







Southern African GEOMORPHOLOGY

Recent Trends and New Directions

Peter Holmes
Michael Meadows

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Edited by

Peter Holmes

Michael Meadows

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Dedication

This volume is dedicated to Emeritus Professor Margaret Marker, whose passion for southern African geomorphology inspired colleagues and a generation of students, and whose eclectic research interests resulted in numerous papers on southern Africa's geomorphology.

Contents

Foreword

The southern African landscape is an incomparable natural laboratory for the geomorphologist. There is a vast range of landscape types ranging from the periglacial of the high mountain areas to the arid sandy deserts, from landscapes that exhibit the dominance of structural or lithological control to areas where erosional or depositional processes have dominated. There are, also, landforms of a prodigious range of ages from the very ancient planation surface remnants, developed since the fragmentation of Gondwana, to those on recently developed floodplains, wetlands and on the coastal margin.

The landscapes and landforms of southern Africa have fascinated earth scientists for more than a century. Initially those involved were geologists, but progressively geomorphology has become a discipline in its own right and during the last few decades specialist geomorphologists, both from abroad and locally trained, have made the field their own. Signal work was undertaken by the likes of Alex du Toit, John Wellington, Frank Dixey and Lester King who made invaluable world-class contributions to macroscale geomorphology. In the last five to six decades the focus has become more thematic and directed at a smaller scale, and there has been a significant increase in both the interest in different landscapes, and in the number of geomorphologists working in the region.

By the 1980s there was a plethora of introductory text books on geomorphology (largely of British or American origin), but little in the way of a compact account of southern African geomorphology. Students of the discipline had to rely on the excellent (but by then somewhat dated) contributions in John Wellington's (1955) *Southern Africa*, and Lester King's (1963) *South African Scenery*. *The Geomorphology of Southern Africa* (Moon and Dardis, 1988) was produced to fill the gap and to provide a contemporary and informative account of the landscapes around us.

Since the late 1980s geomorphology in southern Africa has advanced dramatically and there has been a marked increase in the tempo of research. Some of the developments are reflected in Tim Partridge and Rodney Maud's (2000) *The Cenozoic of Southern Africa* in which the focus is on the last 65 million years. The time is ripe, however, for a new synthesis. In a volume such as this there is a need to demonstrate the developments in the discipline and to provide a contemporary benchmark reflecting the state of the art. In the pages that follow these goals are admirably achieved. There is clearly improved understanding in virtually every sphere of southern African geomorphology, and this attests to the high quality of the research that has been undertaken in this part of the world. Further, it is evident that there is an increased awareness within the discipline with respect to environmental problems, environmental planning, and the effects of climate change. Geomorphologists are contributing not only to the pursuit of their esoteric science, but also to interdisciplinary efforts in tackling environmental issues. The present volume is evidence of a mature discipline and it will ensure the continuation of the great geomorphological tradition in southern Africa.

Bernard Moon
Balgowan, 2012

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Contributing authors

Charles H Barker	Department of Geography, University of the Free State, Bloemfontein
Greg A Botha	Council for Geoscience, Pietermaritzburg
John Boardman	Environmental Change Institute, University of Oxford
Andrew S Carr	Department of Geography, University of Leicester
Simon J Carr	Queen Mary, University of London
Franck D Eckardt	Department of Environmental and Geographical Science, University of Cape Town
Andrew S Goudie	Saint Catherine's College, University of Oxford
Gerald G Garland	Durban
Stefan W Grab	School of Geography, Anthropology and Environmental Sciences, University of the Witwatersrand, Johannesburg
Kevin J Hall	Department of Geography, University of Northern British Columbia, Prince George
M Timm Hoffman	Plant Conservation Unit, University of Cape Town
Peter J Holmes	Department of Geography, University of the Free State, Bloemfontein
K Ian Meiklejohn	Department of Geography, Rhodes University, Grahamstown
Michael E Meadows	Department of Environmental and Geographical Science, University of Cape Town
Rodney R Maud	Drennan Maud and Partners, Durban
Stephanie C Mills	Kingston University, London
David J Nash	Department of Geography, University of Brighton
Werner Nel	Department of Geography, University of Fort Hare, Alice
Kate M Rowntree	Department of Geography, Rhodes University, Grahamstown
Paul A Shaw	University of the West Indies, Saint Augustine, Trinidad and Tobago
Paul D Sumner	Department of Geography, Geoinformatics and Meteorology, University of Pretoria
C Rowl Twidale	School of Earth and Environmental Sciences, University of Adelaide
David SG Thomas	School of Geography, University of Oxford
Giles FS Wiggs	School of Geography, University of Oxford

Preface

Southern African Geomorphology: Recent Trends and New Directions

Southern African Geomorphology: Recent Trends and New Directions

Peter J Holmes and Michael E Meadows

It is now twenty four years since the publication of *The Geomorphology of Southern Africa*, edited by Bernie Moon and George Dardis. That volume was the first and, until now, the only text to attempt an up-to-date synthesis of southern African geomorphology. In the words of the editors, it sought "...to bridge the gap between general geomorphological theory and the landscapes of southern Africa" (Moon and Dardis, 1988: preface). Although regional geomorphologies of South Africa have subsequently been published (see Lewis, 1996; Lewis, 2008), this volume presents the first comprehensive synthesis of the geomorphology of southern Africa to appear since Moon and Dardis (1988).

What motivated this new text? Examination of the literature suggests that, since the final decade of the previous century, there has been an exponential increase in published research investigating form and, more significantly, process relating to southern African landforms. We identify four factors that may have contributed to this. Firstly, there has been a notably increased focus on the physical environment of the sub-continent, not only from the perspective of the purely academic, but also in terms of growing public environmental awareness. The massive growth in public concern over climate change has certainly been an important element of this and has been expressed in research dealing with the association between landforms and environmental change. Secondly, the past two decades have also been characterised by a phenomenal increase in collaborative research between southern African (in particular South African) geomorphologists and colleagues from international institutions, particularly in the United Kingdom, but also in Europe and America. The *internationalisation* of geomorphic research in southern Africa has been facilitated by the political developments of the early nineties that saw the end of the academic boycott, greater interest in conferences hosted within the region and the subsequent development of diverse collaborative relationships between active researchers here, and those from further afield. This was further facilitated through international funding sources coming on stream, at least for some projects. To some extent, these emergent research partnerships are reflected in the authorship of chapters in this book. Thirdly, and partly enabled by the institutional and disciplinary diversity of the collaborations, there has been a substantial escalation in truly interdisciplinary research whereby geomorphologists team with scientists with disciplinary backgrounds from, for example botany and palaeobotany, geology, geochemistry, sedimentology, geomatics, and archaeology to produce interdisciplinary studies. Fourthly, there has been a virtual revolution in the range and sophistication of research tools and methodologies and data generating capacity available to geomorphologists. These are especially prominent in the form of new, sophisticated dating methodologies such as optically stimulated luminescence and cosmogenic isotope dating, but also in other laboratory techniques which, increasingly, utilise organic material from geomorphic inventories for interpreting depositional environments. There have also been huge advances globally in data capture and processing and, perhaps most importantly, in data availability and in the use of related technologies including GIS and satellite imagery in geomorphology.

In this volume we have attempted to present a thematic and systematic overview of recent trends and new directions within the discipline of geomorphology in southern Africa. With the possible exception of one or two chapters where inventories of different landforms are necessary to develop an appreciation of related landscapes, this is, definitively, *not* a presentation of descriptions of the diverse range of southern African landforms and landscapes. Intriguingly, perhaps, that would be an interesting challenge for a somewhat different text altogether. Instead, authors here were requested to focus on recent developments in the way we investigate, comprehend, and account for the range of different landscapes. The result, we hope, is a text that places emphasis on current (and emerging) research directions in southern African geomorphology rather than on generalised generic geomorphic processes and landforms. Inevitably, there are gaps – indeed some may argue flaws – in the scope of the book. The content is, to some extent, constrained by what might be called author capacity or critical mass. For example, from a landscape (rather than, say, an engineering geology perspective) little or no systematic academic work has been undertaken on karst geomorphology in southern Africa over the last two decades. For this reason, as well as space constraints – and so as not simply to re-hash outdated material – we have not included a chapter on karst. For similar reasons, the chapter on geologically controlled landscapes and slopes provides but a fleeting overview; we feel that the current thrust of geomorphic research in southern Africa does not warrant the inclusion of separate chapters on these topics in the context of a volume which defines itself in terms of recent trends. This is not to say that aspects of these topics are not included within the thematic chapters; indeed, they are (see for example chapters on Applied Geomorphology and Soil Erosion and Land Degradation). Some may search in vain, then, for the *equivalents* of chapters in Moon and Dardis (1988) but, equally, there are new inclusions that reflect current research directions. For example, the notion of landscape inventories, incorporating remote sensing imagery and GIS has seen significant advancements in the way we perceive and appreciate landforms, as well as the ways in which we monitor landscape evolution and change. This, then, is reflected here in the form of a dedicated chapter.

It was not possible to invite chapters from every expert from within the southern African geomorphic community, nor indeed from all those researchers beyond the sub-continent who have contributed so significantly to academic geomorphology within southern Africa. We acknowledge that the argument is weaker without such inclusions. In appropriate cases, we have invited these experts to review and comment on the submitted chapters and, elsewhere in the volume; we express our appreciation for these most valuable inputs. It would also be fair to say that the work reflects predominantly Anglophone contributions to the understanding of southern African landscapes. While the important work of German colleagues in, for example, Namibia is both recognised and referenced, the book does not contain direct individual contributions from these researchers. We also wish to point out, with some measure of concern, that there are only two female authors, and not a single black contributor. We do not wish to delve into the possible reasons for this, and this is probably not the appropriate place to introduce a debate on the issue of representivity, other than to state the obvious; in the southern African geomorphic research arena, most senior academics are (still) white males.

We must also acknowledge that, in most cases, the chapters illustrate a bias towards research conducted in South Africa in particular. Indeed, this reflects the predominance of South African-based authors among the contributors. At least in part this situation has been given impetus through the strongly developed sense of collegiality between the South African and international academic community and is reflected in the membership of the Southern African Association of Geomorphologists. It strikes us that geomorphic research in this part of the world is greatly strengthened by an atmosphere of openness, transparency and willingness to share ideas and data: the Southern African Association of Geomorphologists' biennial meetings have proved pivotal in that sense.

In conclusion, while this book attempts to be a worthy and valuable successor to Moon and Dardis (1988), the approach in practice is somewhat different. The latter included a degree of geomorphic theory, as well as explanations of process. This text provides, in the form of its fourteen chapters, synopses of recent trends and research directions as currently reflected in the literature and the on-going research into southern African landforms and landscapes. We trust that the text reflects the vibrancy and vigour (and rigour) of current geomorphic research agendas in southern Africa, and that it will be of value to an informed readership with an interest in southern African geomorphology. We earnestly hope too that it will encourage a new generation of geomorphologists to follow in the footsteps of those who can, with some degree of satisfaction, look back at what has been achieved thus far.

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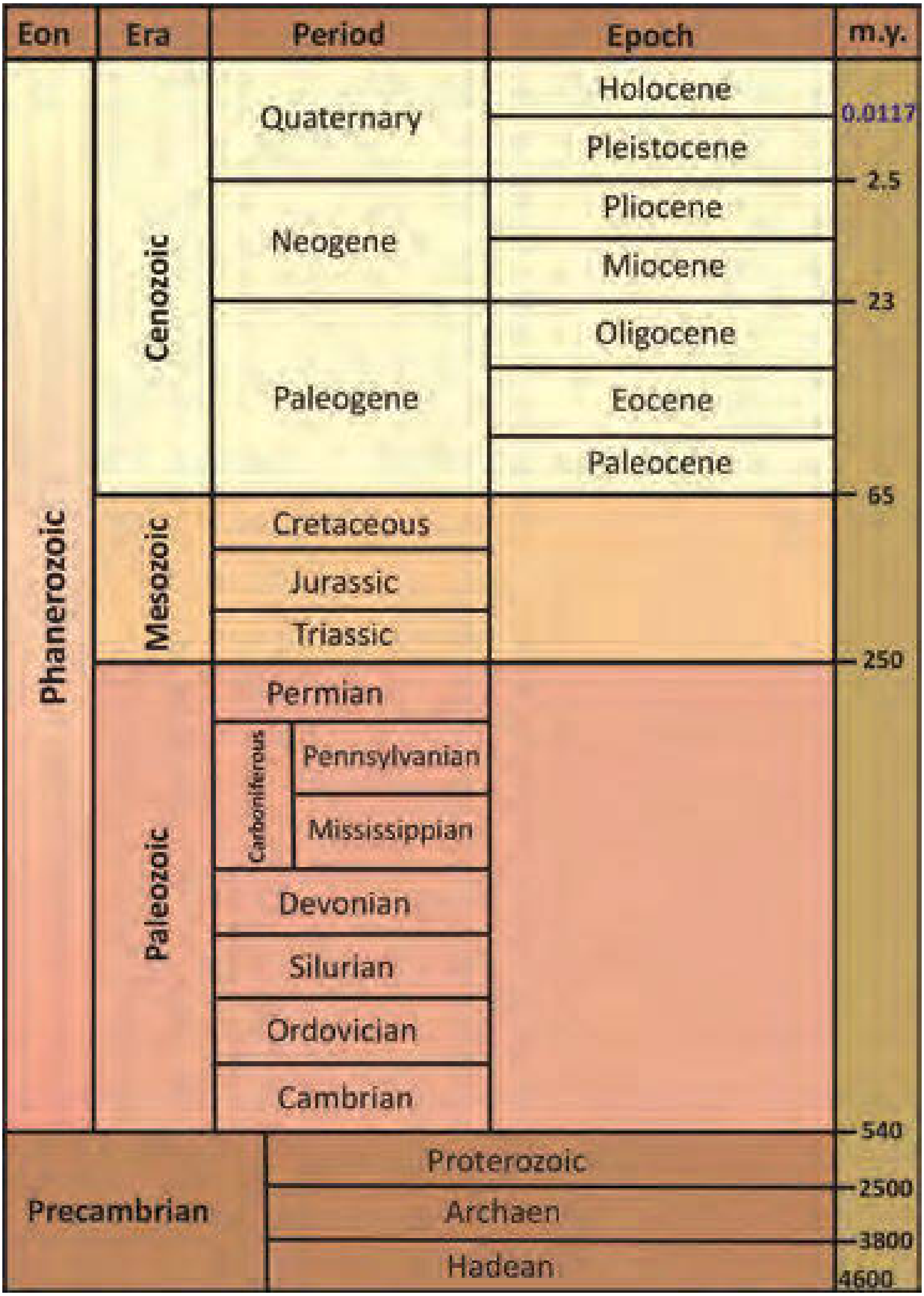
Geological Time Scale



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Macroscale Geomorphic Evolution

Macroscale Geomorphic Evolution

Rodney R Maud

This chapter is dedicated to the memory of my very good long-time friend and colleague Tim Partridge (1942-2009). The chapter is based in large part, on our jointly-authored chapter in Partridge and Maud, 2000.

