most of southern Africa, with the exception of a few higher points, was covered with the sediments of the Karoo Sequence (Partridge and Maud, 1987). Sediments of the Karoo Sequence in the Molopo drainage

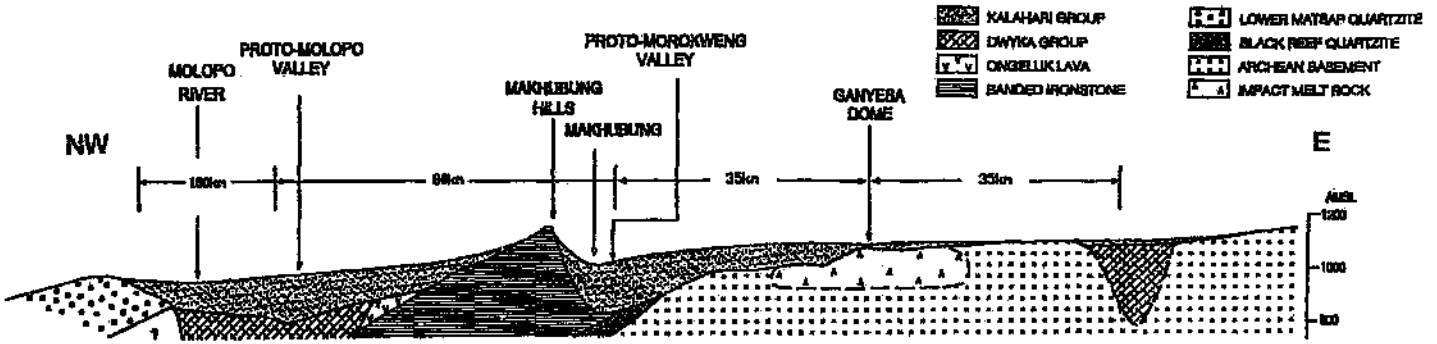


Figure 7.2 Schematic geological cross section through the Morokweng impact structure (after Reimold *et al.*, in press).

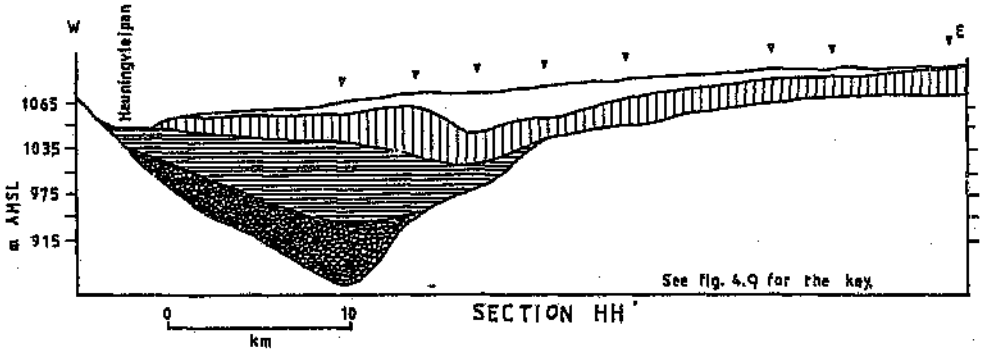


Figure 7.3 Alluvial cross section across a palaeovalley around the Morokweng impact structure near Heuningvlei. This represents the sediment-filled proto-Morokweng valley in Figure 7.2. Based on Smit (1977). See Figure 3.9 for the location of the cross section.

are now, however, only present in the lower parts of the Kalahari (*Karoo*) Basin and mostly covered with sediments of the Kalahari Group. Resistant hills in the area such as the Kuruman Hills, Langeberge, and Skurweberge were exhumed after a substantial period of erosion to become again a major influence on the development of the Mesozoic and Cainozoic drainage systems of the area as was also suggested by du Toit (1910).

7.3 Mid to Late Cretaceous Drainage Phase

The evidence for the reconstruction of the Cretaceous drainage phases is based upon the shape of the Kalahari Group sediment body within the Molopo sub-basin (Fig. 4.6) and the revealed pre-Kalahari topography (chapter 4), as well as from sediments and landforms in selected areas of the eastern rim of the Kalahari Basin. Two selected areas were discussed in chapter 5 (upper Molopo drainage area) and chapter 6 (Griqualand West).

7.3.1 Implications of the Kalahari Group isopach pattern

The isopach pattern of the Kalahari Group sediments revealed a pre-Kalahari drainage system which is interpreted as having been back-tilted with respect to the Griqualand-Transvaal axis and filled in with Cainozoic sediments. The present surface of the Kalahari Group sediment body is tilted to the north-west, and the sediment has partially submerged three sub-parallel resistant ridges. It is suggested that these three ridges together with the regional tilting can explain a step-like shift of the previous courses of the Molopo through time. The drainage lines have shifted in a step-like fashion from east to west while making use of gaps in the ridges, in response to both the lowering of the Kalahari Basin and the warping of the Griqualand-Transvaal axis. It is thus logical to look for the oldest drainage evidence of the Molopo in the most easterly section of the drainage area.

The pre-Kalahari drainage lines show a strong relationship with the structure of the basement rock, but it is clear that with the infilling of the Molopo sub-basin that affinity with the basement structure has been greatly reduced. The most recent phase of the

Molopo drainage has developed in response to the tilted, duricrusted surface of the sediments within the Molopo sub-basin and bears little relation with the pre-Kalahari geology.

7.3.2 The proto-Morokweng River

Prior to the basining of the Kalahari the area of the present Molopo drainage was dominated by a southward flowing drainage system, that was part of de Wit's (1993) upper proto-Orange or Karoo River (Fig. 7.4). The Karoo River was made up of the proto-Vaal, the proto-Harts, the proto-Morokweng, and the proto-Molopo River. The previous existence of the proto-Morokweng River is based on the evidence of two remnants of its fluvial activity: a semi-circular sediment body with a radial distance of 30-35 km from the centre of impact of the Morokweng impact structure, as well as a preserved channel section of a southerly flowing river near Mahura Muthla on the Ghaap Plateau The proto-Morokweng River would have joined the Karoo River just north of the present Vaal/Harts confluence.

The semi-circular sediment body is interpreted as a sediment-filled section of a previous river channel, which was affected by the ring structure of the Morokweng impact structure. The proto-Morokweng River flowed partially around the impact structure in a very similar version as the present Vaal River through the Vredefort Dome (Reimold and Gibson, 1996). Tangential faults associated with the Morokweng ring structure would have further controlled the proto-Morokweng River. The more than 180 m thick Kalahari Group sediment body around the circular Morokweng impact structure, substantially thicker than the sediment-filled tributary channel of the proto-Molopo within a gap of the Kuruman-Makhubung hills range, can only be the westward-tilted remnant of a N-S trending ancient river system.

The meandering channel remnant on the Mahura Muthla farm forms the second piece of the reconstruction of the proto-Morokweng River. Partridge *et al.* (in prep.) indicated the presence of a late Cretaceous drainage system on the Griqualand-Transvaal axis. The

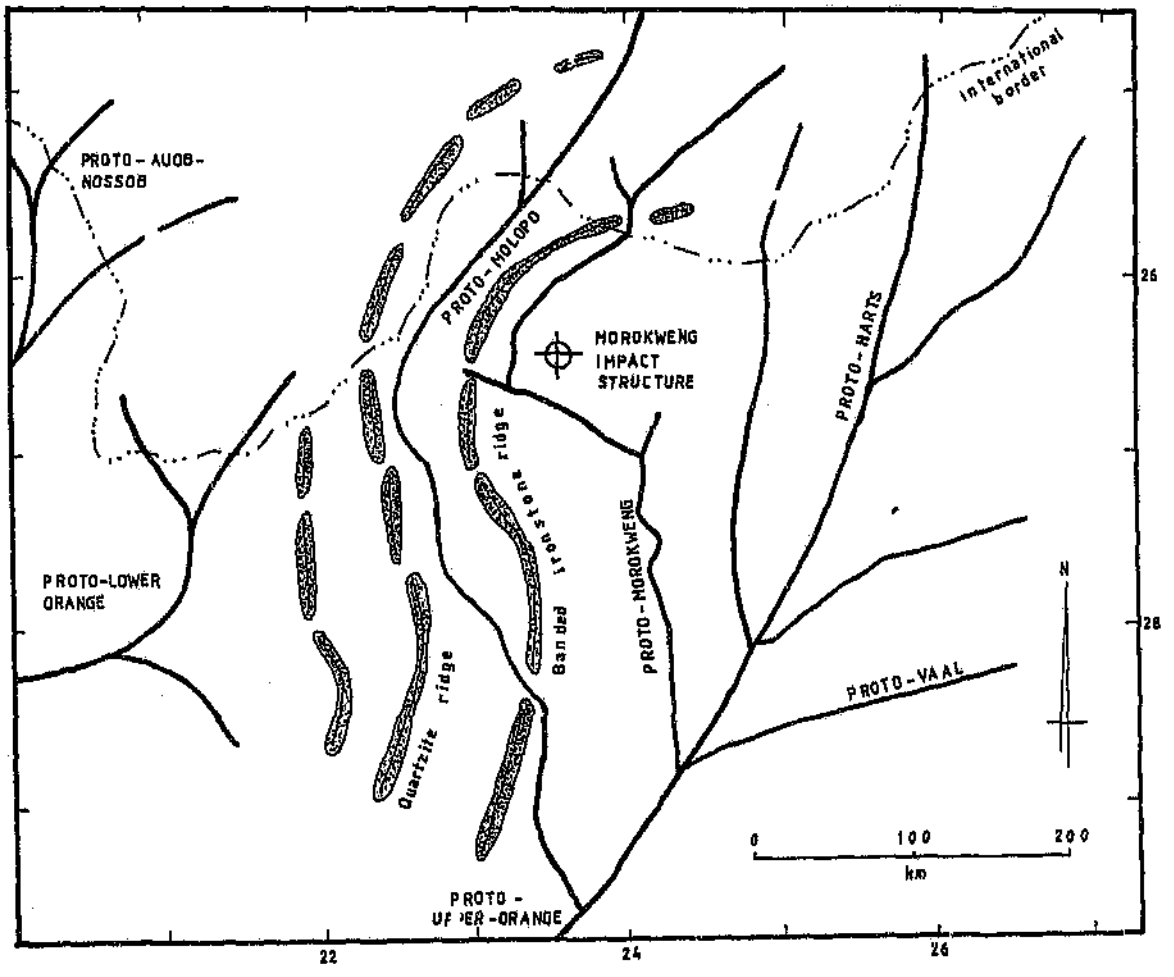


Figure 7.4 Middle to late Cretaceous drainage.

date of the active channel is fixed by the presence of silicified wood of an Upper Cretaceous species. Southward flow is indicated by fluvial structures and the provenance of well-rounded orange agates of the Ongeluk Formation. The presence of subangular clasts of the Kanguru layer in the channel deposits, indicate a reversed flow after the first basining movements had started. These basining movements disrupted the Morokweng River, and a temporary lake resulted north of the Ghaap Plateau. Overflow from this lake at some point broke through a gap in the Kuruman-Makhubung Hills range, to join the proto-Molopo as a tributary.

7.3.3 The proto-Molopo River

Concurrent with the proto-Morokweng River, the proto-Molopo existed further to the west between the erosion resistant ridges of the banded ironstone rocks of the Kuruman Hills-Makhubung range in the east, and the quartzites of the Langeberge-Korannaberg and its extensions across the border in Botswana to the west (Fig. 7.5). The major part of the sediment body of the Kalahari Group is interpreted as having formed in a back-tilted section of the pre-Kalahari proto-Molopo River (section 4.3). The proto-Molopo formed a major artery of the Karoo River (upper proto-Orange), which flowed through the Hotazel Valley, and used a gap through the Asbestos Hills to connect with the Karoo River south of Griquatown. Evidence at the confluence indicates that this proto-Molopo, which McCarthy (1983) termed the Trans-Tswana, could have carried a volume of water four times that of the present Orange River.