1. Pre-Cenozoic geology and geomorphology

1.1 Geology

Some of the geomorphic features on the face of southern Africa are the legacy of its long and turbulent geomorphic history that extends back to the Archean. The most extensive of these geomorphic features relate to the time that postdates the breakup of Gondwana, which took place about 145 million years ago on the east coast of southern Africa and approximately 125 million years ago on its west coast. As a result of this breakup, the African continent, as it is known today, came into being.

A summary of the major stages in the geological development of southern Africa is presented in Table 1.1 (modified after Tankard *et al.*, 1982; Moon and Dardis, 1988). This table also shows the times of the formation of the major geological units in their southern African lithostratigraphic sequence. Some knowledge of these units and their timeframes is essential in the understanding of the geomorphology of southern Africa.

The geological evolution of southern Africa has taken place by means of a series of accretionary events that followed the stabilisation of the granitic Kaapvaal Craton (which forms the structural basement of the subcontinent) at approximately 2 600 Ma. Accretion occurred during a number of extensional and compressional periods, the most important of which, during the period approximately 2 000 to 1 000 Ma, gave rise to the addition of the Namaqua-Natal mobile belt on the southern side of the Kaapvaal craton, it culminating in the stabilisation of the Kalahari Craton, to its northwest, around 1 000 Ma. On to this amalgam was imposed a series of orogenic belts, creating *swells*, between intercratonic *basins* in the course of the Pan-African tectonic cycle which ended about 600 Ma ago. As Burke (1996) has pointed out, this basin-and-swell structure is unique to the African continent. The swells were repeatedly rejuvenated and uplifted during the Phanerozoic in relation to the thick and mechanically strong intervening cratons. These recurrent movements persisted into the Neogene (post 23 Ma) when uplift spread from the inter-cratonic mobile welts to affect large areas of the cratons themselves. (For more detail on this subject, see Figure 1.1 in Partridge and Maud, 2000).

In the early Paleozoic (after approximately 500 Ma) a passive margin developed along the southern edge of the Kalahari Craton into which offshore situation the shelf sediments of the Cape Supergroup were deposited. An active margin then developed to the south, giving rise to the Cape Fold Mountains around the end of the Permian (approximately 250 Ma). As a result, the Karoo Basin between the Cape Mountains and the Kalahari Craton was formed which extended well beyond the present margins of the subcontinent into adjoining areas of Gondwana. Sedimentation thickness in the Karoo Basin totalled up to 7 000 m in places. The sedimentation in the Karoo Basin culminated in extensive outpourings of basalt (with some rhyolite-dacite in the northeast) approximately 180 Ma ago as incipient rifting began to rough out the later southern sub-continents and India. When Karoo volcanism ceased, most of southern Africa south of 15°S was covered by Karoo strata, including the volcanics. Somewhat later, at approximately 130 Ma, the Etendeka basaltic lavas were extruded in the west that was associated with the developing Atlantic rift margin in Namibia, see Serra Geral basal of the Parana Basin in Brazil (Duncan and Marsh, 2006).

Table 1.1. Summary of the macroscale geomorphic evolution of southern Africa since the Gondwana break-up (after Moon and Dardis, 1988).

GEOLOGICAL EON	AGE (Ma)	GEOLOGICAL ERA (SACS)*	EVOLUTIONARY STAGE	GEOLOGICAL UNITS
PHANEROZOIC			POST GONDWANA	Cretaceous-Holocene sediments
			GONDWANA	Karoo Volcanics Karoo Supergroup Cape Mountain folding Cape Supergroup
LATE PROTEROZOIC	500	NAMIBIAN (N)	PROTEROZOIC OROGENIC MOBILE BELT ACTIVITY	Gariep Supergroup
MIDDLE PROTEROZOIC	1000	NAMAQUAN (JN)		Namaqualand-Natal metamorphic rocks (Metamorphism and intrusion of younger granite rocks)
EARLY PROTEROZOIC	1500	KHEISIAN (K)		
	2000	VAALIAN (V)	EARLY PROTEROZOIC SUPRACRUSTAL CRATON DEVELOPMENT	Transvaal and Olifantshoek Supergroups Waterberg Group Bushveld Complex
LATE ARCHEAN	2500	RANDIAN (R)		Ventersdorp Supergroup Witwatersrand and Pongola Supergroups
MIDDLE ARCHEAN	3000	SWAZIAN (Z)	ARCHEAN CRUSTAL CRATON DEVELOPMENT	Barberton Supergroup (Development of granite crust)
EARLY ARCHEAN	3500			
	4000			

* South African Commission for Stratigraphy

1.2 Geomorphology

The faulted rifts along which Gondwana breakup occurred were almost certainly associated with pre-existing Pan-African welts (Partridge, 1998). As in the case of the existing much younger, rift system of East Africa, uplift of the flanks of the incipient rifts preceded separation. The remains of these elevated rift shoulders are preserved in the cordon of high ground, which is still present, especially inland of the Great Escarpment of Lesotho and the Eastern Cape Province, South Africa. On the morphological evidence of intrusive Kimberlite pipes, relatively small thicknesses of material have been removed from these elevated areas by Cretaceous and Cenozoic erosion since their emplacement approximately 90 Ma ago (Hawthorne, 1975). On the evidence of the late occurrence of Jurassic marine sediments in Tanzania and Mozambique, rifting on the northeastern coast commenced prior to approximately 140 Ma, it having extended south to the KwaZulu-Natal Province, South Africa coast by 130 Ma, while on the west coast, tensional rift faulting and separation took place somewhat later between 129 and 121 Ma. Along the southern Cape coast the lowermost marine rocks indicate that continental separation there had also occurred by approximately 130 Ma (Watkeys, 2006). Along the east coast, rifting took the form of tensional tilted block faulting, tilting being mainly in a seaward direction. Further south, rifting was of a more transcurrent nature in the Agulhas-Falkland Fracture Zone.

As a result of the disruption of Gondwana, the present oceans were admitted to what had thereby become margins of southern Africa. As a result of the erosion by rivers working headward from the new ocean coastlines, thick sequences of marine Cretaceous sediments were deposited offshore on the continental shelf, and in faulted structural basins offshore. Terrestrial deposits of this age are also present in such basins onshore, particularly in the faulted basins of the southern coast of the Cape, and in the intermontane valleys of Cape Fold Mountains, in places. The marine Cretaceous sedimentary piles are of considerable thickness (in excess of 8 km in places), which supports the findings of fission track analyses onshore in respect of erosion amounts (Brown, Summerfield and Gleadow, 1994) that by the mid-Cretaceous, between one and three kilometres had been removed from the post-rifting surface of the subcontinent.

Recent work by Hanson *et al*. (2009) on the basis of Kimberlite pipe morphology and the xenoliths therein of country rock intruded by the diatreme pipes at various levels in the North West, Northern Cape and Free State Provinces, South Africa in the central portion of the sub-continent, has shown that nearly 1 000 m of Drakensberg basalt was removed between the surface at 180 Ma, and the erosion surface of mid-Cretaceous age at 120 Ma. A similar amount of erosion took place between that time and the end of the Cretaceous at 65 Ma. From the end of the Cretaceous to the present time, a further approximately 500 m of erosion has occurred. All this erosion was achieved by eastward scarp retreat across this region. As a result of this erosion, a great of the covering of Karoo rocks over the central elevated portion of the subcontinent has been stripped from all but the main Karoo Basin, with the imposition of some structural control by formerly underlying older strata on erosion occurring as a result.

All of this evidence points to the fact that prior to the breakup of Gondwana, southern Africa stood high within it, with pre-rifting elevations probably ranging from approximately 2 400 m in Lesotho to approximately 1 500 m in the western interior (Partridge and Maud, 1987). Hanson *et al*. (2009), however, propose that the pre-rift surface at 180 Ma, across the eastern half of the subcontinent at least, stood at an elevation somewhat in excess of 3 000 m.

These high elevations were due in part, to uplift along the rift shoulders and were responsible for the development, just after the times of continental separation, of a substantial marginal escarpment. This was the forerunner of today's Great Escarpment, which occurs as a rampart located about 50 to 200 km inland of the coastline around virtually the entire southern African marginal hinterland. This initial escarpment, formed by rifting, was rapidly driven back by erosion during the early Cretaceous, partly because of the elevation of the interior, and partly as a result of the humid tropical climate that prevailed at that time. This favoured weathering and active fluvial erosion by a dense, well integrated, drainage net (Partridge, 1998).

As the Great Escarpment receded from the coast, erosion to the oceanic base level cut a gently sloping bench-across the coastal hinterland. This erosion was followed closely by two marine transgressions, on the resulting marine-cut bench of which, strata of Upper Cretaceous to Eocene age were deposited in places (see Figures 1.1 and 1.2). On the evidence of the size-range of the definitive amygdaloidal basalt clasts derived from the basalt capping of parts of the Great Escarpment, in near-shore sediments of Early and Late Cretaceous age on the east coast, Matthews and Maud (1988), found that a position within approximately 20 km of the position of the Great Escarpment today, had been reached by the scarp face around the end of the Cretaceous. Using apatite fission track thermochronology, Brown *et al.* (2002) came to the same conclusion.

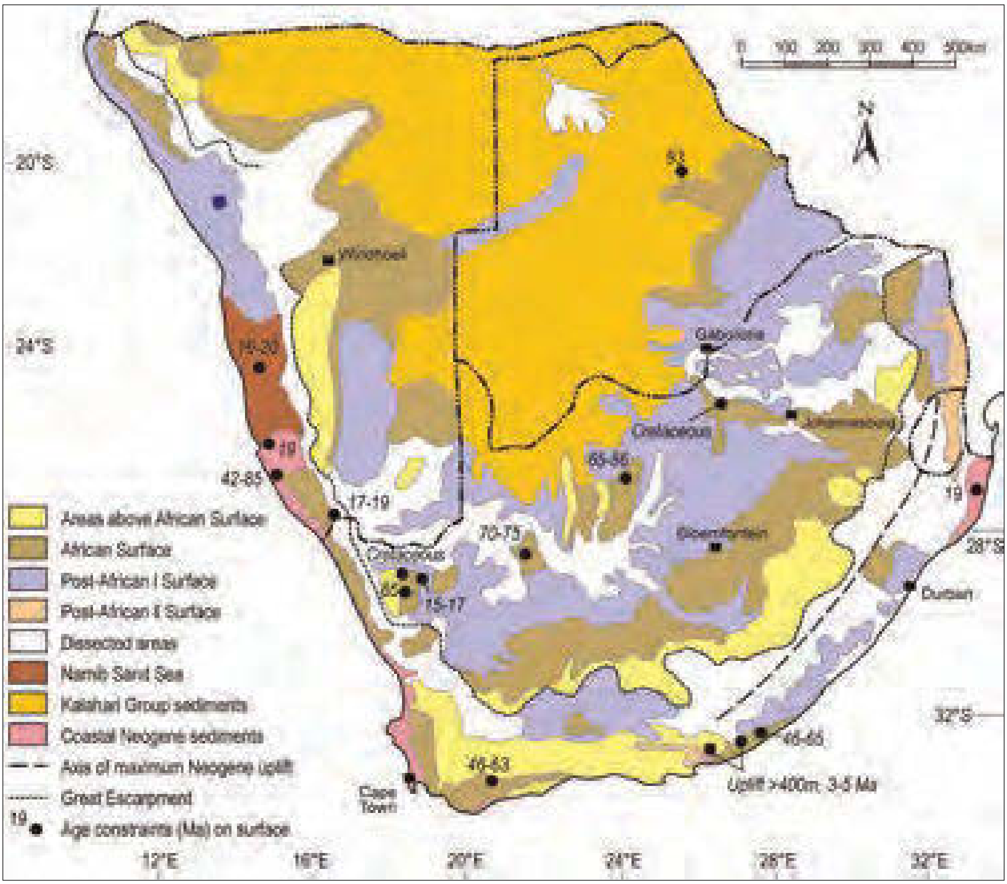

Figure 1.1. Simplified map of the erosion surfaces in southern Africa, showing localities where age constraints on these surfaces are available (after Partridge, 1998).

At the same time erosion was proceeding on the elevated plateau area inland of the escarpment line. Here the base level for erosion was provided by limited number of major river systems at their points of exit from the plateau via a series of waterfalls and rapids. This created the unusual situation that land erosion surfaces of essentially the same age were cut at different levels above and below the Great Escarpment (Partridge and Maud, 1987). As in the case of the coastal platform, geochronological evidence (here in the form of Kimberlite diatreme facies of known age (Figure 1.1) indicates that this inland planation surface was formed no later than the end of the Cretaceous (Partridge and Maud, 1989; Partridge, 1998; Hanson *et al.*, 2009). The outcome of this cycle of erosion and sediment removal from the continental interior was the creation of two vast erosion surfaces, both above and below the Great Escarpment, which have been grouped together as the *African Surface* by L.C. King in a number of seminal papers, for details of which, see Partridge and Maud (1987; 2000).*

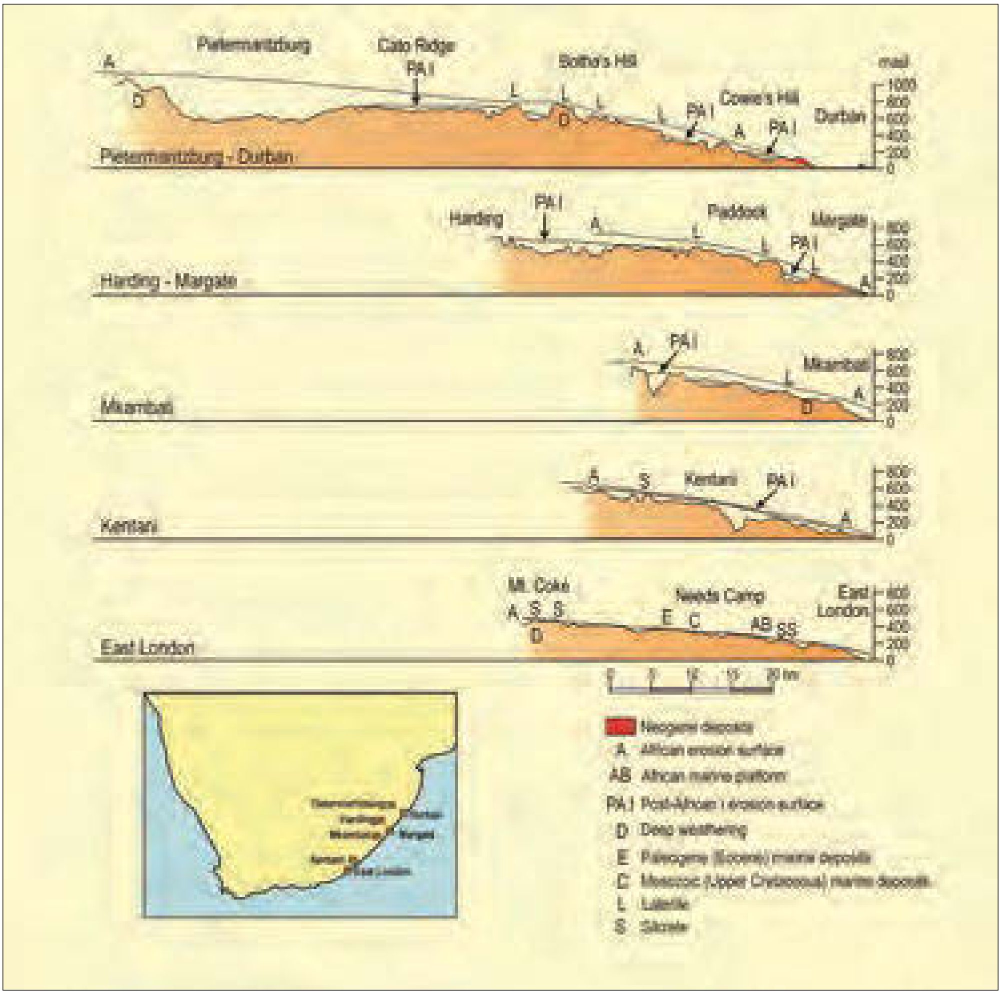

Figure 1.2. Sections drawn inland from the coast across the south-eastern hinterland of southern Africa, showing erosional remnants, duricrusts associated with deep weathering and marine deposits of Late Mesozoic and Early Cenozoic age (after Partridge and Maud, 1987).

Above this African Surface of probable multicyclic origin, a number of mountain massifs was preserved, including the ranges of the Cape Fold Mountains, the Namaqualand Highlands, the mountains behind the Great Escarpment in the Eastern Cape, Lesotho, eastern Mpumalanga and Limpopo Provinces, South Africa and eastern Zimbabwe, as well as the high ranges inland of the Namibian and Angolan escarpments. For a recent review of the geomorphology of Zimbabwe, including the occurrence of the African Surface and younger erosion surfaces there, see Moore *et al*. (2009).

Despite the controversy which has, at times, attended King's interpretations of landscape development, especially on questions of whether the African Surface should be regarded as a peneplain, a pediplain, or an etchplain, (for a review see Twidale, 1988), and whether it can be regarded as a single geomorphic entity, there is no longer any doubt concerning its existence and antiquity. Details of the criteria on which remnants of it have been identified and correlated, through the medium of sections surveyed in the field, are given in Partridge and Maud (1987), from which typical sections on the east coast of the subcontinent are reproduced here as Figure 1.2. The overall results are summarised together with bracketing age data, in map illustration form, in Figure 1.1.

The drainage system of the subcontinent that developed in the Cretaceous had a marked influence on the drainage pattern that has continued subsequently, although some notable changes have occurred, primarily as a result of river capture effects on some of the major rivers such as the lower Orange (or Gariep). For details of these drainage patterns and their evolution with time, the influence of axes of flexuring, the influence of exhumed pre-Karoo features, along with palaeorivers, which no longer exist, in the central and western portions of the country, see De Wit *et al*. (2000). Astonishing is the survival of some of the former ancient river channels on the present landscape from the Late Cretaceous until the present, (a period of approximately 70 Ma) a matter that is of significant economic importance because of the association of alluvial diamonds with many of these channels in this region. One of the more notable of these, on account of it having been able to have been reliably dated on account of the macro-plant fossils (silicified wood) preserved therein, is the locality of Mahura Muthla, near Vryburg, which is of Late Cretaceous age (Partridge, 1998; De Wit *et al*., 2009).

* (Also included in the Partridge and Maud 1987 publication is a review of earlier work on this subject both in southern Africa, and farther north, by King, as well as a number of other notable pioneers in the study of African landscapes, such as Dixey, some of whose findings anticipated those described here).

2. Cenozoic geomorphology

2.1 Cretaceous and Palaeocene landscapes

Apart from the Gondwana breakup events which initiated the development of the current geomorphology of southern Africa, primarily with the development of the river drainage pattern and the erosionally derived African Surface, much of the geomorphology of the subcontinent relates to epeirogenic and climatic events that have taken place in Cenozoic time, that is, since the end of the Cretaceous about 65 Ma ago.

The beginning of the Cenozoic saw southern Africa transected by a widespread planation surface – the African – cut at two levels above and below the Great Escarpment and overlooked in some places by higher-standing mountain massifs. There is strong evidence that climates during much of the Cretaceous had been warm and humid, which not only facilitated rapid erosion through the medium of a well-developed drainage network, but allowed deep weathering mantles to develop over susceptible lithologies (Partridge and Maud, 1987). As planation advanced and erosion rates declined during the latter part of the Cretaceous, so did the African Surface with its deeply weathered saprolites come to occupy much of the landscape. Above the Great Escarpment, the *Highveld* plains of the northeastern Free State, Mpumalanga, and the North West now remain the most extensively preserved of these areas. In the coastal hinterland of the Swartland, inland of the Atlantic coast of the Western Cape Province, South Africa, and the platform extending from Caledon to Mossel Bay along its southern coast, the survival of the African Surface is attested to by the presence of innumerable silcrete-capped remnants of it. Similarly armoured benches are extensively preserved in some of the major valleys of the Cape Fold Mountains, particularly on the lower flanks, as around Uniondale, and also around Grahamstown in the Eastern Cape. Although less humid climatic conditions may have begun to affect some areas of the western interior before the end of the Mesozoic, deep weathered kaolinised saprolite profiles up to 50 m in thickness were evidently of widespread occurrence beneath the African Surface at the time of the *catastrophic* events which brought the Cretaceous to a close (Partridge and Maud, 1989).

The major global change in climates and mass extinctions which accompanied these events have been attributed by most researchers to the global effects of a major meteorite impact, but the evidence for such an event at exactly the relevant time, is becoming increasingly less secure (Keller, 2003). The additional and more prolonged, influence of massive volcanism in the Deccan region of India at this time on atmospheric chemistry and climate change should not be overlooked. The development of widespread climatic desiccation at this time was accompanied by the widespread development of duricrust armouring over the deep weathered profiles on the African Surface. Consisting of silcrete and calcrete in the west, where the response towards greater aridity was apparently the greatest, and laterite in the east, these erosion-resistant duricrust pedocrete cappings have protected remnants of the African Surface from subsequent erosional lowering in many places. Their survival no doubt owes much to widespread aridity in the western and central parts of the subcontinent during the latter part of the Cenozoic.

The nature and genesis of the Late Cretaceous and Early Paleocene pedocretes of southern Africa have been described in detail by Botha (2000), while the occurrence and the nature of silcrete in the Cape coastal belt has more recently been described by Roberts (2003). These African Surface-related pedocretes differ fundamentally in thickness and in distribution from their later counterparts, which have developed on the younger post-African erosion surfaces, usually in response to specific, local topographically-controlled moisture regimes within the soil. Pedocrete-capped remnants of the African Surface can be correlated altimetrically over very wide areas (Partridge and Maud, 1987), and they provide an important datum against which subsequent erosional and tectonic events can be measured. Their mesa-like form and diagnostic armourings of silcrete or laterite, underlain by deeply weathered kaolinised saprolite are unmistakable in the field. Silcretes extend along the coast and its hinterland from southern Namibia in the west, along the southern Cape coast and in its intermontane valleys, to the Eastern Cape, beyond which in the KwaZulu-Natal coastal hinterland it is replaced on the African Surface by laterite.

The close similarities between the laterite pedocrete occurrences on the interior plateau and in the coastal hinterlands attest to their common origin in the geomorphic and climatic events that marked the close of the Mesozoic era. Although not abundant, palaeontological and radiometric ages (see Figure 1.1) are sufficiently consistent and widely distributed to confirm that planation within the African erosional cycle was essentially complete before the end of the Cretaceous after operating for a period of some 70 Ma, and that massive duricrust formation had ended by the Early Paleocene, at least in the west of the subcontinent.

As indicated above, there is evidence for a number of tectonic-disturbances during the long course of the African cycle of erosion, not least in the occurrence of a number of hiatuses and overlaps in the Cretaceous marine sedimentary record offshore, as off the KwaZulu-Natal coast. There is, likewise, no reason to suppose that southern Africa remained unaffected by early events that have led to a global increase in continental hypsometry since the Eocene (Partridge *et al.,* 1995). What is clear, however, is that uplift on a regional scale did not occur in southern Africa during the Paleocene. Offshore sedimentation rates during the Paleocene were variable, but generally slow, this being indicative of low rates of erosion onshore at this time. This slow rate of offshore sedimentation does not support the proposition of an inward-migrating flexural bulge of up to 600 m amplitude, which Gilchrist and

Summerfield (1990) envisage as having formed in response to marine sedimentary loading of the continental shelf by the products of Cretaceous and Early Cenozoic erosion onshore.

The absence of regional tectonics and associated terrestrial deposits (except possibly for the lower units of the Kalahari Group) meant that the African landscape, inherited from the Cretaceous, underwent little modification during the Paleocene. The presence of up to 60 m of cross-bedded aeolian sands in the Eocene Upper Buntfeldschuh Formation of coastal southern Namibia indicates that, at least in the west, the earliest part of this time interval was dry (Ward and Corbett, 1990). This aridity would also undoubtedly have contributed to the slow rate of erosion and landscape development at this time.