7.4 Mid-Tertiary Drainage Phase

The beginning of the basining of the Kalahari Basin at some time during the late Cretaceous terminated both the proto-Morokweng and proto-Molopo Rivers. The headwaters of both rivers became back-tilted with respect to their lower courses. The hinge points for the back-tilting were probably positioned close to the present position of the Griqualand-Transvaal axis of uplift. The location of the hinge points could therefore be seen as an incipient expression of the Griqualand-Transvaal axis. The position of the

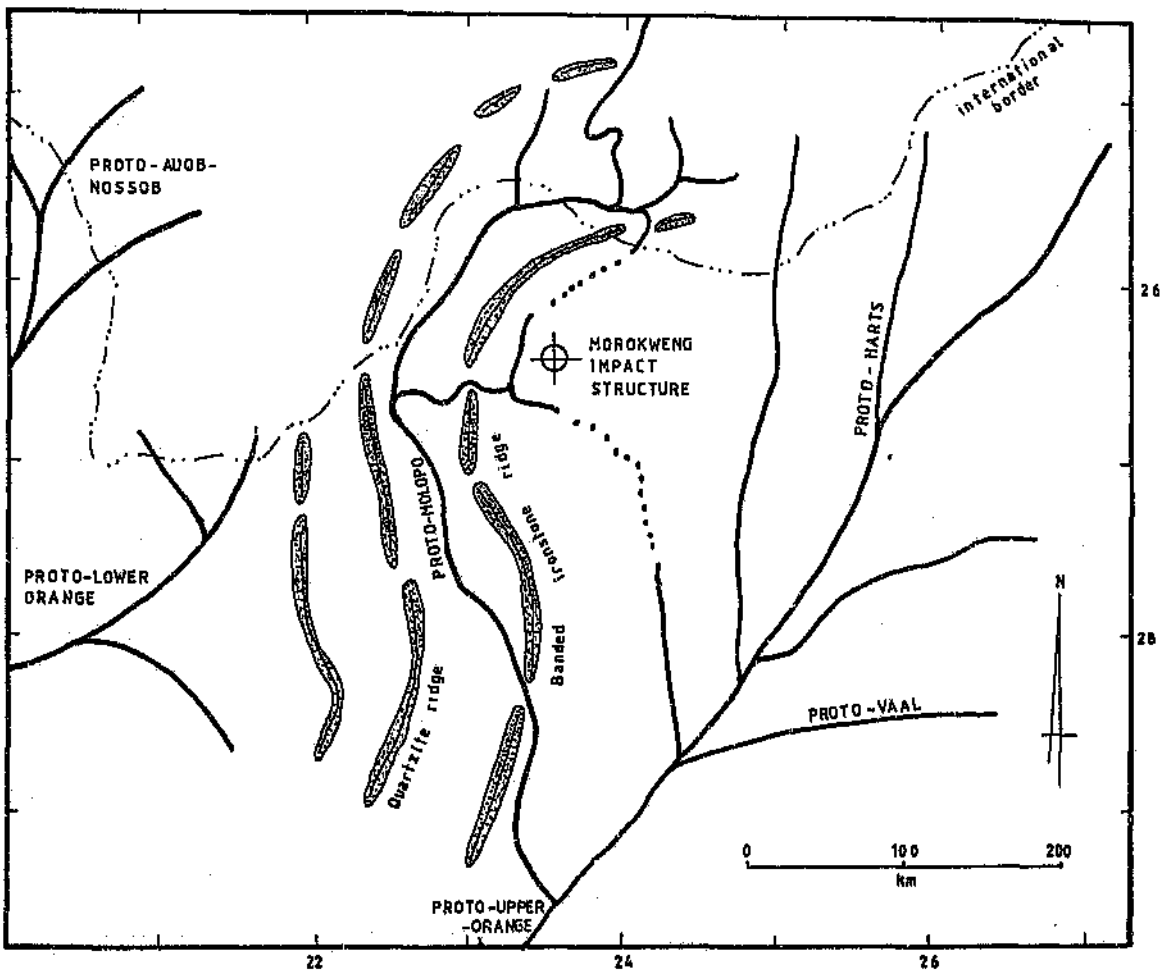


Figure 7.5 Late Cretaceous drainage.

swampy headwater areas of the Soutlcop River (Fig. 6.10), as well as the large pans (e.g. The Great Pan and The Glen), further point to the position of the incipient axis. The exoreic proto-Molopo drainage system had thus been converted into an endoreic lacustrine system.

7.4.1 Wessels Gravel Formation

The disruption of these south-flowing rivers must have resulted in sizeable lakes upstream of their hinge points (Fig. 7.6), as well as reverse river flows from the hinge points towards the newly formed lakes. It is not clear how this basining would have affected the headwaters of the Trans-Tswana further north of the Molopo drainage. The upper Molopo extended in an eastward direction, in a topographical low, which had started to develop between the incipient forms of the Griqualand-Transvaal and Kalahari-Zimbabwe axes of uplift (Bootsman, 1997). During this drainage diversion the diamondiferous gravels of the Lichtenburg area were relocated from the upper proto-Harts system into the proto-Molopo drainage. With the general westward tilting, the water accumulated around the Morokweng impact structure, and eventually broke through a gap in the Kuruman-Makhubung hills near Severn, to join the proto-Molopo as a tributary. Sediments washed into the newly formed lakes and this initially resulted in the formation of the clayey gravels of the Wessels formation, which in some places is over 100 m thick. Diamondiferous gravels, which were earlier transported to the south through the proto-Molopo, now became trapped in this back-tilted section.

7.4.2 Playa environments

Significant desiccation which started at the end of the Cretaceous (Partridge and Maud, 1989), would have changed the water balance of the lake and introduced playa-like conditions, over a length of about 400 km in the back-tilted section. It is envisaged that during a substantial part of the middle Tertiary, such a playa environment persisted. The red-brown Budin Clay Formation, which in places exceeds 100 m in thickness testifies of these conditions. The chemistry of the interconnected string of playas within this 400 km

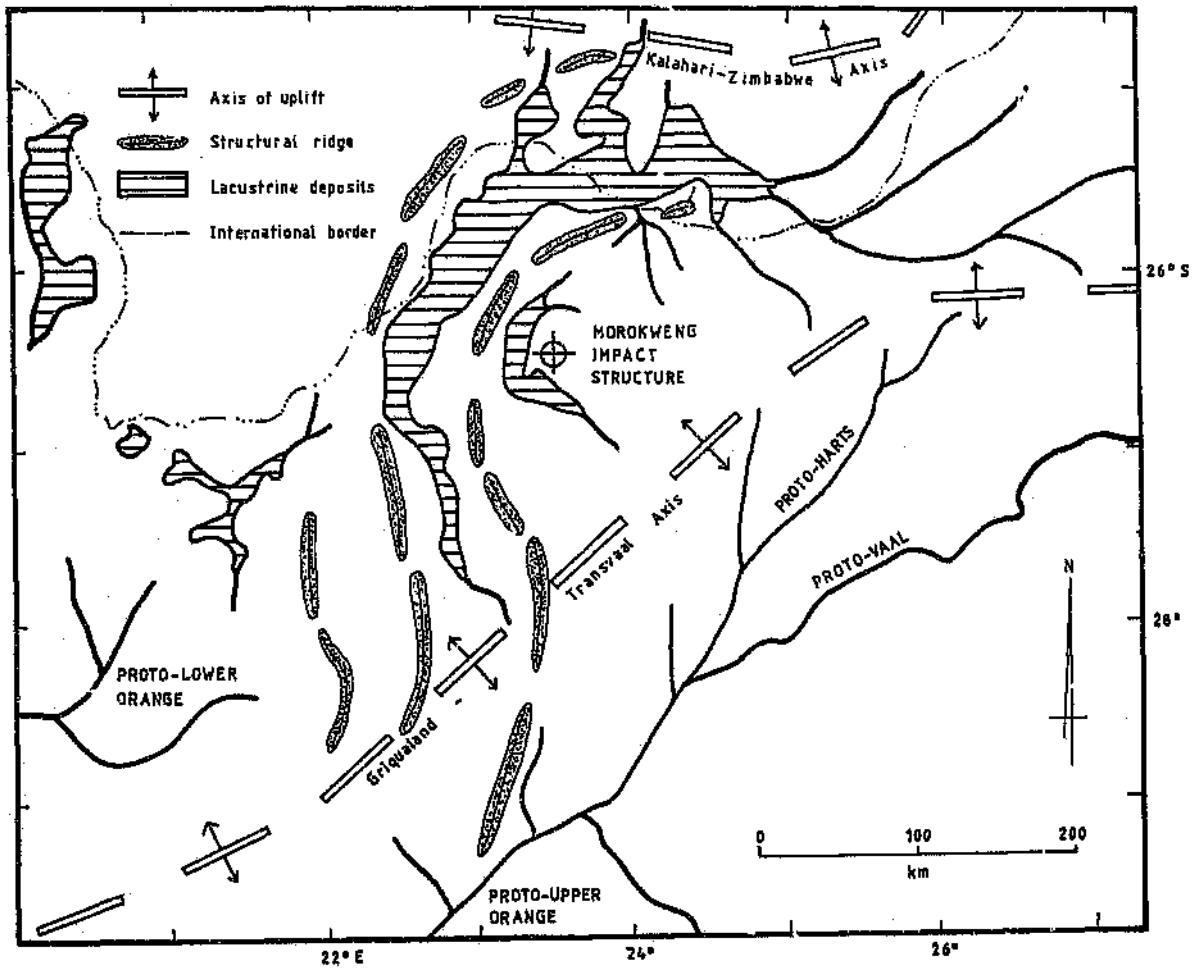


Figure 7.6 Mid-Tertiary drainage.

stretch, would have differed according to local conditions, as was indicated by the chemical analyses of clays.

7.4.3 Rapid infilling of the lacustrine environment

The clays became eventually covered with sandy deposits indicating a coarsening upward sedimentary sequence. It appears that the marginal facies of the playa environment now encroached upon it, indicating that sediment supply, probably under more arid and stable tectonic conditions had increased greatly. These sandy deposits belong to the Eden Sandstone Formation (SACS, 1980). Climatic conditions during the deposition of the sediments also favoured the formation of extensive duricrusts.

Mamatwan rejuvenation phase

The sedimentary sequence of the Kalahari Group exposed in the Mamatwan manganese mine shows one clear channelling phase (unit iii, as described in section 6.3.1), where channels were eroded into the sandy deposits. The cause of this rejuvenation may have been tectonic and/or climatic. An epeirogenic uplift shortly after about 20 Ma, which ended the African erosion cycle has been reported by Partridge (1997b). It was suggested that the Kalahari Basin lagged behind during this epeirogenic uplift with respect to the surrounding areas, thus increasing the gradient. A return to more humid conditions in the early mid-Miocene (about 16 Ma) has however, also been reported by Dingle and Hendey (1984) and de Wit (1993). The channels had a clast-supported gravel lag and a fining-upward sedimentary sequence form there onwards can be observed in the Mamatwan mine. Support for a rejuvenation phase comes from the occurrence of a well-developed gravel-rich layer at depths between 37 and 44 m in the deep borehole (Sacha W216) on the Sacha farm near Kathu. The conglomerate occurs within a 60 m thick calcrete layer and is made up of well-rounded gravel, some of which is heavily weathered. It appears that the originally clast-supported conglomerate has changed into a matrix-supported conglomerate due to heavy calcretisation.

The greenish duricrusted sandy sediments with laminar structures above the gravel layer in the Mamatwan mine suggest a return to pan environments, towards the end of the deposition of the Eden Sandstone Formation. Similar green silcretes are common around the margins of the playas of the Northern Cape Province and Botswana (Summerfield, 1983). Within the Molopo drainage area, such green silcretes have been described from Heuningvleipan (Summerfield, 1982).

7.5 Plio-Pleistocene Drainage Phase

A Pliocene uplift has been dated between 5 and 3 Ma (Partridge, 1997b). Arid conditions prevailed generally during the late Neogene in the Kalahari, which is evidenced by the presence of dolomitic calcretes and silcretes that are normally associated with alkaline lakes or pans. The Pliocene uplift resulted in an increased westward gradient of the southern African subcontinent and uplift in the interior was concentrated along the Griqualand-Transvaal axis. This resulted in a westward shift of the entire Molopo drainage. The Kalahari Basin before the late Neogene tectonics must have been a level region, with the upper section of the Kalahari Group sediments duricrusted. With the tectonic activity, the Molopo drainage responded to follow new gradients determined by the pattern of basining in the Kalahari. A resurgence of old tectonic structures clearly influenced the sediment thickness variations in the Kalahari Basin (e.g. influence of the Tshane hingeline; section 4.4.2). During this period the upper Molopo, Kuruman, and their tributaries started to follow a definite westward direction, and crossed at almost right angles the sediment-filled course of the proto-Molopo. Only the middle and lower sections of the Molopo River managed to maintain a general southward flow.

7.5.1 Humid pulse and a re-established exoreic Molopo drainage

Not only tectonics played a role during the Pliocene but significant climatic changes also occurred. Warmer and moister conditions have been inferred between 5.1 and 4.2 Ma and 3.5 and 2.8 Ma (Partridge, 1997b). It is suggested that the new drainage pattern of the Molopo in response to tectonics was aided by these moister periods, which changed the Molopo drainage from an endoreic system again back to an exoreic system. The

confluence point with the Orange river was this time much further west, about 17 km downstream of the Augrabies Falls (Fig. 7.7).

A joint Auob/Nossob/Kuruman/Molopo System

It appears that the Auob/Nossob Rivers went through a similar evolution as the Molopo drainage, where the proto-Auob/Nossob initially formed a major tributary of the Kalahari River (de Wit, 1993) or lower proto-Orange. It appears that also the Auob/Nossob system became back-tilted with reference to the rim of the Kalahari Basin and was subsequently completely filled in with Kalahari Group sediments. A synthesis of the changing courses of the Molopo/Kuruman drainage (Fig. 7.8) shows a definite shift towards the central axis of the Kalahari Basin. In response to both tectonics and climatic change the new courses of the Auob/Nossob similarly crossed the courses of the proto-system at near right angles, and a major part of the Auob/Nossob drainage system also shifted towards the centre of the Kalahari, but in an eastward direction. The combined Auob/Nossob/Kuruman/Molopo drainage system became thus re-established as an exoreic drainage system (Fig. 7.9), with sufficient energy to cut through the southerly rim of the Kalahari Basin. With the deeper basining of the Kalahari the two river systems shifted towards the central NE-SW axis of the Kalahari Basin, to find a common exit, in response to the surviving overall N-S gradient of the southern Kalahari. This common exit was probably facilitated by a pre-existing tributary of the lower Orange near Riemvasmaak, just downstream of the Augrabies Falls. The combined Molopo and Nossob drainages formed a major drainage system, which contributed large volumes of water to the Orange River downstream of the falls, and resulted in a relatively quick knick point migration up the lower Molopo, which cut the impressive, now mainly dry, gorge near Riemvasmaak.

The joining of the Molopo and Nossob drainages was a response to both tectonics and climatic change. The shift towards the NE-SW structural line of the Kalahari Basin was due to tectonics, while the change from endoreic to exoreic system, as well as the incision

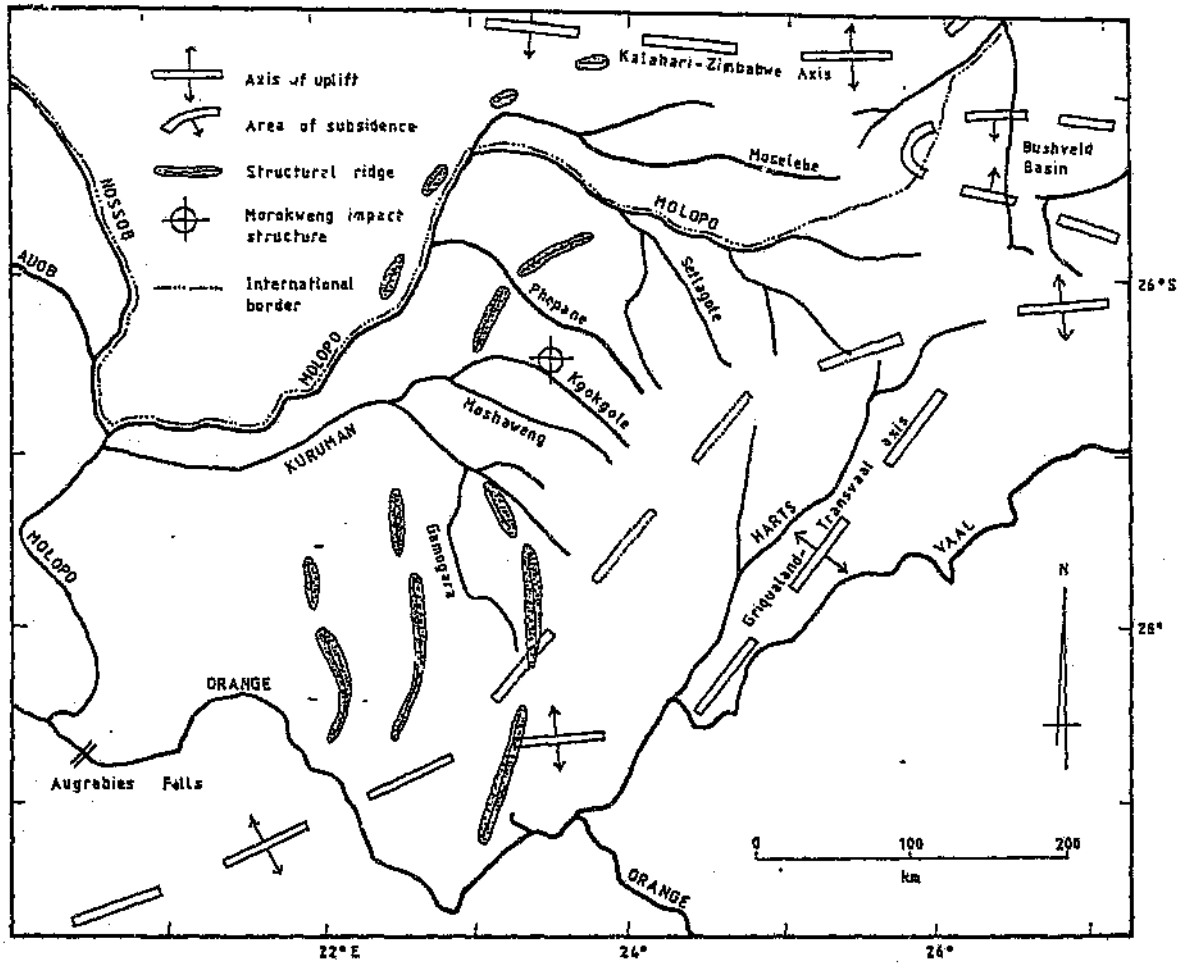


Figure 7.7 Late Tertiary drainage.

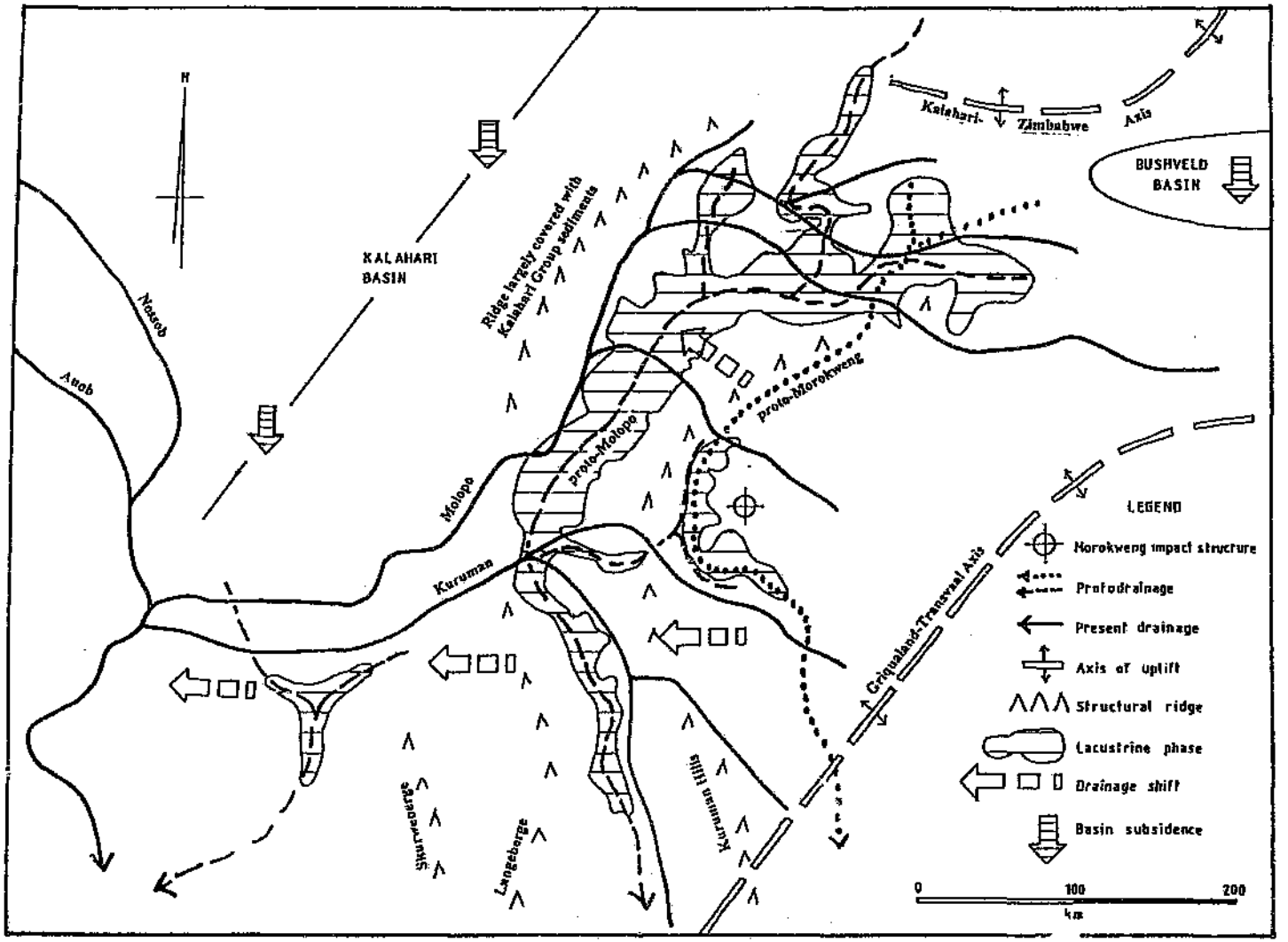


Figure 7.8 Synopsis of Molopo drainage shifts.

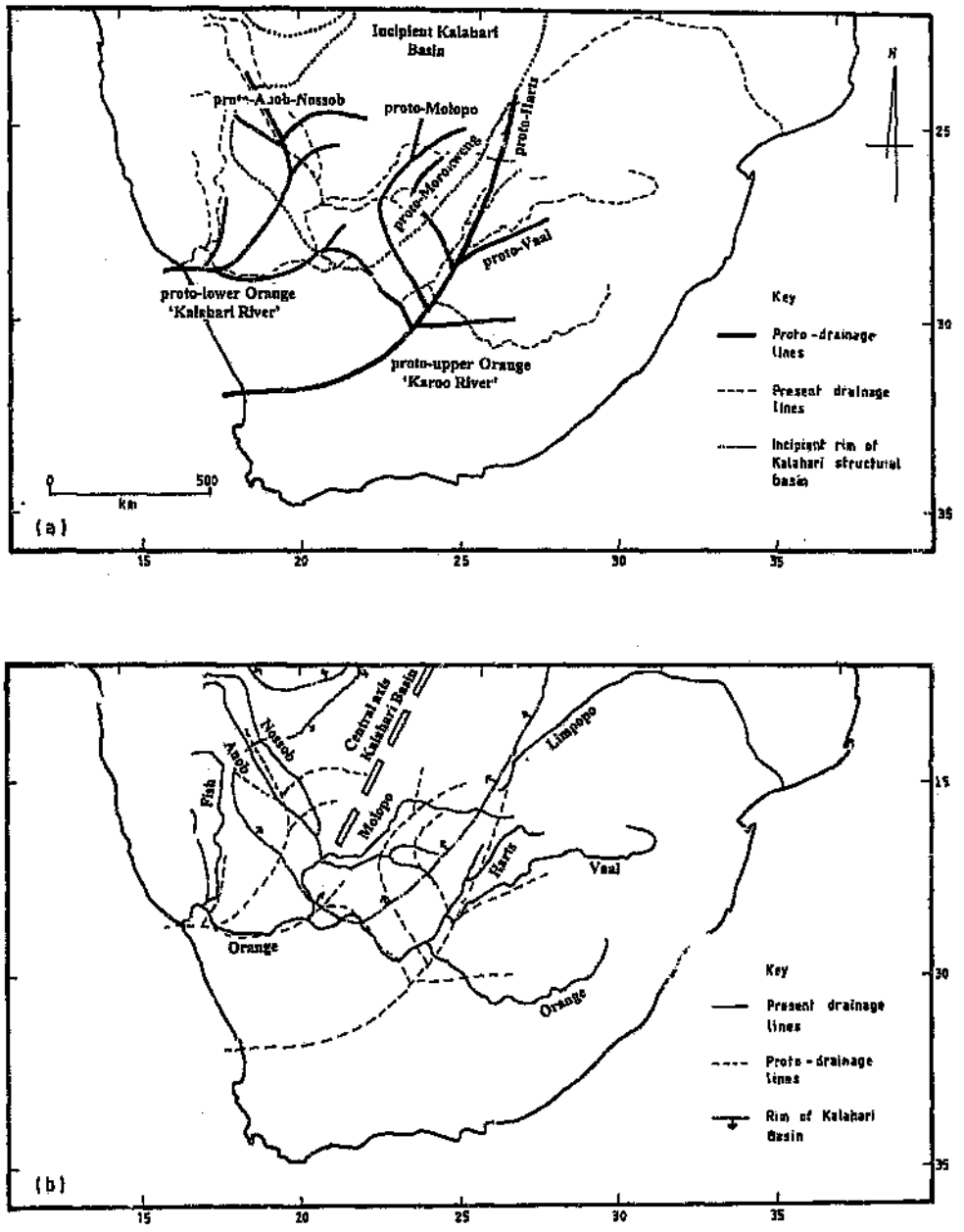


Figure 7.9 Palaeodrainage reconstructions for (a) late Cretaceous and (b) Quaternary. Karoo and Kalahari Rivers based on de Wit (1993).

of the Molopo and Nossob drainage systems into the duricrusted Kalahari Group sediments, should be ascribed to increased rainfall.