3. The Miocene disturbances and uplift

The African cycle of erosion was brought to an end by modest uplift of the sub-continent after the beginning of the Miocene, the maximum uplift of some 250 to 300 m being concentrated in its eastern part. This uplift was of an epeirogenic, non-faulting and non-volcanic nature. At the time of its termination approximately 18 Ma ago by sub-continental uplift, the African cycle of erosion had been operative for a time span of some 120 Ma since its inception at the breakup of Gondwana, although it had been subject to a number of tectonic disturbances during this time. An absence of materials suitable for radiometric dating has necessitated that reliance in respect of the timing of the onset of this end-Early Miocene uplift has to be placed on palaeontological and geomorphic indicators.

The best evidence for Early Miocene uplift comes from the western areas of southern Africa where high terraces of the major rivers such as the Orange (Gariep), and the now abandoned Koa Valley, indicate that the sluggish drainages here during the early Cenozoic responded to the uplift by incising their channels, by headward erosion to the extent of some 100 to 200 m below then existing African Surface pediplain. This incision initiated a new cycle of landscape development, producing a less well-planed surface than that of the African, studded with residual koppies. King and King (1959) called this younger erosion surface the Post-African I Surface, a nomenclature followed by Partridge and Maud (1987), and subsequently. Age control comes from fossils preserved in associated fluvial deposits. Among these are terraces of the proto-Orange (Gariep) River some 40 to 50 m above its present channel which have yielded fauna dating to between 19 and 17 Ma (Corvinus and Hendey, 1978; Pickford *et al*., 1996). Van Niekerk *et al*. (1999a) have provided the first absolute radiometric dates for the Post-African I Surface in the area between Johannesburg and Lichtenburg in the Gauteng Province, South Africa and in the North West. Their $^{40}Ar/^{39}Ar$ ages of 12 to 15 Ma define the time of formation of potassium-bearing cryptomelane in a pedogenic manganese crust immediately below this surface which, therefore, post-dates the surface itself by a small margin. Van Niekerk *et al*. (1999b) have also investigated the relationship between the African Surface and younger surfaces, ancient soil profiles, and palaeoclimatic change in the same area. All this evidence provides confirmation of the earlier proposal by Partridge and Maud (1987) that the Post-African I Surface was formed in response to Early Miocene uplift of the subcontinent.

The epeirogenic uplift of the sub-continent which initiated this new Post-African I erosion cycle was asymmetrical across its width, it reaching a maximum of some 250 to 300 m in KwaZulu-Natal in the southeastern hinterland (Figure 1.3), this amount of uplift being based on the amount of separation of the remnants here of the African and Post-African I erosion surfaces (Figure 1.2). The uplift of the subcontinent gave rise to its seaward marginal tilting, on the east coast the axis of this tilting being located parallel to the coastline, from about Port Elizabeth in the south to Swaziland in the north, and approximately 80 km inland from it. On the Cape south coast, the corresponding uplift was approximately 200 m, while on the west coast it was more or less 150 m (Figure 1.3). As river incision due to this and the later greater uplift proceeded through Neogene time, drainage patterns became increasingly controlled by exhumed pre-Karoo topography that became exposed by Cretaceous erosion of the Karoo rock cover.

At much the same time as the subcontinent (and, indeed, much of the eastern seaboard of Africa) was subjected to this modest uplift, additional localised flexural movements along the Griqualand-Transvaal Axis and the Kalahari-Rhodesia Axis farther to the north (Figure 1.3) accentuated the southern rim of the Kalahari Basin and beheaded southward-flowing drainages, causing the continental divide to shift southwards to the southern edge of the Bushveld Basin. As a result, reorganisation within the relevant drainage networks at this time was considerable.

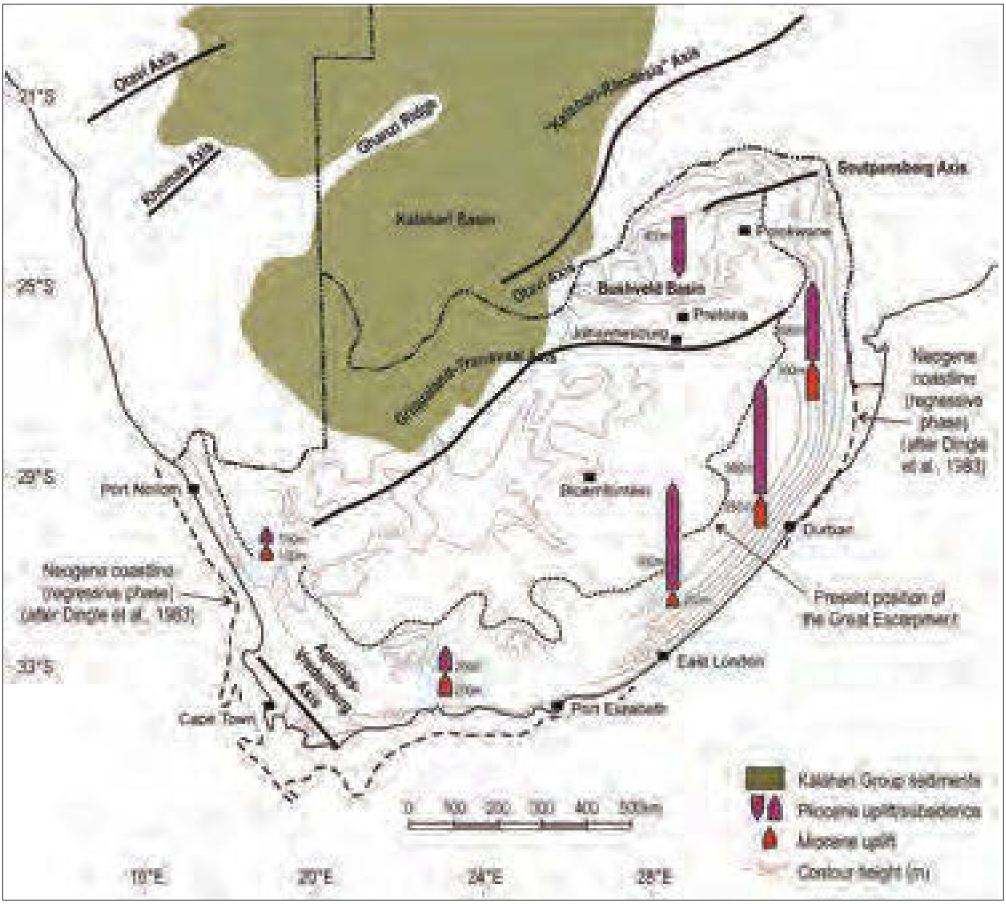

Figure 1.3. Generalised elevation contours in metres and on the Post-African I (Early Miocene) erosion surface (after Partridge and Maud, 1987).

Relative wetness that characterised western (and probably also eastern) southern Africa during the Early Miocene was short-lived, and by about 15 Ma aridity had returned to this region (De Wit, 2000; Marshall and Partridge, 2000; Pickford, 1996). This desiccation which caused the onset of true desert conditions in the Namib, and the aeolian conditions which led to the deposition of the Tsondab Sandstone there at this time, was, in part, associated with the initiation of cold upwelling within the Benguela Current system, which accompanied the establishment of the East Antarctic ice-sheet about

14 Ma. Except during a humid interval in the Pliocene, only major rivers such as the Orange (Gariep) maintained their flow through the arid hinterland of Namaqualand and the Namib Desert after this time.

The Post-African I Surface, which resulted from the cycle of erosion initiated in the Miocene, dominates much of the southern African landscape today (Figure 1.3). Advanced planation was achieved only in a few areas such as the Lowveld portions of Mpumalanga and adjoining northern KwaZulu-Natal, and the Springbok Flats in the Bushveld Basin. In most regions, landscape development was limited to the removal of the deep weathering mantles of the earlier African Surface, leaving scattered residuals of the latter above a rolling surface cut shallowly into the underlying rocks. In areas subject to pronounced structural influences, bouldery koppies commonly provide the modest relief which characterises the Post-African I Surface of the interior, as in the Karoo and Northern Cape. In some coastal areas of the Western Cape, which experienced minimal uplift in the end-Early Miocene, the full thickness of the earlier kaolinised saprolite was not everywhere removed.

4. The Pliocene uplift

In contrast to the relatively small amplitude of the end-Early Miocene uplift and the limited incision that it produced, elevation of the eastern hinterland in the Early Pliocene was likewise of an epeirogenic, but major nature. Uplift and marginal seaward tilting took place along the same axes in the coastal hinterland, as had been the case previously. This time, however, the uplift was massive and was accompanied by a major new pulse of offshore sedimentation. As can be seen in Figure 1.3, uplift along the southeast coast of the subcontinent varied in the range 600 to 900 m, which added to the uplift that had already occurred in the end-Early Miocene gave a cumulative total uplift of over 1 000 m in this region. Uplift along the Cape southern coast was only some 200 m, while along the west coast it amounted to only some 100 m. At the same time the Bushveld Basin subsided by about some 400 m. It was this major asymmetrical uplift of the subcontinent which gave its present topographic configuration, as may be seen in Figure 1.4.

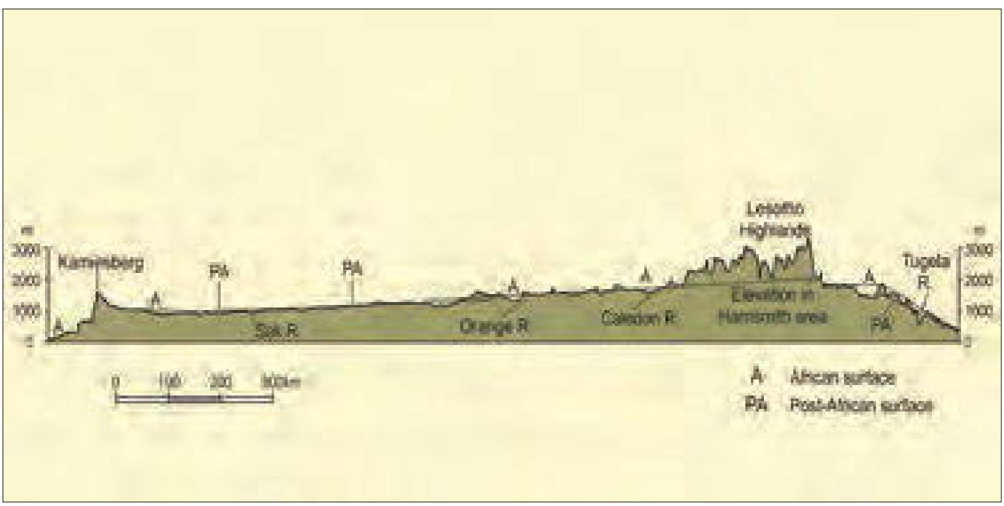

Figure 1.4. Cross-section, east-west, across southern Africa showing erosion surfaces and slight westward tilt of the subcontinent (simplified after Partridge and Maud, 1987).

The amount of uplift that occurred at this time and the location of its axis can be assessed from the form of profiles across remnants of the African and Post-African I Surfaces measured at right angles to the coast. Across the axis of movement, which is located approximately 80 km from the coast and parallel to it, the long profiles of rivers are convex upward, reflecting the influence of this recent uplift and warping in spite of the presence of significant lithological and structural controls. Successive profiles of the African Surface, reconstructed from the remnants of its diagnostic pedocrete cappings and deep weathering profiles, are similarly deformed and gradients on the seaward side of the axis of maximum uplift are as high as 40 metres per kilometre (Figures 1.2 and 1.3). This steepened seaward inclination contrasts with values of approximately one to three metre/s per kilometre in areas of advanced planation which have been unaffected by warping or tilting. Gradients across remnants of the Post African I Surface are only marginally less steep, the separation between the two sets of profiles across the maximum uplift axis being a measure of the much smaller movements which took place earlier in the Early Miocene (Figure 1.2). In the area between Paterson and Bathurst in the Eastern Cape, marine deposits identified as being of Early Pliocene age (King, 1973) have been raised to elevations of up to 400 m above present sea-level along the seaward flank of the axis, less than 20 km from the coast. Similarly Late Cretaceous and Eocene marine sediments have been elevated to altitudes of approximately 350 and 400 m respectively at Needs Camp, west of East London, (Figure 1.2) approximately 22 and 26 km from the coastline (Partridge and Maud, 1987).

Since maximum Pliocene sea-levels are now accepted as having been significantly less than 100 m these lines of evidence can be used severally and in combination to bracket the amplitude of uplift on the east to between 700 and 900 m. Along considerable stretches of both the southern and western coastlines, however, uplifts at this time were limited to 90 to 110 m.

Absolute elevations above the Great Escarpment, to the west of the uplift axis, decline progressively from 1 750 m to the east of Harrismith, to 1 180 m on the Bushmanland Plain east of Port Nolloth on the Atlantic coast (Figure 1.4). The effect of the Pliocene uplift was, therefore, to impart a second slight increase to the gradients of westward flowing rivers. In contrast, their eastern counterparts were significantly steepened leading to major gorge cutting and dissection, the 500 m deep Valley of a Thousand Hills and the nearly 1 000 m deep valley of the Thukela River in the hinterland of coastal KwaZulu-Natal being spectacular examples of this.

A further consequence of the Pliocene uplift was the increase of relief along the eastern and southern parts of the Great Escarpment as a result of incision in the headwater reaches of most rivers. The response offshore was an increase in sedimentation rates to near the high values of the Cretaceous, this change occurring above a well-defined seismic acoustic reflector whose age has been put at around 5 Ma on micropalaeontological evidence (Martin, 1987).