A renewed period of cooling and aridity began about 2.8 Ma ago (Partridge, 1997a). The result of this global interval of cooling was the spreading of the Kalahari sand sea to form the mega-Kalahari. Discontinuous dunes spread far beyond the limits of the Kalahari Basin as far south as the Cape Fold Mountains (Partridge and Maud, 1987; Partridge, 1997a). It is inferred that perennial flow was terminated from that time onwards, to occur again only periodically in relation to the climatic changes of the Pleistocene. It has been suggested that during this period of aridity encroaching dunes caused the lower course of the Molopo drainage to become defunct (Partridge, 1997b). This would mean that once again the Molopo drainage had become endoreic. Absolute dating of these events is still lacking and the events can presently only be slotted in, into the general time frame of the geomorphological evolution of southern Africa.

7.6 Quaternary Drainage Conditions

Globally the Pleistocene has been dominated by recurring glacial and interglacial intervals, but well-calibrated palaeoenvironmental data for the early and middle Pleistocene is very limited in southern Africa. It appears that cool glacial maxima correlated with greater aridity in the Kalahari (Partridge, 1997b), and that during such times the dune field became reactivated. Limited palaeoenvironmental data of the later Pleistocene within the Molopo drainage is available from the Lobatse Caves, the Wonderwerk Cave and the Kathu Pan doline and could help to determine periods of perennial flow during the late Pleistocene.

7.6.1 Terraces

Lower terraces have been reported by a number of authors and have been linked with climatic changes during the Pleistocene, but also here absolute dating is lacking. The terraces are younger than the incision of the Molopo drainage into the duricrusted

Kalahari Group sediments. Heine (1982) indicated the last possible perennial flow in the lower Molopo between 17 000 and 15 000 BP and around 12 000 BP.

7.7 Demise of the Molopo Drainage

The Molopo drainage is at present a defunct drainage system with very limited spring flow in the upper sections of the Molopo and Kuruman River and periodic flows along certain sections of it, due to high-intensity precipitation events from summer thunderstorms. It is still unclear when integrated flow last occurred in this drainage system, but most likely during interglacial periods with generally moister conditions. It appears then that after the Pliocene humid period(s) of incision, pulses of river activity occurred during the Pleistocene, in response to climatic ameliorations during the interglacials.

* * * * *

A summary of the drainage evolution as well as the conclusions and implications of this model of the Molopo drainage are contained in the next and last chapter.

CHAPTER 8 SUMMARY, CONCLUSIONS, AND IMPLICATIONS

8.1 Summary of the Evolution of the Molopo Drainage

The analysis of the Molopo drainage has indicated that the evolution of the Molopo drainage is complex, and that it is the result of a long-term interaction between tectonics and climatic change (Table 8.1). There is evidence of great antiquity of drainage activity in the Molopo drainage area. The earliest drainage indicators (more than 300 Ma) come from a pre-Karoo valley system, that has been preserved through the in-fill of Permo-Carboniferous glacial sediments (Dwyka Formation) under a cover of Cainozoic Kalahari Group sediments.

The fragmentation of Gondwana around the beginning of the Cretaceous resulted in the determination of the general coastline of the present southern African subcontinent, with most of the landsurface between 2000 and 2500 m AMSL. This condition triggered massive erosion, especially in the coastal areas and is considered as the start of the formation of the African planation surface. The beginning of the Cretaceous was also the time of a large meteorite impact near Morokweng, in the central eastern section of the present Molopo drainage, which had a substantial influence on the existing drainage lines in the area.

A more detailed picture of the various drainage phases of the Molopo drainage on the basis of remnant landforms and sediments can, however, be reconstructed from the

middle to late Cretaceous onwards. The long-term development of the Molopo drainage during the Cainozoic can be explained both in terms of tectonics and climatic change. The initial basining during the late Cretaceous may well have been an isostatic response related to the deposition of large quantities of sediment in the coastal waters around the subcontinent, but late Neogene uplifts have also been related to a large negative S-wave velocity within the lower mantle below southern Africa; the Great African Plume.

Table 8.1: Indicators to the evolution of the Molopo River (Modified from Bootsman, 1997).

Geomorphic manifestation / event	Landscape / sedimentary evidence	Main influence tectonics/ climate/other	Estimated Age
Demise of Molopo River.	Meander-in-meander landforms, sandy-clay floodplain backfill. No perennial flow in living memory.	Climate	late Pleistocene to Holocene
Lower terraces.	Rock-cut terrace remnants ~ 6 and 14 m above floodplain level.	Climate	Pleistocene
River rejuvenation.	20 - 30 m incision of Molopo River in present position. Reactivated palaeodrainage lines. Re-instated exoreic drainage toward Orange River.	Climate	Pliocene (5.1-4.2 Ma and/or 3.5-2.8 Ma)
Shift of Molopo	Most westerly position of Molopo River	Tectonic	5-3 Ma
Hydrological barrier formed by exposed Malmani Dolomites and lowering of groundwater levels.	Calcretised sediment-filled palaeo-drainage lines.	Tectonic	~ Tertiary
Bushveld Basin subsidence.	River piracy and drainage reversals by the Limpopo River.	Tectonic	End of early Miocene (~ 18 Ma)
Endoreic drainage phase (palaeo-Lake Molopo) created through back-tilting of headward drainage section. Expanded upper Molopo.	Isopach pattern of Kalahari group sediments. Thick deposits of Wessels Gravel formation, and lacustrine clays of Budin Clay Formation.	Tectonic ?	late Cretaceous to mid-Tertiary ?
Pre-Kalahari proto-Molopo drainage.	Isopach pattern of Kalahari Group sediments	Climate	Upper Cretaceous
Morokweng meteorite impact	Annular drainage components.	Extra-terrestrial	Jurassic-Cretaceous boundary (145±2 Ma)
Dwyka glaciation.	Nature of basal tillite and glacio-marine, glacio-fluvial and glacio-lacustrine sediments	Climate	Permo-Carboniferous (~ 300 Ma)
Pre-Dwyka river valley	Pattern of Dwyka Tillite suboutcrops under Kalahari Group sediment cover.	Climate	Carboniferous

Evidence of a mid to late Cretaceous, pre-Kalahari drainage phase comes from a reconstructed pre-Kalahari surface, beneath the Kalahari Group sediments. The interpretation of this pre-Kalahari surface must include a correction for the total surface lowering, that has occurred after the Kalahari basining had started. Without this correction an erroneous interpretation of drainage direction can result. The proto-Molopo and proto-Morokweng formed a south-flowing system, which joined the upper proto-Orange near the present Vaal/Orange and Vaal/Harts confluences respectively. The African erosion surface, which was initiated at the time of the break-up of Gondwana, was towards the late Cretaceous almost completely levelled, and these low-gradient proto-rivers flowed on the nearly-completed African erosion surface.

The disruption of both the proto-Molopo and proto-Morokweng Rivers during the late Cretaceous-early Tertiary, indicated the end of the pre-Kalahari drainage phase, and the start of the infilling of the Molopo sub-basin, as formed by the back-tilting of the headwaters of the protodrainage system. The proto-Molopo had now become an endoreic system with a radial inflow. The flow direction of a section of the proto-Molopo above the hinge point of the back-tilting mechanism, now experienced a reversal of flow direction into the endoreic system. Concurrently with the deepening Kalahari Basin a topographic low formed, which allowed the upper Molopo to extend itself by about 200 km in an eastward direction. During this phase diamondiferous gravels of the neighbouring proto-Harts system were reworked and redirected towards the sump of the Molopo sub-basin.

The sub-basin became filled up with substantial amounts of gravel, clay and sand under generally aridifying conditions from the Cretaceous onwards. Neogene tectonics during the mid Miocene resulted in the lowering of the Bushveld Basin and the warping of the intra-continental axes. The post-African surface was initiated and resulted in the erosion of the African surface. The Malmani Dolomites emerged from beneath a Karoo cover to become a hydrological barrier for most of the extended upper Molopo. The result was a

much shortened upper Molopo and a drowning of the endoreic drainage system in its own sediment. Generally aridifying conditions, which had started since the late Cretaceous, encouraged the formation of duricrusts in the Kalahari Group sediments. A major rejuvenation phase under more humid conditions resulted in such a reactivation of the Molopo drainage, that once again it became an exoreic drainage system, but now together with the Auob/Nossob system. The Molopo had shifted westward due to Pliocene warping of the Griqualand-Transvaal axis, while the Auob/Nossob had shifted eastward to the central axis of the deepening axis of the Kalahari Basin, to find a common exit along a gentle north-south gradient across the southern basin rim. A return to more arid conditions during the Pliocene enhanced aeolian activity, and the red Kalahari Sand spread far beyond the margins of the Kalahari Basin.

Fluctuating climatic conditions during the Q ary resulted in limited episodes of perennial flows, which gave rise to the formation of rock-cut terraces in the upper Molopo west of Mafikeng. At present the Molopo drainage is largely an inactive drainage system since integrated perennial flow has not been known during living memory, and the lower Molopo has been blocked by sand dunes for a considerable time.

8.2 Major Conclusions

In this study of the integrated Molopo drainage system it has been shown that the evolution of the drainage is complex and involves the influence of both tectonics and climatic change. The position of the Molopo drainage across the south-eastern rim of the Kalahari Basin and near the intra-continental axes of uplift has strongly influenced its evolution. There is further ample evidence that climatic change has greatly influenced the competence and nature of the river system. In general there has been a decrease in competence with evidence of fluctuating flows.

The most important conclusions of this study of the evolution of the Molopo drainage are presented in the same order as the objectives (section 1.2) as stated at the beginning of the

thesis. The shifted emphasis during the research towards the development of a larger scale model of the Molopo drainage evolution has, however, influenced the emphasis in the final conclusions.

8.2.1 Structural influences and the drainage evolution

(i) Resurgence of tectonic structures has occurred in several instances. The Cainozoic Kalahari Basin coincides to a large extent with the Kalahari (*Karoo*) Basin, which has been described as a half-graben, and there is evidence of earlier basins (possibly as old as Precambrian). The Cainozoic Griqualand-Transvaal axis of uplift coincides greatly with the axis of the Permo-Carboniferous Cargonian Highlands. The Griqualand-Transvaal axis of uplift is, therefore, considered to be a rejuvenation of the much older axis of the Cargonian Highlands. The Precambrian Tshane hingeline also coincides with a basement high in the Middelputs-Khuis area within the Cainozoic Kalahari Basin. Kalahari Group sedimentation within the Molopo sub-basin has been greatly influenced by the resurgence of older tectonic structures.