The Pliocene uplift terminated the cycle of erosion giving rise to the Post-Africa I Surface, and initiated the succeeding Post-African II erosion cycle and surface. By contrast to the very long duration of the African Surface planation of some 120 Ma, whereby extreme planation conditions were realised, the relatively short duration of the Post-African I Surface cycle of erosion of only some 15 Ma, ensured that it was not possible to achieve the same degree of planation during the time of its currency as had been possible in the earlier African Surface multicyclic erosion period. Even less planation was possible in the Post-African II erosion cycle of only some 3 Ma duration before this was terminated by drainage changes consequent on climatic and glacio-eustatic sea-level fluctuations related to the waxing and waning of ice-sheets in high latitudes after about 2.6 Ma.

In the interior of the subcontinent the Pliocene uplift was accompanied by rejuvenation along the Griqualand-Transvaal and Kalahari-Rhodesia axes and simultaneous subsidence of up to approximately 400 m within the Bushveld Basin (Figure 1.3), here evidently, partly along major boundary faults. At much the same time, an axis developed between Vredendal and Cape Agulhas in the Western Cape

(Figure 1.3), subsidence to the west of this axis resulting in the depression of coastal remnants of the African Surface to below sea-level in places here, at Noordhoek south of Cape Town (Partridge and Maud, 1987).

While the most extensive result of the Pliocene was renewed incision within river valleys, in some weaker unprotected lithologies, a considerable degree of Post-African II planation was achieved. Examples of this are the eastern Lowveld areas of Mpumalanga, Swaziland, and northeastern KwaZulu-Natal.

For a discussion of the possible mechanisms of continental uplift, where such elevated areas are flanked by rifted passive continental margins, continuing neotectonics affecting the subcontinent, etc. (see Partridge and Maud, 2000). In the matter of isostatic uplift of the marginal areas of the subcontinent, there is minimal, if any surviving evidence for this. Thi is not surprisingas the major period of erosion of the subcontinent was during the Cretaceous.

5. Later Plio-Pleistocene geomorphic events

After the major Pliocene uplift and the initiation of the Post-African II erosion, approximately 2.6 Ma ago, the geomorphic development of the subcontinent became dominated by climatic and glacio-eustatic sea-level movements consequent on the onset of succeeding glacial and inter-glacial conditions in global regions of high latitude, with no major events of a diastrophic nature affecting it thereafter, although locally minor deformations occurred.

In coastal areas, sea-level declines during the glacial periods exposed the continental shelf and aeolian activity gave rise to the Maputaland, Algoa, and the Bredadorp Groups, as well as the Knysna Formation and the Wilderness Dune Cordons along the east and southern coasts, (Maud and Botha, 2000). On the West Coast, the comparable deposits of the Sandveld Group were deposited (Pether, Roberts and Ward, 2000). Commencement of the deposition of some of the Groups had already commenced in the late Miocene. During periods of low sea level, as at 18 ka, when during the Last Glacial period, sea level was approximately 120 m below that of the present, the rivers in the immediate coastal zone incised their lower courses accordingly, these subsequently being filled, or partially filled, with alluvial and estuarine sediments during and after the rise of sea level to that of the present in the Holocene. On the exposed continental shelf, dune deposits accumulated. The onshore fringing coastal deposits were laid down on a series of marine-cut benches which were cut at successively lower levels as still-stands occurred during the overall decline in sea level that took place during the Late Pliocene and the Pleistocene. The most notable of these are the diamond-bearing terraces on the West Coast. A buried boulder bed overlying the well-developed marine-cut bench at an elevation of 70 m above present sea level in the Durban area has recently been cosmogenically (^{26}Al / ^{10}Be) dated to 4.3 +/- 0.68 Ma. Corresponding climatic changes during this time affected vegetation cover in most places markedly.

In the interior, the beginning of the Kalahari Desert probably dates to the beginning of this time (Partridge, Botha and Haddon, 2006). Climatic changes gave rise to the formation of flights of terraces on major rivers particularly in the western interior as river flows changed in response thereto, the most notable of these being the diamondiferous Vaal River gravels. A number of changes to drainage patterns by river capture also took place at this time. Climatic changes in the Late Pleistocene are also evidenced in the colluvial palaeosol sequences which characterise most of the now drier basins of the major rivers below the Great Escarpment in the east (Botha and Partridge, 2000). The formation of the soil cover over much of the subcontinent took place during the most recent part of this period, the Holocene.

6. Summary

The geomorphic evolution of southern Africa since the breakup of Gondwana is summarised in Table 1.2 below (after Partridge and Maud, 1987).

Table 1.2. Summary of the macroscale geomorphic evolution of southern Africa since the Gondwana break-up (after Partridge and Maud, 1987).

TIME	EVENT	GEOMORPHOLOGY
Late Pliocene to Holocene	Climatic fluctuations glacio eutatic sea level changes	Marine benches, coastal and interior dunes, river terraces
	Post-African II erosion	Post-African II Surface formed (limited extent), incision of gorges
Mid Pliocene (~ 5 Ma)	Major Uplift (up to 900 m in the east)	Asymmetric uplift and westward tilting of the sub-continent, marginal seaward tilting of erosion surfaces
Mid Miocene (~ 8 Ma) to Mid Pliocene	Post-Africa II erosion	Post-African I erosion surface formed (imperfectly planed) major deposition in Kalahari Basin
Mid Miocene (~ 18 Ma)	Moderate Uplift (200-300 m)	Termination of African Surface (multicyclic) erosion, westward tilting of African Surface in interior, marginal seaward tilting of African Surface
Late Jurassic – Early Cretaceous (~ 140 Ma) to Mid Miocene	Multicyclic African erosion	Extensive advanced planation of African Surface (at different levels above and below the Great Escarpment, deep weathering and duricrust formation on surface).
Late Jurassic – Early Cretaceous (~ 140-125 Ma)	Gondwana Fragmentation	Subcontinent formed by rifting, rapid erosion to new base levels, initiation of Great Escarpment in marginal areas

Note: It is of interest to note that, although there are some relatively minor time differences, the post-Mesozoic geomorphic evolution of southern Africa as set out above, is generally paralleled by that of East and West Africa, and the other Gondwana fragments, eastern South America, India and Australia, not least in the presence there also of uplifted pronounced marginal escarpment features.

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2

Lithological and Structural Controls on Landforms

Lithological and Structural Controls on Landforms

Peter J Holmes

1. Introduction

The previous chapter provided a review of the geomorphic evolution of southern Africa from the pre-Cenozoic to the present. Because of the thematic structure of this book, and because it focuses primarily on current trends in geomorphic research in southern Africa, there are aspects of landform and landscape development which are not necessarily dealt with individually. This chapter touches on aspects of the present-day geomorphology of the subcontinent not covered elsewhere in this book. The rationale is to provide the reader with some background in terms of landforms where there is significant endogenic influence or control. Such landforms frequently demonstrate strong lithological and/or structural control. The chapter further looks at how these controls, and the resultant landforms, relate to the southern African landscape. Readers with a particular interest in geological controls are referred also to Chapter 3.

Clearly, the present-day landscapes of southern Africa are a function of its past evolution. In the more than two decades since the original *Geomorphology of Southern Africa* was published (Moon and Dardis, 1988) some aspects of the subcontinent's geomorphology have continued, as in the past, to receive attention from researchers, some have enjoyed significantly more attention (for reasons which will become apparent in the introductions to many of the individual chapters which follow), and some, for a variety of reasons, have received little or no further attention. This chapter, therefore, fulfils a dual role. Firstly, as mentioned, it provides a broad overview of those aspects of the southern African geomorphic landscape that are not explicitly dealt with elsewhere in this book and, secondly, it provides a brief synopsis of those facets of geomorphology which no longer necessarily enjoy, within the southern African context, the attention they previously did. Without mention of them, this book could, arguably, be incomplete. The two objectives are not seen as mutually exclusive, and are here considered via an integrated approach.

The original *Geomorphology of Southern Africa* (Moon and Dardis, 1988) included separate chapters on slopes and structural control on landforms, and karst landscapes. Within the southern African context, neither of the above has continued to enjoy, *sensu stricto*, the academic attention which they previously did. Reasons for this are suggested in the introduction to this book. Similarly, surficial karst geomorphology has all but disappeared from the southern African geomorphic research agenda in spite of the fact that some two percent by area of the southern African landscape comprises soluble limestone or dolomite (Marker, 1988). In this book, karst is subsumed in this chapter for the simple reason that southern African karst (speleology excluded) is not currently high on the research agenda of any member of the geomorphological community, either local or foreign.

2. The morphology of landforms

The morphology, or shape of individual landforms is primarily controlled by the inherent geological characteristics of the region (endogenous, or endogenic controls), or by surficial processes of weathering and erosion acting on the landscape (exogenous or exogenic controls), or by a combination of both. This being a geomorphology text, exogenic controls and processes form the foci of most chapters, although, as can clearly be appreciated from Chapters 1 and 3, geological controls cannot be ignored. Exogenic controls are, to a greater or lesser extent, dependent on local environmental conditions. These include climatic conditions, albeit the extent to which climate exercises a significant control on the morphology of a landscape is a topic of debate (Summerfield, 1990). This book devotes an entire chapter (Chapter 12) to Landscapes and Environmental Change, an area that, currently, probably enjoys more attention from the southern African geomorphological community than any other. Because exogenic controls on landform development enjoy attention within the context of virtually every chapter of this book, they are not further pursued here. This chapter focuses primarily on endogenic, or geological controls on the development of southern African landscapes. Both the inherent geological properties of a rock type, and the way in which rock types are arranged relative to one another in a particular region may have a prominent, indeed even a defining, influence on the morphology of individual landforms, or an entire landscape.

It is useful at this juncture to consider the concept of geomorphic provinces. A geomorphic province (see also Chapter 14) is "…the highest level of organization of the geomorphology hierarchy" (Partridge *et al*., 2010:2). The notion of landforms and assemblages of (recurring) landforms is a familiar one. Geomorphic provinces may be regarded as spatially delimited areas in which a certain landform assemblage prevails. By definition, a geomorphic province is unique, or distinguishable from its neighbouring province(s). King (1942) identified 26 geomorphic provinces for southern Africa and these were later re-constituted as 16 provinces (King, 1951) and still later to 18 (King, 1967). King's work incorporated earlier work by a number of pioneers in southern African physical geography, including Wellington (1944, 1955). For a fuller explanation, see Partridge *et al*. (2010) as well as Chapter 14 (Geomorphology and Remote Sensing) in this book. Geomorphic history, geological structure, altitude, location and climate all play a role in identifying a unique province. Partridge *et al*. (2010) delimited 34 geomorphic provinces and 12 sub-provinces across South Africa, Lesotho and Swaziland. Geomorphic provinces are, of course, not constrained by geopolitical borders. Thus, the Kalahari province also covers a considerable portion of southern Namibia and Botswana, while the Lowveld reaches into Mozambique, and the Limpopo Flats into Botswana and Zimbabwe.

Partridge *et al*.'s (2010) delineation is based primarily on statistically defined changes in slope, as well as valley cross-sectional width, along each of 99 southern African longitudinal river profiles. The delineation is designed to serve as a tool for a range of catchment management applications, including the protection of freshwater ecosystems, and biodiversity. It is a complex, detailed model, and not necessarily the most appropriate in terms of a relatively simple geomorphic descriptor. Nevertheless, it is an exhaustive work, to which the reader with an interest in detailed morphological descriptors of the geomorphology of this part of southern Africa is referred.

A simpler model, not referenced by Partridge *et al*. (2010), is the terrain morphology map of southern Africa produced by the Soil and Irrigation Research Institute (Kruger, 1993). Kruger (1993) identified 12 broad terrain patterns (see Figure 14.5 in Chapter 14 of this book) covering South Africa, Lesotho and Swaziland, and went on to define six terrain morphological classes, ranging from plains with low relief, to mountains and table lands. Each class has a brief descriptor, a description of slope form, and a note on drainage density.

The post-Gondwana geomorphic evolution of southern Africa is described in the preceding chapter. It is abundantly clear from a drainage map (see for example Map 1 in Partridge *et al*. (2010)) what influence the break-up of Gondwana had on the flow direction of southern Africa's rivers. In spite of its relative aridity, fluvial processes have undoubtedly played the leading role in shaping the southern African landscape. These processes are dealt with in Chapter 5 (Fluvial Geomorphology) and, to an extent, in Chapter 7. The remainder of this chapter is devoted to an overview of the inherent physical properties of the subcontinent; in other words, the *passive* (this discounts neotectonic forces) geological environment which, influenced by exogenic processes, has yielded the landscapes we see today.

Rock type, referred to as lithology (Figure 2.1), is the primary geological characteristic of any landscape. Landforms are, typically, the end or ongoing manifestation of processes operating on specific lithology or lithologies. In certain landscapes, lithology exercises a dominant control on the morphology of the landscape, to the extent that it either overrides the direct influences of surface processes or relegates such processes to a secondary role. This seems frequently to be the case with granite (see Chapter 3) and limestone (calcium carbonate) or dolomite (magnesium carbonate) rocks. The latter two are referred to as *karst*. Both granite and karst occur to a significant extent in southern Africa (Twidale, 1988, 2012; Marker, 1988). In this book, granite landscapes are dealt with in a separate chapter, primarily because of significant interest and progress in research into these landscapes over the last two decades. Karst landscapes have, from a geomorphological perspective, not received the same attention.