(ii) The meteorite impact near Morokweng at the Jurassic-Cretaceous boundary (145 Ma) created a circular ring structure of at least 70 km and possibly as much as 340 km in diameter and coincided with a dramatic change in the general drainage direction of the area. The drainage patterns of both the proto-Molopo and the proto-Morokweng have annular components which seem related to the meteorite impact. The full effect of the meteorite impact on the drainage evolution is the subject of further research.

(iii) The pre-Kalahari proto-Molopo drainage lines showed a strong relationship with geological structures (e.g. fault lines and meteoritic impact structure). With the filling of the Kalahari Basin this affinity was greatly reduced.

(iv) Westward tilting of the sediment-surface of the Molopo sub-basin towards the centre of the Kalahari Basin occurred due to Neogene warping of the Griqualand-Transvaal axis. Miocene and Pliocene uplifts further accentuated the lowering of the Kalahari Basin. The Miocene uplift affected the sedimentation in the sub-basin, and Pliocene uplift caused a substantial westward shift of the Molopo drainage.

(v) The existence of three parallel ridges of erosion resistant rocks (Asbestos Hills-Kuruman Hills-Makhubung range, Lan[illegible]erge-Korannaberg range, and Skurweberge) have determined the position of the Molopo drainage at various stages during its westward shifts in respond to tectonics.

8.2.2 The diamondiferous gravels of the Mafikeng-Bakerville-Lichtenburg area

(i) The diamondiferous gravels of the headwaters of the proto-Harts drainage were reworked by the extended upper Molopo, and redirected towards the endoreic Molopo drainage.

(ii) Mid-Miocene lowering of the Bushveld Basin and the emergence of the Malmani Dolomites from underneath a Karoo rock cover caused a reduction in the extension of the upper Molopo, and eventually stopped the transport of the diamondiferous gravels.

8.2.3 The Molopo drainage and the Griqualand-Transvaal and Kalahari-Zimbabwe axes of uplift.

(i) The 200 km eastward extension of the upper Molopo is due to a topographic low or tectonic trough, which developed between the incipient forms of both the Kalahari-Zimbabwe and the Griqualand-Transvaal axes of uplift. An extension of the Kalahari Basin formed between these incipient axes simultaneously with the deepening Kalahari Basin.

(ii) An incipient form of the Neogene Griqualand-Transvaal axis of uplift determined the position of the south-easterly section of the rim of the Kalahari Basin. Due to tectonic backtilting of the headwaters of the proto-Molopo in relation with the rim of the Kalahari Basin, the Molopo sub-basin, as an endoreic drainage system, has been a sediment-trap for a considerable period of the early and middle Tertiary,

(ii) The proto-Molopo formed the lower section of McCarthy's (1983) Trans-Tswana river. Warping of the Griqualand-Transvaal axis disrupted the Trans-Tswana drainage. The relationship between the Trans-Tswana and the Kalahari-Zimbabwe axis is not clear. The buried valley of the proto-Molopo suggests that the upper proto-Limpopo formed part of the Trans-Tswana, rather than the earlier suggested remains of a drainage line coming direct from the Makgadikgadi Depression and the Okavango Delta.

8.2.4 Climatic fluctuations and terrace formation

Terrace formation has not been the emphasis of this research but the following conclusions regarding climatic change and its influence on the Molopo drainage evolution were reached.

(i) A decreasing clast size can be observed from the oldest reworked late Cretaceous gravels in the upper Molopo area, through the lower terrace gravels to the sandy-clay backfill in the present Molopo River floodplain. The evolution of the Molopo drainage in general is characterised by a decreasing competence from late Cretaceous to the present.

(ii) A major rejuvenation phase is connected with a humid climatic pulse which occurred after the infilling of the sub-basin, the formation of major duricrusts, and the westward tilting of the Molopo drainage area. The rejuvenation resulted in the flushing out of a section of the upper Molopo, the incision of the Molopo drainage into the duricrusted Kalahari Group sediments, and the re-establishment of the

Molopo drainage as an exoreic drainage, by cutting a valley through the southern rim. The timing of the rejuvenation is placed in the Pliocene after the warping of the Griqualand-Transvaal axis, but before the spreading of the red Kalahari Sands.

(iii) The sediment-filled Molopo sub-basin and the preserved headwater sections of the palaeo-Molopo (calcrete tongues) in the upper Molopo drainage area provide evidence of at least one major episode of complete sediment infilling of the Molopo drainage. The infilling of the drainage was due to an increased sediment supply in relation to the lowering of the Kalahari Basin. The increased sediment supply is related to aridification during the Tertiary.

8.2.5 Duricrust formation and the Molopo drainage

(i) Thick duricrusts, mainly calcretes, developed in the sediments of the of the Molopo sub-basin under aridifying conditions during the Cainozoic. Although undoubtedly the majority of the calcretes are of the pedogenic type, the thick calcretes that have developed as calcrete tongues from the Malmani Dolomites in the sediment-filled drainage sections of the Molopo, primarily developed because of channelled groundwater flows under semi-arid to arid climatic conditions.

(ii) The thick duricrusts developed in the sediment filled drainage lines caused lateral drainage shifts to the margins of the crusts during more humid climatic periods.

8.2.6 The Kalahari Group stratigraphy in the Molopo drainage area

(i) The Molopo sub-basin constitutes a discrete sub-basin of the Kalahari Basin. The stratigraphy as discussed by Smit (1977), SACS (1980), and Botha *et al.*(1986) has been determined by regional tectonics and the Molopo drainage evolution.

8.3 Implications

Implications of the results of this research are presented under three sections below, namely, for regional-scale geomorphology, for subcontinental-scale geomorphology, and for further studies within the Molopo drainage area.

8.3.1 Implications for regional-scale geomorphology

(i) The roles of both tectonics and climatic change should be considered in a reconstruction of the evolutionary history of drainage lines in the subcontinental interior of southern Africa.

(ii) The evolution of the Auob-Nossob tributary drainage system of the Molopo has experienced a very similar history of drainage basin back-tilting, sediment infilling, shift of drainage lines, rejuvenation, and demise. The Auob/Nossob system shifted in mirror fashion towards the central line of the Kalahari Basin.

8.3.2 Implications for subcontinental-scale geomorphology

(i) The proto-Molopo was a major south-flowing tributary of the upper proto-Orange, and possibly the main branch prior to the formation of the Kalahari Basin. The proto-Auob/Nossob was a major south-flowing tributary of the lower proto-Orange prior to the formation of the Kalahari Basin. The formation of the Kalahari Basin thus caused a major disturbance of the south-flowing protodrainage lines of the interior of the southern African subcontinent, which resulted finally in a joint Auob/Nossob/Kuruman/Molopo drainage system

8.3.3 Further studies in the Molopo drainage basin

The focus of this investigation into the development of the Molopo drainage has been mainly on the larger structural aspects and climatic framework. Further testing and fine-tuning of this model of the evolution of the Molopo drainage will proceed through: in-depth clast lithological analysis of the gravels in the upper Molopo and Griqualand West

areas, C-14 dating of shells as indicators of perennial flow in the Molopo system, pollen analysis of the sandy-clay backfill that is generally present in the floodplain of the Molopo drainage, and an investigation into the role of local knickpoints on the formation of silcretes along the Molopo River.

REFERENCES

Altermann, W. and Siegfried, H. P., 1997: Sedimentology and facies development of an Archaean shelf: carbonate platform transition in the Kaapvaal Craton, as deduced from a deep borehole at Kathu, South Africa, *Journal of African Earth Sciences*, 24, 391-410.

Andreoli, M. A. G., Ashwal, L. D., Hart, R. J., Smith, C. and Hambleton-Jones, B. B., 1995: The impact origin of the Morokweng ring structure, southern Kalahari, South Africa, in: *Extended Abstracts Centenary Geocongress (Geocongress '95), Geological Society of South Africa*, Johannesburg 541-543.

Anthony, R., 1996: *Vacation Work at Mamatwan Mine*, Unpublished project report, Samancor, Hotazel.

Arakel, A. V., 1986: Evolution of calcrete in palaeodrainages of the Lake Napperby area, central Australia, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 54, 283-303.

Arakel, A. V., 1991: Evolution of Quaternary duricrusts in Karinga Creek drainage system, central Australian groundwater discharge zone, *Australian Journal of Earth Sciences*, 38, 333-347.

Arakel, A. V. and McConchie, D., 1982: Classification and genesis of calcrete and gypsite lithofacies in palaeodrainage systems of inland Australia and their relationship to carnotite mineralization, *Journal of Sedimentary Petrology*, 52, 1149-1170.

Arakel, A. V., Jacobson, G., Salehi, M. and Hill, C. M., 1989: Silicification of calcrete in palaeodrainage basins of the Australian arid zone, *Australian Journal of Earth Sciences*, 36, 73-89.

Barnard, W. S., 1992: "Om die Ngami en Makarikari meren opnieuw te vullen": Die Schwarzplan as dwaling, *The South African Geographer*, 19(1/2), 59-75.

Beaumont, P. B., Van Zinderen Bakker, E. M. and Vogel, J. C., 1984: Environmental changes since 32 000 BP at Kathu Pan, Northern Cape, in: Vogel, J. C. (ed.), *Late*

Cainozoic Palaeoclimates of the Southern Hemisphere, Balkema, Rotterdam, 329-338.

Behr, S. H., 1989: The thickness off the Kalahari Group sediments in the proto-Molopo river basin. *Abstracts, 9th Biennial SASQUA Conference*, Durban, p. 22.

Bishop, P., 1995: Drainage rearrangement by river capture, beheading and diversion, *Progress in Physical Geography*, 19 (4), 449-473.

Bishop, P. and Brown, R., 1992: Denudational isostatic rebound of intraplate highlands: The Lachlan River Valley, Australia, *Earth Surface Processes and Landforms*, 17, 345-360.

Blank, H. R. and Tynes, E. W., 1965: Formation of caliche in situ, *Bulletin of the Geological Society of America*, 76, 1387-1392.

Blatt, H., Middleton, G., and Murray, R., 1972: *Origin of Sedimentary Rocks*, Prentice-Hall, Englewood Cliffs.

Boocock, C. and van Straten, O. J., 1962: Notes on the geology and hydrogeology of the central Kalahari region, Bechuanaland Protectorate, *Transactions of the Geological Society of South Africa*, 65, 125-171.

Bootsman, C. S., 1997: On the evolution of the upper-Molopo drainage, *South African Geographical Journal*, 79(2), 83-92.

Bootsman, C. S. and Reimold, W. U., 1998: The significance of the Morokweng meteorite impact for the long-term Molopo drainage evolution (extended abstract), in: Rowntree, K. M. (ed.), *Proceedings of the Southern African Association of Geomorphologists Biennial Conference, 28th June - 1st July*, Grahamstown, 217-219.

Botha, B. J. V., Thomas, M. A., Malherbe, S. J., and Thomas, R. J., 1986: Aspects of the Tertiary and Quaternary geology of Gordonia, Northern Cape Province, *Annals of the Geological Survey of South Africa*, 20, 41-44.

Bowler, J. M., 1978: Quaternary climate and tectonics in the evolution of the Riverine plain, south-eastern Australia, in Davies, J. L. and Williams, M. A. J. (eds.), *Landform Evolution in Australia*, Australian National University Press, Canberra, 70-112.

Brinn, C., 1991: *Cainozoic Deposits of the Eastern Molopo Drainage Basin*, Internal report, Bophuthatswana Geological Survey, Mmabatho.

Bruno, S. A., 1985: Pan genesis in the southern Kalahari, in: Hutchins, D. G. and Lynam, A. P. (eds.), *The Proceedings of a Seminar on the Mineral Exploration of the Kalahari, October 1983*, Bulletin 29, Botswana Department of Geological Survey, Lobatse, 261-277.

Buckley, R., Wasson, R. and Chubb, A., 1987: Phosphorous and potassium status of arid dunefield soils in central Australia and southern Africa, and biogeographic implications, *Journal of Arid Environments*, 13, 211-216.

Burchell, W. J., 1824: *Travels in the Interior of Southern Africa*, Vol. 2, Longman, Hurst, Rees, Orme, Brown and Green, London.

Burollet, P. F., 1984: Intracratonic and pericratonic basins in Africa, *Sedimentary Geology*, 40, 1-11.

Butzer, K. W., 1984: Late Quaternary environments in South Africa, in: Vogel, J. C. (ed.), *Late Cainozoic Palaeoclimates of the Southern Hemisphere*, Balkema, Rotterdam, 235-264.

Cahen, L. and Lepersonne, J., 1952: Equivalence entre le systeme du Kalahari du Congo Belge et les Kalahari beds d'Afrique Australe, *Memoires du Societe Belge, Palaeontologie et Hydrologie*, 8, 1-64.

Cahen, L., Snelling, N. J., Delhal, J. and Vail, J. R., 1984: *Geochronology and Evolution of Africa*, Clarendon Press, Oxford.

Carter, D. B., 1954: Climates of Africa and India according to Thornthwaite's 1948 classification, *Laboratory of Climatology, Publ. Climat.*, 7, 455-474.

Clifford, T. N., 1970: The structural framework of Africa, in: Clifford T. N. and Gass, I. C. (eds.), *African Magmatism and Tectonics*, Oliver and Boyd, Edinburgh, 1-26.

Cooke, H. B. S., 1946: The development of the Vaal River and its deposits, *Transactions of the Geological Society of South Africa*, 49, 243-260.

Cooke, H. J., 1975: The Lobatse Caves, *Botswana Notes and Records*, 7, 29-33.

Cooke, H. J., 1980: Landform evolution in the context of climatic change and neo-tectonism in the middle Kalahari of northern central Botswana, *Transactions of the Institute of British Geographers*, NS 5, 80-99.

Cornell, D. H., 1985: Petrological stratigraphy of sub-Kalahari basaltic formations, in: Hutchins, D. G. and Lynam, A. P. (eds.), *The Proceedings of a Seminar on the Mineral Exploration of the Kalahari, October 1983*, Bulletin 29, Botswana Department of Geological Survey, Lobatse, 320-337.

Corner, B., 1994: Crustal Framework of the Kaapvaal province from geophysical data, *Abstracts Proterozoic Crustal and Metallogenic Evolution Conference*, Windhoek, p. 9.

Corner, B., Reimold, W. U., Brandt, D. and Köberl, C., 1997: Morokweng impact structure, North West Province, South Africa: geophysical imaging and shock petrographic studies, *Earth and Planetary Science Letters*, 146, 351-364.

Crockett, R. N. and Jones, M. T., 1975: Some aspects of the geology of the Waterberg System in eastern Botswana, *Transactions of the Geological Society of South Africa*, 78, 1-10.

Dardis, G. F., Beckedahl, H. R. and Stone, A. W., 1988: Fluvial systems, in: Moon, B. P. and Dardis, G. F. (eds.), *The Geomorphology of Southern Africa*, Southern Book Publishers, Johannesburg, 30-56.

de Wit, M. C. J., 1981: *A geographical Investigation and Geological Interpretation of Part of the Diamondiferous Gravels on the Farm Grasfontein*, Unpublished M. Sc. Thesis, University of Pretoria.

de Wit, M. C. J., 1993: *Cainozoic Evolution of Drainage Systems in the North-Western Cape*, Unpublished Ph. D. Thesis, University of Cape Town.

de Swardt, A. M. J. and Bennet, G., 1974: Structural and physiographic evolution of Natal since the Late Jurassic, *Transactions of the Geological Society of South Africa*, 77, 309-322.

Dingle, R. V. and Hendey, Q. B., 1984: Late Mesozoic and Tertiary sediment supply to the Eastern Cape basin (SE Atlantic) and palaeo-drainage systems in south-western Africa, *Marine Geology*, 56, 13-26.

Dingle, R. V., Siesser, W. G. and Newton, A. R., 1983: *Mesozoic and Tertiary Geology of Southern Africa*, Balkema, Rotterdam.

Dixey, F., 1938: Some observations on the physiographic development of central and southern Africa, *Transactions of the Geological Society of South Africa*, 41, 113-172.

Dixey, F., 1942: Erosion cycles in central and southern Africa, *Transactions of the Geological Society of South Africa*, 45, 151-181.

Dixey, F., 1955: Erosion surfaces in Africa: some considerations of age and origin, *Transactions of the Geological Society of South Africa*, 58, 265-280.

Droste, J. B., 1961a: Clay minerals in sediments of Owens, China, Searles, Panamint, Bristol, Cadiz and Danby lake basins, California, *Bulletin of the Geological Society of America*, 72, 1713-1722.

Droste, J. B., 1961b: *Clay Minerals in the Playa Sediments of the Mojave Desert, California*, Special report 69, California Division of Mines.

Dury, G. H., 1964: *Principles of Underfit Streams*, Geological Survey Professional Paper 452-A, United States Government Printing Office, Washington.

Dury, G. H., 1965: *Theoretical Implications of Underfit Streams*, Geological Survey Professional Paper 452-C, United States Government Printing Office, Washington.

du Toit, A. L., 1907: *Geological Survey of Portions of Mafeking and Vryburg*, Annual report of the Geological Commission of the Colony of the Cape of Good Hope, pp. 123-158.

du Toit, A. L., 1910: The evolution of the river system of Griqualand West, *Transactions of the Royal Society of Southern Africa*, 1, 347-362.

du Toit, A. L., 1916: Notes on the Karoo System in the southern Kalahari, *Transactions of the Geological Society of South Africa*, 19, 1-13.

du Toit, A. L., 1921: The Carboniferous glaciation of South Africa, *Transactions of the Geological Society of South Africa*, 24, 188-227.

du Toit, A. L., 1927: The Kalahari and some of its problems, *South African Journal of Science*, 24, 88-101.

du Toit, A. L., 1933: Crustal movements as a factor in the geographical evolution of South Africa, *South African Geological Journal*, 16, 3-20.

du Toit, A. L., 1951: *The Diamondiferous Gravels of Lichtenburg*, Memoir No. 44, Geological Survey, Pretoria.

du Toit, A. L., 1954: *The Geology of South Africa*, Oliver and Boyd, Edinburgh.

Frakes, L. A. and Crowell, J. C., 1969: Late Palaeozoic Glaciation: I, South America, *Bulletin of the Geological Society of America*, 80, 1007-1042.

Frakes, L. A. and Crowell, J. C., 1970: Late Palaeozoic Glaciation: II, Africa exclusive of the Karoo Basin, *Bulletin of the Geological Society of America*, 81, 2261-2286.

Fried, A. L. and Smith, N., 1992: Timescales and the role of inheritance in long-term landscape evolution, northern New England, Australia, *Earth Surface Processes and Landforms*, 17, 375-385.

Frodeman, R., 1995: Geological reasoning: Geology as an interpretative and historical science, *Bulletin of the Geological Society of America*, 107, 960-968.

Gale, S. J., 1992: Long-term landscape evolution in Australia, *Earth Surface Processes and Landforms*, 17, 323-343.

Gerrard, I., 1965: *The lower Karroo Succession at Lelanamagadi Pan, South-Western Bechuanaland, and its Correlation with the Molopo River Sequence*, Unpublished report of the Geological Survey of Bechuanaland.

Goudie, A. S., 1973: The geomorphic and resource significance of calcrete, in: Board, C., Chorley, R. J., Hagget, P. and Stoddart, D. R. (eds.), *Progress in Geography*, Vol. 5, Edward Arnold, London, 77-118.

Gould, D. and Rathbone, P. A., 1985: The geological structure of the Molopo Farms Project area, in: Hutchins, D. G. and Lynam, A. P. (eds.), *The Proceedings of a Seminar on the Mineral Exploration of the Kalahari, October 1983*, Bulletin 29, Botswana Department of Geological Survey, Gaborone, 160-186.

Gould, D., Rathbone, P. A. and Kimbell, G. S., 1987: *The Geology of the Molopo Farms Complex, Southern Botswana*, Volume 1 - Text, Botswana Department of Geological Survey, Bulletin 23, Gaborone.

Gould, D., Rathbone, P.A. and Kimbell, G. S., 1989: *The Geology of the Molopo Farms Complex, Southern Botswana*, Volume 2 - Appendices, Botswana Department of Geological Survey, Bulletin 23, Gaborone.

Green, D., 1966: *The Karroo System in Bechuanaland*, Botswana Department of Geological Survey, Bulletin 2, Lobatse.

Green, L. G., no date (± 1948): *To the River's End*, Timmins, Cape Town.

Greenwood, P. G. and Carruthers, R. M., 1973: *Geophysical Surveys in the Okavango Delta, Botswana*, Geophysical Division, Institute of Geological Science, London, Report 15, 1-23.

Grove, A. T., 1969: Landforms and climatic change in the Kalahari and Ngamiland. *Geographical Journal*, 135, 191-212.

Harland, W. B., Cox, A. V., Llewellyn, P. G., Pickton, C. A. G., Smith, A. G. and Walters, R., 1982: *A Geologic Time Scale*, Cambridge University Press, Cambridge.

Harland, W. B., Armstrong, R. L., Cox, A. V., Craig, L.C., Smith, A. G. and Smith, D. G., 1989: *A Geologic Time Scale*, Cambridge University Press, Cambridge.

Hart, R. J., Andreoli, M. A. G., Tredoux, M., Moser, D., Ashwal, L. D., Eide, E. A., Webb, S. J. and Brandt, D., 1997: Late Jurassic age for the Morokweng impact structure, southern Africa, *Earth and Planetary Science Letters*, 147, 25-35.

Haughton, S. H., 1927: Notes on the river system of south-west Gordonia, *Transactions of the Royal Society of South Africa*, 14, 225-231+ plates.

Haworth, R. J. and Ollier, C. D., 1992: Continental rifting and drainage reversal: the Clarence River of eastern Australia, *Earth Surface Processes and Landforms*, 17, 387-397.

Hecht, J., 1997: Big blast brought in the Cretaceous, *New Scientist*, 153, 21.

Hegenberger, W., and Seeger, K. G., 1980: *The Geology of the Gobabis Area: Explanation of Sheet 2218*, South West Africa/Namibia Department of Geological Survey, Windhoek.

Heine, K., 1978a: Radiocarbon chronology of late Quaternary lakes in the Kalahari, southern Africa, *Catena*, 5, 145-149.

Heine, K., 1978b: Jungquartäre pluviale und interpluviale in der Kalahari (südliches Afrika), *Palaeoecology of Africa*, 10, 31-39.

Heine, K., 1979a: Reply to Cooke's discussion of "K. Heine: Radiocarbon chronology of the late Quaternary lakes in the Kalahari", *Catena*, 6, 259-266.

Heine, K., 1979b: Studien zur Jungpleistozänen Klima- und Landschafts entwicklung der Kalahari, südliches Afrika, *Deutscher Geographentag*, 42, 281-283.

Heine, K., 1980: Wann hat es in der Kalahari geregnet?, *Umschau*, 80, 250-251.

Heine, K., 1981: Aride und pluviale bedingungen während der letzten kaltzeit in der Südwest-Kalahari (südliches Afrika), *Zeitschrift für Geomorphologie*, Supplementband, 38, 1-37.

Heine, K., 1982: The main stages of late Quaternary evolution of the Kalahari region, southern Africa, *Palaeoecology of Africa*, 15, 53-76.

Heine, K., 1990: Some observations concerning the age of the dunes in the western Kalahari and palaeoclimatic implications, *Palaeoecology of Africa*, 21, 161-177.

Helgren, D. M., 1979: *River of Diamonds - an Alluvial History of the Lower Vaal Basin, South Africa*, Department of Geography, Research Paper 185, University of Chicago.

Holmgren, K., 1995: *Late Pleistocene Climatic and Environmental Changes in Central Southern Africa*, Dissertation Series, No. 3, Department of Physical Geography, Stockholm University, Stockholm.

Hutchins, D. G., Hutton, S. M. and Jones, C. R., 1976: The geology of the Okavango Delta, *The Proceedings of the Symposium on the Okavango Delta and its Future Use*, Botswana Society, Gaborone, 15-19.

Hutchinson, I. P. G. and Midgeley, D. C., 1973: A mathematical model to aid management of outflow from the Okavango Swamp, Botswana, *Journal of Hydrology*, 19, 93-112.

Jacobs, N., 1996: The flowing eye: water management in the upper Kuruman Valley, South Africa, *Journal of African History*, 37, 237-260.

Kalahari District Council and McGregor Museum (KDC-MM), undated: *Wonderwerk Cave*, A leaflet for the Wonderwerk Cave display, McGregor Museum, Kimberley.

King, L. C., 1949: On the ages of African land-surfaces, *Quarterly Journal of the Geological Society of London*, 104 (4), 439-453.

King, L. C., 1951a: *South African Scenery*, 2nd Edn., Oliver and Boyd, Edinburgh.

King, L. C., 1951b: The geology of Makapan and other caves, *Transactions of the Royal Society of South Africa*, 33, 121-150.