Figure 2.1. Differences in lithology. Robberg Formation conglomerate unconformably overlain by sandstone in the southern Cape.

Structure or stratigraphy refers to the way in which lithological units occur in relation to one another. The term stratigraphy refers to the juxtaposition or sequential arrangement or layering of, typically, sedimentary and volcanic rocks. The primary manifestation of such control is frequently observed in terms of slopes and slope morphology. Stratigraphy (a geological phenomenon) provides for structural (a geomorphological phenomenon) control. Structural control is an important determining element in many southern African landscapes at the macro- to mesoscale. The macroscale geomorphic evolution of southern Africa is described in Chapter 1; this chapter focuses on the influence of structural controls on landforms and the occurrence of slopes at the local (landscape or landform specific) scale. Note too that lithological and structural controls are frequently inseparable. Thus, for example, the typical Karoo landscape of flat-topped mesas and concave slopes is a function of both lithology (resistant sandstone and dolerite and less-resistant mudstone) and structure (the way in which the rock types are horizontally interspersed). This is further discussed below.

Figure 2.2. Structural control on a Karoo landscape near Ceres. Note the dipping bands of sandstone.

3. Landscapes where lithology exercises the primary control

3.1 Karst

The term *karst* derives from the Krs area of Slovenia (once part of northwestern Yugoslavia) and which is typified by stony, barren rock (Marker, 1988). The term was subsequently applied globally to regions of limestone and dolomite, where large-scale dissolution of these soluble rocks has created unique surface and sub-surface features. Geomorphologically, the driver of karst is atmospheric CO_2 dissolved in rainwater to form a weak acid. This in turn can dissolve carbonate rocks to form a distinctive suite of geomorphic features, both at the surface, and underground. Lithology is, therefore, the dominant control in that, provided it is a humid environment, the inherent solubility of the rock type largely determines the morphology of the landscape. Marker (1980) proposed a systems model for the development of karst (Figure 2.3).

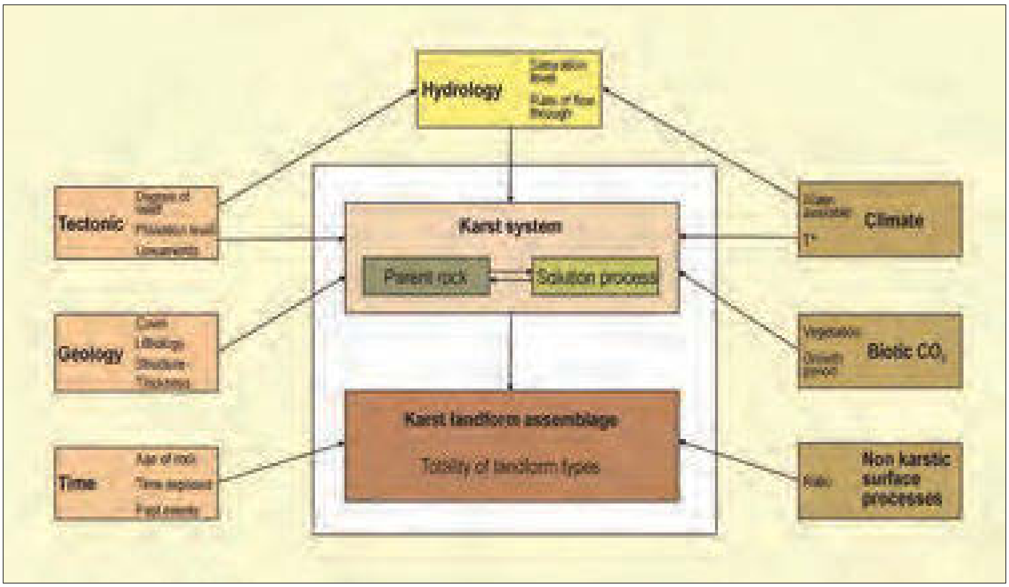

Figure 2.3. A karst systems model (after Marker, 1980).

As mentioned, some two percent of southern Africa is underlain by carbonate rocks (Marker, 1988). In the interior of South Africa, karst landform assemblages typically occur on dolomites of the Transvaal Supergroup (Gauteng, North West, Limpopo and Mpumalanga Provinces), and on the Ghaap Plateau west of Kimberley (Griqualand West Sequence of the Transvaal Supergroup). The most famous South African examples are undoubtedly the Cradle of Mankind (Sterkfontein Caves and surrounds). Younger (Tertiary) coastal limestone occurs along the southern Cape, and northern KwaZulu-Natal Province coastlines. The Cango Caves of the Little Karoo are also formed in karst. Karst also occurs in Botswana and Namibia (Marker, 1988). South African karst received considerable attention in the 1960s 1970s and 1980s, not least because of the dangers associated with collapse, particularly where underground water is extracted on a large scale (Brink, 1979). From a geomorphic perspective, most work on karst occurred prior to 1990 and was reviewed by Marker (1988; see also Figure 2.4).

Figure 2.4. The distribution of carbonate rocks in southern Africa (after Marker, 1988) and the position of various features mentioned under 4. Landforms and landscape where structure exercises significant control.

Karst landscapes are characterised by dissolution of limestone and dolomite to create a unique suite of surface features that include potholes, sinkholes, dolines (circular hollows or depressions), uvalas (very large depressions, often formed when adjacent dolines coalesce), poljes (enlarged, level areas subject to lateral solution), swallow holes (where surface drainage disappears underground), springs (where underground drainage reappears) and dry valleys. Underground features include caves and caverns (complete with a range of speleothems such as stalactites, stalagmites and pillars) and underground streams. Southern African caves and speleothems have been extensively researched (e.g. Holmgren *et al.*, 1999) because they are a valuable archive for palaeoenvironmental reconstruction. Finally, an absence of surface drainage is a distinctive and characteristic feature of karst. A review of research into South African karst prior to 1988 was reported by Marker (1988). Occasional sub-regional (e.g. Russell, 1989) and local (Marker and Swart, 1995; Marker and Craven, 2002) studies on karst and pseudokarst (karst type morphology developed on a resistant rock type such as Table Mountain Group sandstone) have appeared subsequently.

3.2 Granites, flood basalts and dolerite

Granites are ubiquitous in southern Africa's geology, occurring from Zimbabwe in the north, to the Cape Peninsula in the extreme south. For this reason, and because they have continued to receive attention from a research perspective, they form the subject of a separate chapter (Chapter 3) in this book. Twidale (2012) mentions that weathering on a range of rock types that differ lithologically from granite but which are physically similar (impermeable, usually well-jointed) can produce etching forms similar to those on granite. Such rocks include gabbro, diorite, dolerite and basalt (Figure 2.5). Dolerites are ubiquitous in the Great Karoo, and common in other regions as well. They typically manifest themselves as prominent dykes and sills (Figure 2.6).

Figure 2.5. Drakensberg Formation basalt capping Clarens Formation sandstone at high altitude in the extreme north-eastern Free State Province, South Africa (after Holmes and Barker, 2006).

Figure 2.6. Dolerite dyke intruding Stormberg Group sedimentary rock (including prominent Clarens Formation sandstones) in the eastern Free State Province, South Africa.

4. Landforms and landscape where structure exercises significant control

4.1 Domes

Geologically, domes are anticlinal structures or upfolds, where all the strata dip away from a central point. In other words, domes are structurally controlled. Geomorphologically, the term may also be used for a rounded hill, mountain or outcrop. The well-known Paarl Rock (Figure 3.3, and Figure 3.22 in Chapter 3), which exhibits lithological control, is an example. Granite domes are discussed in more detail in Chapter 3.

4.1.1 Vredefort Dome

South Africa's best-known dome is the Vredefort Dome. This unique geological feature represents the largest and oldest meteorite impact structure on earth. The 80 km wide Vredefort Dome is located 260 km north of Bloemfontein (Figure 2.4). It represents the eroded remnant of the original Vredefort impact structure. Following the impact there was, in the centre, a massive upwelling of magma, which created an upheaval dome, with the crater radiating outwards in a series of rings beyond this dome to form the entire structure. The original width was ~250 to 300 km (Gibson and Reimold, 2001; Lana *et al.*, 2004), and the event occurred ~ 2 020 Ma ago (Kamo *et al.*, 1996). The present day feature comprises a 40 km wide core of Archean (see geological timescale on page 4) gneisses. This is rimmed by a ~20 km wide collar comprising subvertically dipping Late Archean to Paleoproterozoic supracrustal strata (Lana *et al.*, 2003). Phanerozoic sedimentary strata and dolerite sills to the south and southeast largely obscure the structure (Lana *et al.*, 2003). The Vredefort Dome, which continues to provide vital insights into the geological evolution of the earth, was declared a World Heritage Site in 2005.

4.2 Basins and cuestas

Basins are, typically, large-scale structural down-folds or depressions formed by tectonic activity, or through fluvial erosion. At the macroscale, Karoo basins in southern Africa represent vast accumulations of sedimentary fill under the dual control of tectonism and climate (Catuneanu *et al.*, 2005; Figure 2.7). The Kalahari Basin is a very large (2.5 million km^2) depression covering most of Botswana, and parts of South Africa, Namibia, Zimbabwe and Zambia. It serves as a sediment sink in terms of the sand, which comprises the Kalahari Desert, and is discussed in Chapter 6. River basins at sub-continental scale include the Limpopo and Orange River Basins.

Cuestas are erosional features resulting in asymmetrical ridges comprising a long, gentle slope, which corresponds to the direction of dip, and a short, steep slope forming an escarpment, such as the Lebombo Ridge in eastern southern Africa. In the case of a hogback, the angle of the dip and strike (scarp) slopes are more or less equal. The *Hogsbacks* of the Amatola Range in the Eastern Cape Province, comprising dipping Karoo dolerite sills, are South Africa's best-known example. The term *homoclinal ridge* is sometimes used to define a structural landform where the dip angle is greater than that of a cuesta, and less than that of a hogback. The Magaliesberg ranges of Gauteng are typically described as homoclinal ridges. Used in the generic sense, both cuestas and hogbacks are homoclinal (uniform dip along the slope).

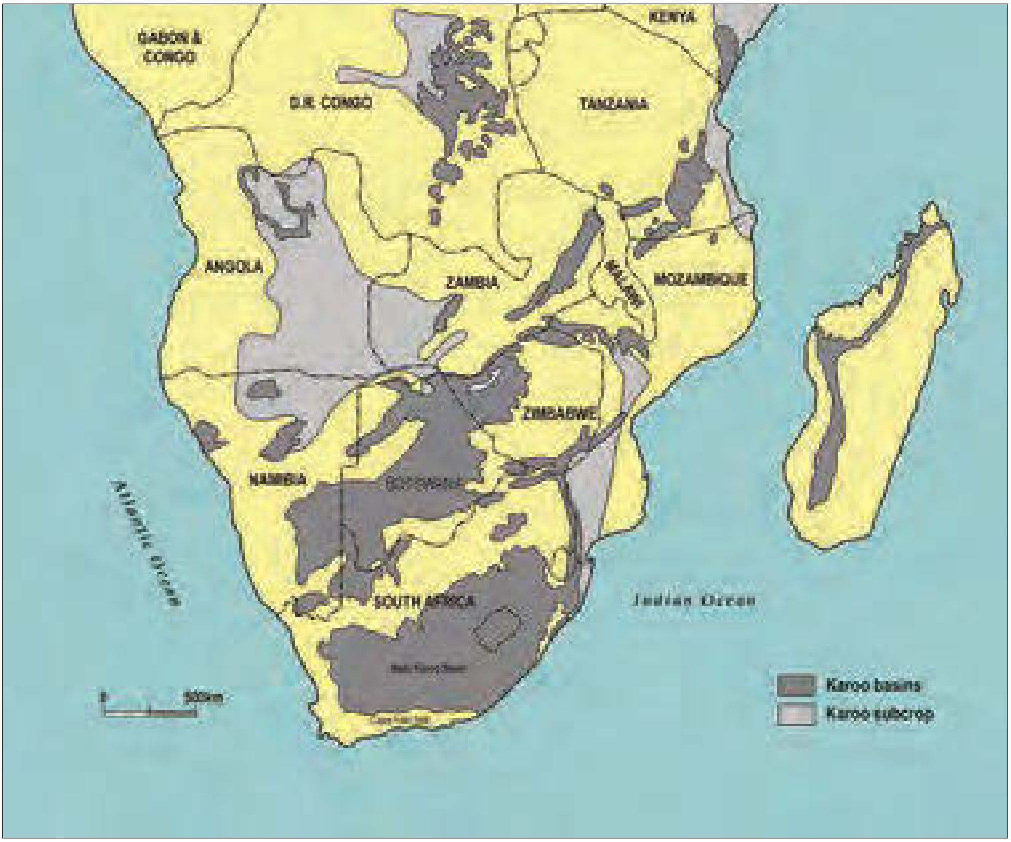

Figure 2.7. The distribution of Karoo basins in southern and central Africa. Note the situation of the Main Karoo Basin (after Catuneanu *et al.*, 2005).