King, L. C., 1955: Pediplanation and isostasy: an example from South Africa, *Quarterly Journal of the Geological Society of London*, 111, 353-359.

King, L. C. and King, L. A., 1959: A reappraisal of the Natal monocline, *South African Geographical Journal*, 41, 15-30.

Knight, K., Hall, E., Hendry, R. W. and Associates, 1993: *Mmabatho Structure Plan, General Geological and Geotechnical Conditions with Special References to Proposed New Developments*, Stewart Scott Inc., Johannesburg.

Köberl, C., Armstrong, R. A., and Reimold, W. U., 1997: Morokweng, South Africa: A large impact structure of Jurassic-Cretaceous boundary age, *Geology*, 25 (8), 731-734.

Köberl, C., Reimold, W. U., Armstrong, R. A., Brandstätter, F. and Kruger, F. J., 1997: Morokweng, South Africa: the Jurassic-Cretaceous boundary impact structure?, *Abstract, Lunar and Planetary Science Conference XXVIII, March 15-19*, Houston.

Levin, M., 1980: *A Geological and Hydrochemical Investigation of the Uranium Potential of an Area between the Orange and Kuruman Rivers, North-Western Cape Province*, Report PEL-272, Atomic Energy Board, Pelindaba.

Levin, M., Hambleton-Jones, B. E. and Smit, M. C. B., 1985: Uranium in the groundwater of the Kalahari regions south of the Molopo River, in: Hutchins, D. G. and Lynam, P. A. (eds.), *The Proceedings of a Seminar on the Mineral Exploration of the Kalahari, October 1983*, Bulletin 29, Botswana Department of Geological Survey, Gaborone, 234-250.

Lewis, A. D., 1936: Sand dunes of the Kalahari within the borders of the Union, *South African Geographical Journal*, 19, 23-32.

Lister, L. A., 1979: The geomorphic evolution of Zimbabwe-Rhodesia, *Transactions of the Geological Society of South Africa*, 82, 363-370.

Livingstone, D., 1858: *Missionary Travels and Researches in South Africa*, Harper and Brothers, New York, Reprinted 1971, Johnson Reprint Corporation, New York.

Loock, J. C. and Visser, J. N. J., 1985: South Africa, in: Diaz, C. M. (ed.), *The Carboniferous of the World. Volume II, Australia, Indian subcontinent, South Africa, South America and North Africa*, Inst. Geol. Min. Espana, IUGS Publ., 20, pp 167-174.

Ludtke, G., 1985: Mineral exploration in two selected areas of the Kalahari. Preliminary results, in: Hutchins, D. G. and Lynam, A. P. (eds.), *The Proceedings of a Seminar on the Mineral Exploration of the Kalahari, October 1983*, Bulletin 29, Botswana Department of Geological Survey, Lobatse, 81-105.

Main, M., 1987: *Kalahari: Life's Variety in Dune and Delta*, Southern Book Publishers, Johannesburg.

Maizels J., 1990: Raised channel systems as indicators of palaeohydrologic change: a case study from Oman, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 76, 241-277.

Makara, N., 198- *ntribution to the Weather and Meteorological Studies of Bophuthatswa... ...ith Special Emphasis on the Rainfall Distribution and its Applications to Agrometeorological Activities*, Internal Report, Bophuthatswana Department of Transport, Mmabatho.

Mallick, D. I. J., Habgood, F. and Skinner, A. C., 1981: A geological interpretation of LANDSAT imagery and air photography of Botswana, *Overseas Geological and Mineralogical Resources*, 56, Institute of Geological Sciences, London.

Mann, A. W. and Horwitz, R. C., 1979: Groundwater calcrete deposits in Australia: some observations from western Australia, *Journal of the Geological Society of Australia*, 26, 293-303.

Marshall, T. R., 1990: *The Nature, Origin, and Evolution of the Diamondiferous Gravels of the South-Western Transvaal*, Unpublished Ph. D. Thesis, University of the Witwatersrand, Johannesburg.

Martin, H., 1973: The Atlantic margin of southern Africa between latitude 17^0 south and the Cape of Good Hope, in: Nairn, A. E. M. and Stehli, F. G. (eds.), *The Ocean Basins and Margins*, Vol. 1, Plenum, New York.

Martin, H., 1981: The late Palaeozoic Gondwana glaciation, *Geologische Rundschau*, 70, 480-498.

Martin, H. and Wilczewski, N., 1970: Palaeoecology, conditions of deposition and palaeogeography of the marine Dwyka beds of South West Africa, *Proceedings Papers of the Second Gondwana Symposium, South Africa*, 225-232.

Martinelli, E. and Associates, 1981: *Report on Engineering Geological Investigations of the Proposed Township of Mmabatho Unit 8*, Department of Economic Affairs, Geological Survey Branch, Republic of Bophuthatswana, Mmabatho.

Matthews, D. L., 1960: The clay mineralogy of some Mojave Desert Playas, *The Compass*, 38 (1), 28-38.

Mayer, J. J., 1973: Morphotectonic development of the Harts River valley in relation to the Griqualand-Transvaal axis and the Vaal and Molopo Rivers, *Transactions of the Geological Society of South Africa*, 76, 183-194.

Mayer, J. J., 1985: Interpretation of heavy-mineral patterns in sandy soils of the region drained by the upper reaches of the Harts River, *Transactions of the geological society of South Africa*, 88, 27-32.

McCarthy, T. S., 1983: Evidence for the former existence of a major, southerly flowing river in Griqualand West, *Transactions of the Geological Society of South Africa*, 86, 37-49.

McCarthy, T. S., Moon, B. P. and Levin, M., 1985: Geomorphology of the western Bushmanland Plateau, Namaqualand, South Africa, *South African Geographical Journal*, 67, 160-178.

McKinnon, W. B., 1997: Extreme Cratering, *Science*, 276, 1346-1348.

Meigs, P., 1953: World distribution of homoclimates, in: *Reviews of Research on Arid Zone Hydrology*, UNESCO, Paris, 203-209.

Meixner, H., Peart, R. J. and Hutchins, D. G., 1985: Some geophysical and geological results of follow-up drilling of the reconnaissance aeromagnetic survey project, in:

Hutchins, D. G. and Lynam, A. P. (eds.), *The Proceedings of a Seminar on the Mineral Exploration of the Kalahari, October 1983*, Bulletin 29, Botswana Department of Geological Survey, Lobatse, 39-80.

Michaluck, E. and Moen, H. F. G., 1991: *The Geology of the Mafikeng Area*, Geological Survey, Pretoria.

Milliman, J. D. and Mead, R. H., 1983: World-wide delivery of river sediment to the oceans, *Journal of Geology*, 91, 1-21.

Moen, H. F. G., 1977a: Explanatory notes, Sheet 2722 Kuruman, 1: 250 000 Geological Series, Geological Survey, Pretoria.

Moen, H. F. G., 1977b: Explanatory notes, Sheet 2822 Postmasburg, 1: 250 000 Geological Series, Geological Survey, Pretoria.

Moon, B. P. and Dardis, G. F., 1988: Introduction, in: Moon B. P. and Dardis, G. F. (eds.), *The Geomorphology of Southern Africa*, Southern Book Publishers, Johannesburg, 1-11.

Moon, B. P. and Partridge, T C., 1993: Geomorphology in southern Africa, in: Walker, H. J. and Grabau, W. E. (eds.), *The Evolution of Geomorphology*, John Wiley, New York, 383-395.

Nash, D. J., 1992: *The Development and Environmental Significance of the Dry Valleys in the Kalahari, Central Southern Africa*, Unpublished Ph. D. Thesis, University of Sheffield.

Nash, D. J., 1995: Structural control and deep-weathering in the evolution of the dry valley systems of the Kalahari, central southern Kalahari, *Africa Geoscience Review*, 2 (1), 9-23.

Nash, D. J., 1996: On the dry valleys of the Kalahari: documentary evidence of environmental change in central southern Africa, *The Geographical Journal*, 162, 154-168.

Nash, D. J., Thomas, D. S. G., and Shaw, P. A., 1992: Late Quaternary fluvial activity in the dry valleys (mekgacha) of the middle and southern Kalahari, southern Kalahari, *Journal of Quaternary Science*, 7 (4), 273-281.

Nash, D. J., Shaw, P. A. and Thomas, D. S. G., 1994: Duricrust development and valley evolution: process-landform links in the Kalahari, *Earth Surface Processes and Landforms*, 19, 299-317.

Nash, D. J., Thomas, D. S. G., and Shaw, P. A., 1994: Timescales, environmental change and dryland valley development, in: Millington, A. C. and Pye, K. (eds.), *Environmental Change in Drylands: Biogeographical and Geomorphological Perspectives*, John Wiley, New York, 25-41.

Nott, J., 1992: Long-term drainage evolution in the Shoalhaven catchment, south-east highlands, Australia, *Earth Surface Processes and Landforms*, 17, 361-374.

Nugent, C., 1990: The Zambezi River: tectonism, climatic change and drainage evolution, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 78, 55-69.

Nyblade, A. A. and Robinson, S. W., 1994: The African Superswell, *Geophysical Research Letters*, 21, 765-768.

Ollier, C. D., 1995: Tectonics and landscape evolution in south-east Australia, *Geomorphology*, 12, 37-44.

Partridge, T. C., 1993: The evidence for Cainozoic aridification in southern Africa, *Quaternary International*, 17, 105-110.

Partridge, T. C., 1997a: Cainozoic environmental change in southern Africa, with special emphasis on the last 200 000 years, *Progress in Physical Geography*, 21, 3-22.

Partridge, T. C., 1997b: Evolution of landscapes, in: Cowling, R. M., Richardson, D. M. and Pierce, S. M. (eds.), *Vegetation of Southern Africa*, Cambridge University Press, Cambridge, 5-20.

Partridge, T. C., 1997c: Late Neogene uplift in eastern and southern Africa and its palaeoclimatic implications, in: Ruddiman, W. F. (ed.), *Tectonic Uplift and Climate Change*, Plenum Press, New York, 63-86.

Partridge, T. C. and Brink, A. B. A., 1967: Gravels and terraces of the lower Vaal River Basin, *South African Geographical Journal*, 49, 21-38.

Partridge, T. C. and Maud, R. R., 1987: The geomorphic evolution of southern Africa since the Mesozoic, *South African Journal of Geology*, 90, 179-208.

Partridge, T. C. and Maud, R. R., 1988: The geomorphic evolution of southern Africa: a comparative review, in: Dardis, G. F. and Moon, B. P. (eds.), *Geomorphological Studies in Southern Africa*, Balkema, Rotterdam, pp. 5-15.

Partridge, T. C. and Maud, R. R., 1989: The end-Cretaceous event: new evidence from the southern Hemisphere, *South African Journal of Science*, 85, 428-430.

Partridge, T. C., Ward, J. D. and Zavada, M. S.: Late Cretaceous fluvial sediments at Mahura Muthla, Ghaap Plateau, Northern Cape (in preparation).

Passarge, S., 1904: *Die Kalahari*, Dietrich Reimer, Berlin.

Poynton, R. J., 1971: A silvicultural map of southern Africa, *South African Journal of Science*, 67, 58-60.

Pretorius, D. A., 1985: The Kalahari foreland, its marginal troughs and overthrust belts, and the regional structure of Botswana, in: Hutchins, D. G. and Lynam, A. P. (eds.), *The Proceedings of a Seminar on the Mineral Exploration of the Kalahari, October 1983*, Bulletin 29, Botswana Department of Geological Survey, Lobatse, 294-319.

Range, P., 1912: Topography and geology of the German South Kalahari, *Transactions of the Geological Society of South Africa*, 15, 63-73.

Rathbone, P.A. and Gould, D., 1982: *The Geology of the Molopo Farms Project Area*, Report PAR/1/82, Botswana Department of Geological Survey, Lobatse.

Reeves, C. V., 1973: Rifting in the Kalahari?, *Nature*, 237, 95-96.

Reimold, W. U. and Gibson, R. L., 1996: Geology and evolution of the Vredefort Impact Structure, South Africa, *Journal of African Earth Sciences*, 23, 125-162.

Reimold, W. U., Köberl, C., Brandstätter, F. J., Armstrong, R. A., Bootsman, C. S., (in press): The Morokweng impact structure, South Africa: geological, petrological, and isotopic results, and implications for the size of the structure, *Proceedings of the Conference on Large Meteorite Impacts and Planetary Evolution (Sudbury, 1997)*, GSA Special Paper XXX.