5. Landscapes where lithology and structure combine to exercise significant control

The term *Karoo* can be used in a physiographic sense (Kruger, 1983; Partridge *et al.*, 2010) in an ecological sense (Low and Rebelo, 1996) or to describe the rocks of the Karoo Supergroup that cover approximately 60% of southern Africa. For an overview of the Karoo Supergroup and Karoo basins respectively in the context of all of southern Africa see Johnson *et al.*, 1996; and Catuneanu *et al.*, 2005. The sedimentary rocks of the Karoo Supergroup were deposited by a combination of ice, water and wind from the late Paleozoic into the early Mesozoic. They were subsequently intruded by volcanics (dolerites) or, in places, covered by extrusive basalts (Drakensberg Lavas). Although the margins of the Karoo basin suffered extensive deformation with the Gondwana rifting (see Chapter 1), in much of the Main Karoo Basin (Figure 2.7), the horizontal stratigraphy has been maintained. The result is strong lithological, or structural control, or a combination of both. The succession of Karoo sedimentary strata comprise, in sequence from bottom to top, the Dwyka and Ecca Groups, the Beaufort Group, and the Molteno, Elliot and Clarens Formations (Stormberg Group), capped by the Drakensberg Group lavas (Figure 2.5).

Tillites of the Dwyka Group exercise strong lithological control. Weathering along planes of axial cleavage forms distinctive *tombstone* features, which are particularly pronounced in the Laingsburg area of the Western Cape Province, South Africa. The Beaufort Group and, to an extent, the Ecca Group display arguably the best examples of structurally controlled morphology. The largely horizontal stratigraphy has resulted in the classic Karoo landscape of mesas (flat-topped tablelands) buttes (flat-topped hills) and conical hills (typically referred to by the Afrikaans term *koppies*) where the hard, capping dolerite has been removed by weathering. Koffiebus and Teebus near Middelburg in the Eastern Cape, and the Three Sisters of the Western Cape (Figure 2.8) are well-known buttes. Dolerite intrusions are lacking in the southern Karoo, therefore typical Karoo topography is more pronounced in the central and eastern Karoo. Multiple dolerite sills also produce, subsequent to erosion, slopes with a classic structural morphology of cliffs (dolerite) alternating with short, typically concave slopes comprising the older sedimentary rock into which the dolerite intruded.

Figure 2.8. Structural and lithological control; the Three Sisters in the Karoo. Resistant dolerite remnants cap these buttes. Prominent bands of resistant sandstone on the mesa in the foreground interrupt the slightly concave profile of this predominantly mudstone feature.

Sedimentary rocks (mudstones and siltstones) of the Molteno and Elliot Formations produce a more undulating topography with a relatively subdued relief, as rapid weathering precludes the presence of steep slopes. However, where dolerite is present, the landscape may be rugged. Indeed, the highest point in the Upper Karoo, Kompasberg (near Middelburg in the Eastern Cape) is a doleritic feature. Clarens Formation sandstones, typically of the eastern Free State Province, South Africa, produce spectacular sandstone cliffs (Figure 2.8). Differential weathering produces overhangs and cliffs (Moon and Munro-Perry, 1988). Grab *et al.* (2011) identified and studied 27 macro- and microscale landforms on sandstone in the Golden Gate area of the eastern Free State. Drakensberg Group basaltic lavas are generally horizontally stratified as a result of a large number of flows of varying thickness. They cap the Karoo rocks to form the Drakensberg and Maluti Ranges of the eastern escarpment.

6. Slopes

For all practical intents and purposes, the surface of the Earth comprises assemblages of slopes (Young, 1972). Such slopes in turn comprise various elements. Descriptions can be found in virtually any undergraduate textbook on physical geology or geomorphology. More detailed explanations of slope morphology may be found in, *inter alia* Young (1972); Carson and Kirkby (1972); Small and Clark (1982); Selby (1982); and Chowdhury *et al.* (2009). Slope form, slope evolution, slope processes and the relationship between form and process have received attention in southern Africa (King, 1953; Moon, 1988), particularly during the last century. Similarly, the influence of lithology on slope (see above) has received attention. For a comprehensive, but elementary text on slope, the reader is referred to Young (1972). Slopes may vary from simple, single-element forms (convex, concave, and rectilinear) to complex profiles comprising a number of elements or facets. Classic slope elements are illustrated in Figure 2.9.

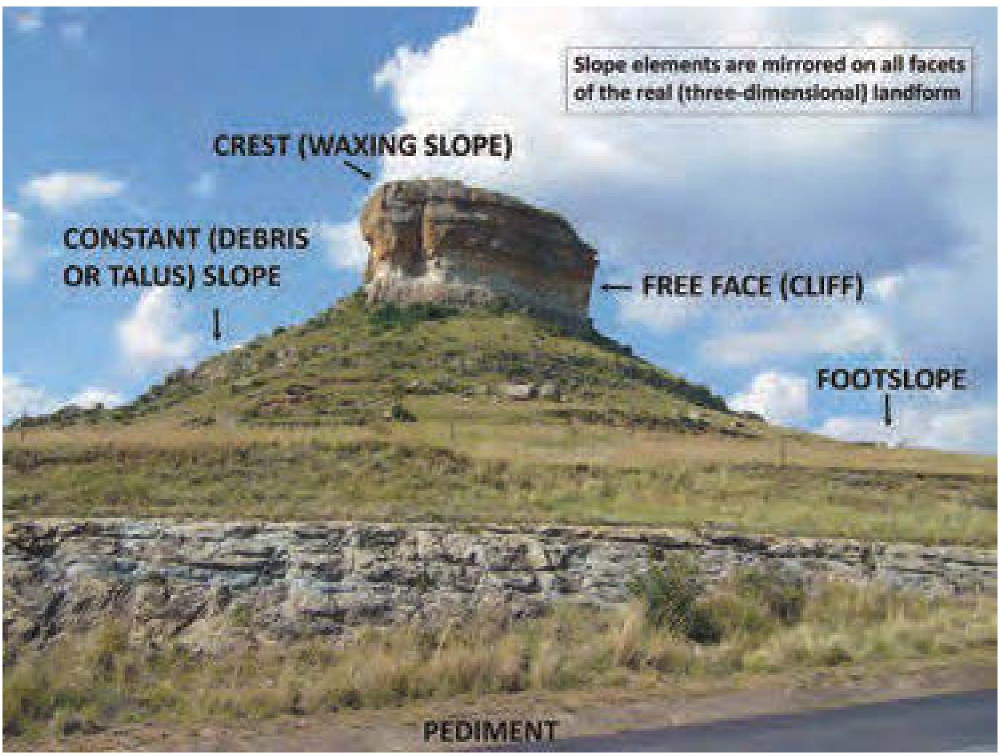

Figure 2.9. Slope elements here illustrated on a butte in the eastern Free StateProvince, South Africa. Resistant cliffs comprise Clarens Formation sandstone. The footslope extends into a broad pediment, from where the photograph was taken.

Slopes are, ultimately, a product of endogenic (geological) forces, which originate in the Earth's mantle (folding, faulting, and volcanism) and the lithosphere, creating uplift in the crust. In the long term, the exogenic or surface processes of rock weathering, erosion and mass movement, which are driven by the atmosphere, and by the force of gravity, will act to balance the endogenic forces. Slopes are a manifestation of both endogenic and exogenic drivers. Slopes can develop on bare rock (see granites above) as well as on surfaces initially covered in rock debris or soil. Similarly, rock debris can accumulate on slopes, and soil can develop on such debris.

Rates at which slope processes occur depend on environmental conditions, the nature of the material on which the slope is formed, local relief, and the gradient of the slope. Slopes are dynamic, and may be stable or unstable. This in turn has significance for the way slopes are managed (see Chapter 13).

Theories of slope development have been around since the end of the nineteenth century. Davis (1899; 1902) proposed a theory of *slope decline*, with the slope being lowered by weathering and erosion. Debris may be deposited to mantle the bedrock slope. Penck (1924) suggested a model of *slope replacement*, in that steep slopes will retreat from their initial positions and be replaced by slopes of lower gradient. Again, debris will cover and protect the lower slope, so the length of the exposed, retreating face becomes shorter until, with time, it may be completely buried. In effect, a gentler slope replaces a steep slope. Yet another important theory is that of *parallel retreat* of slopes. Here, all the elements of the slope retreat simultaneously. This influential theory was put forward by the South African geomorphologist King (1953, 1957) and is widely referred to as the pedimentation model of landscape evolution. The crest, cliff and talus, or debris slope retreat parallel to themselves (see Figure 2.9), maintaining the original form of the slope and leaving a pediment at the base. Naturally, processes at the base of the slope must remove rock debris at a rate equal to that at which it is being generated upslope, as well as debris that is busy weathering out of the slope beneath the talus material. Figure 2.10 shows an example where accumulation exceeds removal. The front (main) face of Table Mountain represents a classic example of debris mantling, with the thickness of the debris mantle increasing towards the base of the slope. Debris mantling is also very evident in the Golden Gate area of the Free State (Figure 2.11).

Figure 2.10. Massive sandstone blocks have accumulated on this slope on the Robberg Peninsula at Plettenberg Bay in the southern Cape. Debris accumulation far exceeds the removal rate.

Although King (1953) maintained that the free face or cliff is the key element for parallel retreat, it is widely accepted that such retreat can occur in the absence of a free face. In southern Africa, the analysis of slope development by Fair (1947, 1948) has been noted as being of particular importance (Young, 1972). Fair's comparison of slope profiles indicated that the form of the waxing slopes; free face and talus slope elements of a particular slope remain constant as the pediment broadens. Consequently, the slopes develop by parallel retreat. Fair also demonstrated that such parallel retreat occurred only when two conditions were satisfied; the presence of a resistant (dolerite sill) caprock, and efficient removal of debris from the base of the slope by sheetwash. The classic mesa, butte, conical hill (where the dolerite caprock is entirely removed) sequence of the Karoo appears to confirm Fair's observations.

Figure 2.11. A mantle of weathered debris creates a slope below the cliff, on which further, unweathered Clarens Formation rock debris has come to rest on the flanks of this ridge in the eastern Free State Province, South Africa.

7. Conclusion

This chapter has provided a brief background into lithological and structural controls on landscape evolution and development, as well as a short introduction to slopes. With the possible exception of Antarctica and the southern ocean islands (not covered in this book, largely because of page constraints), none of these currently feature highly on the southern African geomorphic research agenda. This is not to say that lithological and structural controls, and slope evolution are discounted by southern African geomorphologists. Rather, they form part of integrated and often inter-disciplinary approaches to the study of landscape form and function, rather than discrete areas of study *per se*. Much of the classic literature on structural and lithological control, and slope evolution, and in particular the theory underlying macro- to mesoscale landscape evolution, dates back to the first six decades of the last century (see references to, e.g. King and Fair). This chapter does not claim to do justice to the published literature, even in the southern African context. For this reason, the reader with an interest in geological controls on landform evolution, and the development of slopes in the southern African context is referred to Moon (1988).

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3

Granite Landscapes

Granite Landscapes

C Rowl Twidale

1. Introduction

Granitic rocks, including granite gneisses, are widely exposed in southern Africa, and not only in the Rhodesian, Angolan, Richtersfeld, and Kaapvaal cratons (e.g. Truswell, 1977), but also in the intervening mobile zones of which the Cape Fold Belt and the Natal Monocline are examples. Given the extent of granitic outcrops it is not surprising that many basic concepts concerning the origin of granite landforms and landscapes are based in the research of investigators who worked in southern Africa and whose works still are cited not only with reference to southern Africa, but also in the broader context.

Inselberg landscapes are typical of cratonic exposures in southern Africa. Some comprise extensive plains interrupted only by scattered residual hills or *island mountains* of various types standing either in isolation or standing in ordered rows in granitic massifs (Figures 3.1 and 3.2). Occurrences of intrusive granite in fold mountain belts are less extensive but nevertheless distinctive. Paarlberg, in the Western Cape Province of South Africa, for example, is a group of domical hills (Figure 3.3), as are the Valley of a Thousand Hills, in the KwaZulu-Natal Province of South Africa, and the Matopo (Matobo) Hills of Zimbabwe.

Figure 3.1. Inselberg landscape near Twyfelfontein, northwest Namibia.

Figure 3.2. Aligned granitic bornhardts in the Kamiesberg, Namaqualand, Northern Cape Province, separated by fracture-controlled valleys and with sheet fractures and structures well represented.

2. Physical characteristics of granite: composition and fractures

Various characteristic granite landforms reflect the physical properties of the country rock. It consists of quartz, potassium feldspar, and mica, all of which react with water. Because of its molecular structure water is a powerful solvent (Mason, 1966), and most minerals – even quartz, given the right chemical environment, or sufficient time – are soluble. In the middle and low latitudes the alteration and disintegration of granite is caused primarily by water-related processes such as solution, hydration, and hydrolysis (e.g. Yatsu, 1988). Of these, solution is the first stage of many types of rock alteration (Loughnan, 1969). Water charged with chemicals reacts with rock-forming minerals. Though quartz is resistant, the mica and feldspar are rapidly altered to clay (e.g. Caillère and Henin, 1950; Alexander, 1959). Organisms contained in circulating groundwater not only add to the chemical mix, but live microorganisms bore into crystals and so facilitate water access (Trudinger and Swaine, 1979; Folk, 1994). Such weathering produces *grus*, a fine-grained gravel forming a regolith that meets the fresh rock in the weathering front (Mabbutt, 1961).