Rogers, A. W., 1907: *Geological Survey of Parts of Vryburg, Kuruman, Hay and Gordonia*, Annual report, Geological Commission of the Colony of the Cape of Good Hope, 11-22.

Rogers, A. W., 1922: Post-Cretaceous climates of South Africa, *South African Journal of Science*, 19, 1-31.

Rogers, A. W., 1936: The surface geology of the Kalahari, *Transactions of the Royal Society of South Africa*, 24(1), 57-80.

Samancor, undated: *Overview of the Manganese Mines*, Information booklet of Manganese Ores Division, Samancor, Hotazel.

Schulze, R. E. and McGee, O. S., 1978: Climatic indices and classifications in relation to the biogeography of southern Africa, in: Werger, M. J. A. (eds.), *Biogeography and Ecology of Southern Africa*, Junk, The Hague, 19-54.

Schumm, S. A., 1977: *The Fluvial System*, John Wiley, New York.

Schutte, I. C., 1994: *Die Geologie van die Gebied Christiana*, Geological Survey, Pretoria.

Schwarz, E. H. L., 1919: The progressive desiccation of Africa: the cause and the remedy, *South African Journal of Science*, 15, 139-190.

Schwarz, E. H. L., 1920: *The Kalahari, or Thirstland Redemption*, Maskew Miller, Cape Town.

Schwarz, E. H. L., 1923: The Kalahari scheme as the solution of the South African drought problem, *The South African Journal of Science*, 20, 208-222.

Shaw, P. A., 1989: Fluvial systems of the Kalahari - a review, in: Yair, A. and Berkowicz, S., (eds.), Arid and Semi-arid Environments - Geomorphological and Pedological Aspects, *Catena Supplement*, 14, 119-126.

Shaw, P. A. and Cooke, H. J., 1986: Geomorphic evidence for the late Quaternary palaeoclimates of the middle Kalahari of northern Botswana, *Catena*, 13, 349-359.

Shaw, P.A. and Thomas, D. S. G., 1988: Some observations on drainage evolution and related sediments in the Kalahari of Botswana, *Abstract, 9th Biennial SASQUA Conference*, Durban, 23-24.

Shaw, P. A. and Thomas, D. S. G., 1993: Geomorphology, sedimentation, and tectonics in the Kalahari Rift, *Israel Journal of Earth Sciences*, 41, 87-94.

Shaw, P. A. and de Vries, J. J., 1988: Duricrust, Groundwater and valley development in the Kalahari of south-east Botswana, *Journal of Arid Environments*, 14, 245-254.

Shaw, P. A., Thomas, D. S. G. and Nash, D. J., 1992: Late Quaternary fluvial activity in the dry valleys (mekga 'a) of the middle and southern Kalahari, southern Africa, *Journal of Quaternary Science*, 7 (4), 273-281.

Smit, P. J., 1964: *Diep Boorgate vir Water op Kruispad, Drieling en Potsepan, Gordonia Distrik*. Unpublished report 64-05-02. Geological Survey of South Africa, Kuruman Office.

Smit, P. J., 1974a: Sheet 2522, Bray, geological map, scale 1:250 000, with explanatory notes, Geological Survey of South Africa, Pretoria.

Smit, P. J., 1974b: Sheet 2622, Morokweng, geological map, scale 1:250 000, with explanatory notes, Geological Survey of South Africa, Pretoria.

Smit, P. J., 1977: *Die Geohidrologie in die Opvanggebied van die Moloporivier in die Noordelike Kalahari*, Unpublished Ph. D. Thesis, University of the Orange Free State, Bloemfontein.

Snyman, P. H. R., 1992: *Kuruman - Vervloe Pad na Afrika*, Human Sciences Research Council, Pretoria.

South African Committee for Stratigraphy (SACS), 1980: *Stratigraphy of South Africa*, Geological Survey of South Africa, Handbook 8, Pretoria.

Stigand, A. G., 1912: Notes on Ngamiland, *Geographical Journal*, 376-379.

Stratten, T., 1979: The origin of the diamondiferous alluvial gravels in the south-western Transvaal, in: Anderson, A. M. and Van Biljon, W. J. (eds.), Some sedimentary basins and associated ore deposits in South Africa, *Special Publication, Geological Society of South Africa*, 6, 219-228.

Su, W. -J., Woodward, R. L. and Dziewonski, A. M., 1994: Degree 12 model of shear velocity heterogeneity in the mantle, *Journal of Geophysical Research*, 99, 6945-6980.

Sugden, D. E., 1996: The east-Antarctic ice sheet: unstable ice or unstable ideas?, *Transactions of the Institute of British Geographers*, NS 21, 443-454.

Summerfield, M. A., 1982: Distribution, nature and probable genesis of silcrete in arid and semi-arid southern Africa. in: Yaalon, D. H. (ed.), Aridic soils and Geomorphic Processes, *Catena Supplement*, 1, 37-65.

Summerfield, M. A., 1983: Silcrete as a palaeoclimatic indicator: evidence from southern Africa, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 41, 65-79.

Summerfield, M. A., 1985a: Tectonic background to long-term landform development in tropical Africa, in: Douglas, I. and Spencer, T., (eds.), *Environmental Change and Tropical Geomorphology*, Allen and Unwin, London, 281-294.

Summerfield, M. A., 1985b: Plate tectonics and landscape development on the African continent, in Morisawa, M. and Hack, J. T. (eds.), *Tectonic Geomorphology, The Bingham Symposia in Geomorphology, International Series, 15*, Allen and Unwin, Boston, 27-51.

Summerfield, M. A., 1996: Tectonics, geology, and long-term landscape development, in: Adams, W. M., Goudie, A. S., and Orme, A. R. (eds.), *The Physical Geography of Africa*, Oxford University Press, Oxford.

Swanevelder, C. J., 1974: 'n Morfometriese studie van die Tsauchabvallei van die sentrale Namib, *South African Geographer*, 4, 325-334.

Tankard, A. J., Jackson, K. A., Eriksson, K. A., Hobday, D. K., Hunter, D. R., and Minter, W. E. L., 1982: *Crustal Evolution of Southern Africa*, Springer-Verlag, New York.

ten Brink, U. and Stern, T., 1992: Rift flank uplifts and hinterland basins: comparison of the Transantarctic Mountains with the Great Escarpment of southern Africa, *Journal of Geophysical Research*, 97, 569-585.

Thomas, D. S. G., 1988: The nature and depositional setting of arid and semi-arid Kalahari sediments, southern Africa, *Journal of Arid Environments*, 14, 17-26.

Thomas, D. S. G. and Shaw, P. A., 1988: Late Cainozoic drainage evolution in the Zambezi Basin: geomorphological evidence from the Kalahari rim, *Journal of African Earth Sciences*, 7, 611-618.

Thomas, D. S. G. and Shaw, P. A., 1990: The deposition and development of the Kalahari Group sediments, central southern Africa, *Journal of African Earth Sciences*, 10 (1/2), 187-197.

Thomas, D. S. G. and Shaw, P. A., 1991: *The Kalahari Environment*, Cambridge University Press, Cambridge.

Thomas, D. S. G. and Shaw, P. A., 1992: The Zambezi River: tectonism, climatic change and drainage evolution - is there really evidence for a catastrophic flood? A discussion, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 91, 175-182.

Thomas, D. S. G. and Shaw, P. A., 1993: The evolution and characteristics of the Kalahari, southern Africa, *Journal of Arid Environments*, 25, 97-108.

Thomas, R. J. and Thomas, M. A., 1989: *The Geology of the Noenieput Area*, Geological Survey, Pretoria.

Thomas, R. J., Thomas, M. A. and Malherbe, S. J., 1988: *The Geology of the Nossob and Twee Rivieren Areas*, Geological Survey, Pretoria.

Truswell, J. F., 1977: *The Geological Evolution of South Africa*, Purnell, Cape Town.

Twentyman Jones, V., 1993: *Morphological Variation and its Taxonomic Implications for Insular Populations of Pseudocrenilabrus Philander*, Unpublished M. Sc. Thesis, Rhodes University, Grahamstown.

Twidale, C. R., 1968: *Geomorphology, with Special Reference to Australia*, Nelson, Melbourne.

Twidale, C. R., 1976: On the survival of palaeoforms, *American Journal of Science*, 276, 1138-1176.

Twidale, C. R., 1994: Gondwanan (Late Jurassic and Cretaceous) palaeosurfaces of the Australian craton, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 112, 157-186.

Visser, D. J. L., 1958: *The Geology and Mineral Deposits of the Griquatown Area, Cape Province*, Geological Survey, Pretoria.

Visser, J. N. J., 1981: Carboniferous topography and glaciation in the north-western part of the Karoo Basin, South Africa, *Annals of the Geological Survey of South Africa*, 15 (1), 13-24.

Visser, J. N. J., 1983a: An analysis of the Permo-Carboniferous glaciation in the marine Kalahari Basin, southern Africa, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 44, 295-315.

Visser, J. N. J., 1983b: Submarine debris flow deposits from the Upper Carboniferous Dwyka Tillite Formation in the Kalahari Basin, South Africa, *Sedimentology*, 30, 511-523.

Visser, J. N. J., 1987: The palaeogeography of part of south-western Gondwana during the Permo-Carboniferous glaciation, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 61, 205-219.

Visser, J. N. J., 1989: The Permo-Carboniferous Dwyka Formation of southern Africa: deposition by a predominantly subpolar marine ice sheet, *Palaeogeography, Palaeoclimatology, Palaeoecology*, 70, 377-391.

Visser, J. N. J., no date: Explanatory notes, geological map of the Carbonate rocks in the Northern Cape Province, 1: 250 000, Institute for Groundwater Studies, University of the Orange Free State, Bloemfontein.

Visser, W. A., 1980: *Geological Nomenclature*, Martinus Nijhoff, The Hague.

von M. Harmse, H. J., 1978: Schematic soil map of southern Africa, in: Werger, M. J. A. (ed.), *Biogeography and Ecology of Southern Africa*, Junk, The Hague, 71-75 with map.

Wagner, P. A., 1914: *The Diamond Fields of South Africa*, The Transvaal Leader, Johannesburg.

Weare, P. R. and Yalala, A., 1971: Provisional vegetation map, *Botswana Notes and Records*, 3, 131-152.

Wellington, J. H., 1929: The Vaal-Limpopo watershed, *South African Geographical Journal*, 12, 36-45.

Wellington, J. H., 1941: Stages in the process of river-superimposition in the southern Transvaal, *South African Journal of Science*, 37, 79-96.

Wellington, J. H., 1955: *Southern Africa: A Geographical Study*, (Volume 1: Physical Geography), Cambridge University Press, Cambridge.

Wellington, J. H., 1958: The evolution of the Orange River Basin: some outstanding problems, *South African Geographical Journal*, 40, 3-30.

Werger, M. J. A., 1978: The Karoo-Namib region, in: Werger, M. J. A. (ed.), *Biogeography and Ecology of Southern Africa*, Junk, The Hague, 233-295.

Williams, I. I. E., 1993: *Geological Summary of the Kalahari Formation in the North-Eastern Boundary of Mamatwan Open Pit*, Internal memorandum, Samancor, Manganese Mines, Hotazel.

Williams, M. A. J., Dunkerley, D. L., de Decker, P., Kershaw, A. P. and Stokes, T. J., 1993: *Quaternary Environments*, Edward Arnold, London.

Wray, R. A. L., 1997: A global review of solutional weathering forms on quartz sandstones, *Earth-Science Reviews*, 42, 137-160.

Wright, E. B., 1978: Geological studies in the northern Kalahari, *Geographical Journal*, 144, 235-250.

Author: Bootsman, Cornelis Siebe.
Name of thesis: The evolution of the Molopo drainage - Cornelis Siebe Bootsman.

PUBLISHER:
University of the Witwatersrand, Johannesburg
©2015

LEGALNOTICES:

Copyright Notice: All materials on the University of the Witwatersrand, Johannesburg Library website are protected by South African copyright law and may not be distributed, transmitted, displayed or otherwise published in any format, without the prior written permission of the copyright owner.

Disclaimer and Terms of Use: Provided that you maintain all copyright and other notices contained therein, you may download material (one machine readable copy and one print copy per page)for your personal and/or educational non-commercial use only.

The University of the Witwatersrand, Johannesburg, is not responsible for any errors or omissions and excludes any and all liability for any errors in or omissions from the information on the Library website.