Fresh granite is a crystalline rock of low porosity and permeability, but depending on whether they are open or tight, well-developed systems and sets of fractures may impart a degree of perviousness. Active shearing (the *torsion* of earlier writers) has caused granites to be strained and ruptured resulting in the widely distributed systems of orthogonal joints and small displacement faults that subdivide the bedrock into cubic or quadrangular blocks. Recurrent dislocation and fracture propagation have generated more complex patterns. Open fractures are avenues along which water can penetrate, and as water is the primary agent of weathering, fracture control of form and topography are evident in all granitic terrains, with straight valleys and angular drainage patterns typical of such areas.

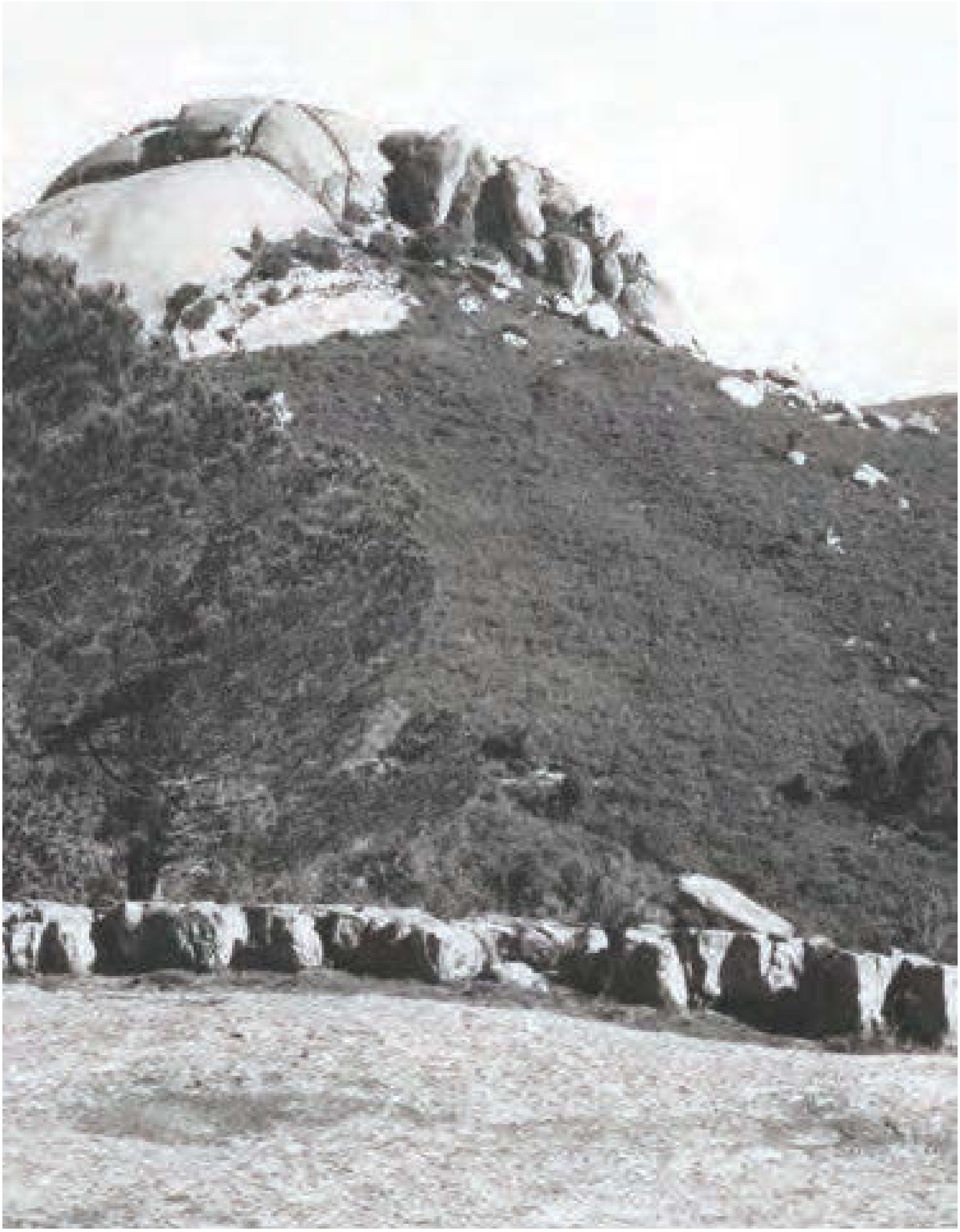

Figure 3.3. Part of Paarl Mountain, Western Cape Province, with sheet structure on crest and in foreground, each subdivided by joints into boulders and blocks.

3. Corestone boulders

Boulders can be defined as rounded, detached rock masses at least 256 mm diameter. They have been formed in various ways. Some boulders are rounded by abrasion during stream or glacier transport, others as a result of the break down, both below ground and after exposure, of sheeting slabs (Figure 3.3), but most granite boulders are of etch origin (Figures 3.4 and 3.5). They are two-stage or etch forms, for *to etch* means to be attacked chemically (Hassenfratz, 1791; MacCulloch, 1814; Branner, 1896; Scrivenor, 1913; Wayland, 1934; Willis, 1936; Linton, 1955). Similarly, the inclined platy or slabby outcrops developed on gneissic granites result from the subsurface exploitation of a well-developed and closely spaced cleavage, and are known variously as penitent rocks, monkstones, tombstones, or *Büssersteine* (e.g. Ackermann, 1962).

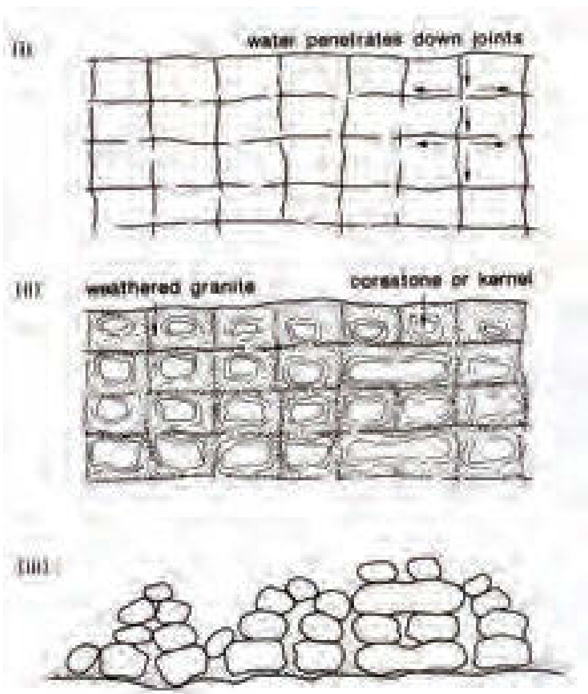

Figure 3.4. Diagram illustrating the two-stage development of corestone boulders.

Meteoric or rain water infiltrating along partings from the surface downward attacks the rock with which it is in contact (Figure 3.4). Mica and some feldspar are altered to clay, which takes in water and expands, forming laminae or flakes (see Larsen, 1948; Hutton *et al.*, 1977; and Twidale, 1986). The corners and edges of the blocks are preferentially weathered because they are attacked from three and two directions, respectively, whereas plane faces are open to only one. The laminae of partly altered rock that are wrapped around the corestones are further weathered and converted to a granular mass of fine gravel or *grus* that consists of granular fragments of quartz and feldspar mixed with clay. In this way the original angular blocks are converted to spheroidal masses of fresh rock known as kernels or corestones, set in a grussy matrix (Figure 3.5). The outlines of the corestones are discrete sectors of the weathering front. As weathering penetrates further into the corestones they are reduced in size. They may be eliminated and be replaced by a puggy clay with a few quartz fragments, though even they are eventually dissolved (Ruxton and Berry, 1957). The relative rates of weathering and erosion determine not only the size, but also the shape of the residual boulders (Lewis, 1955). As the land surface is lowered, the friable *grus* is evacuated, typically under gravity and by rivers, but in coastal settings by waves.

Four general points can be noted. First, corestone boulders are a product of fracture-controlled differential subsurface weathering. That most are formed *in situ* is demonstrated by sites where the granite was intruded by veins of say, quartz or aplite, and despite subsurface weathering, both corestones and intrusive masses remain unweathered (e.g. Kingsmill 1862). Further evidence that the corestones are *in situ* is provided by sites where the corestones retain an outer zone of laminated rock that could never have survived transport either by rivers or by glaciers (Figure 3.5).

Second, weathering acts from the surface downwards (MacCulloch, 1814). Within the regolith the early stages of weathering are evidenced at and near the weathering front, and the most advanced in the near-surface zones, which have been longest in contact with descending meteoric waters. For this reason the corestones there tend to be smaller or even absent, whereas at depth near the weathering front they tend to be larger and less well rounded.

Third, once water enters the fabric of the granite along microscopic fissures, or crystal cleavages and boundaries, selective weathering occurs, resulting in disruption and fractures. This dramatically increases permeability (Kessler *et al.*, 1940) for laminated zones are more readily infiltrated by water than the fresh crystalline rock and the granular *grus* that is next formed even more so. Thus, the more weathered the rock the more readily can water enter, causing even more rapid and intense alteration: a positive feedback or reinforcement effect.

Fourth, although granite is susceptible to alteration when in contact with water, the converse also applies, namely that granite in dry or relatively dry sites is not subject to such alteration and tends to be resistant (Logan, 1851). For this reason, exposed corestone boulders tend to persist in the landscape.

Figure 3.5. Corestones in granite, Paarlberg Quarry, Western Cape Province. Note laminated (weathered) rock on upper side of some corestones.

4. Granite plains

Given a well-fractured rock mass, and either intense and/or long-continued weathering beneath a relatively stable surface, all or most projecting rock masses can be reduced to produce a regolith resting on a weathering front that is a featureless plane surface, which is the shape least susceptible to attack by water. Such etch surfaces occur at various scales from the regional to the local. The extraordinarily flat Bushmanland Surface of Namaqualand and southern Namibia is an etchplain of regional extent cut across various rock types, including granite. Its otherwise unrelieved regularity is interrupted by a few inselbergs based on compartments of particularly massive rock, and occasional clusters of corestone boulders (Figure 3.6). At several sites in Namaqualand, for instance near Platbakkies, kaolinised regolithic remnants protected by silcrete rise above the adjacent plains cut in granitic gneiss (Partridge and Maud, 1987). They form narrow, elongate, mesas, winding in plan, and are construed as inverted former valley floors. The etch character of the adjacent plains is demonstrated by their coincidence with the base of the kaolinitic zone. The extreme flatness of such etchplains is a measure of long-continued subsurface chemical attack during which all but the most durable rock compartments were eliminated.

Figure 3.6. Part of the Bushmanland Surface, an etchplain in granite (boulders in foreground) as well as schist and sandstone and with an isolated granite hill, northern Namaqualand, Northern Cape Province of South Africa.

In addition to etchplains and surfaces, smooth, gently-sloping mantled pediments, cut in bedrock but carrying a veneer of sand or other detritus (Figures 3.7 and 3.8), have been shaped by wash from the adjacent uplands, and are well-developed in the piedmonts of inselbergs (Twidale, 1981a). Rock pediments cut in granite have been attributed to sheet floods (McGee, 1897) but the interpretation has been challenged (Kirk Bryan, in Brock 1977:10-11). Alternatively, the rock pediments may be mantled pediments from which the veneer has been stripped to produce etch forms. Some granite landscapes consist largely of coalesced pediments, or flat cones backed by inselbergs of various types (Figure 3.7). Elsewhere, for instance, in the vicinity of Saldanha in the Western Cape, the coastal plains have been shaped by exoreic rivers to produce broadly rolling landscapes.

Figure 3.7. Inselberg landscape near Polokwane (Pietersburg), Limpopo Province, South Africa, with partly disintegrated bornhardts bordered by gently sloping cones of mantled pediments.

Figure 3.8. Unusually large castellated residual or castle koppie standing on low angle pediment cone, in the Erongo Mountains, near Ameib, some 200 km northwest of Windhoek, Namibia. The upper part of the pediment is strewn with boulders fallen from the backing scarp. Note fracture-controlled precipitous flanks.

5. Bornhardts and related forms

5.1 Characteristics

Corestone boulders are shaped on fracture-defined blocks, as are the smooth, bald, domical hills, known eponymously as bornhardts (Bornhardt, 1900; Willis, 1934). Spectacular turrets and acicular forms occur where prominent steeply inclined fractures have been exploited by weathering (e.g. Rognon, 1967; Seager, 1981). More commonly, however, orthogonal blocks have given rise to rounded residuals that for one reason or another have proved resistant to weathering and hence to erosion. Cratonic bornhardts and other inselbergs rise abruptly from the surrounding plains (e.g. Figures 3.1 and 3.6) for the piedmont angle, the transition from hill slope to plain, and shaped mainly as a result of scarp-foot weathering and erosion, is well developed. Bornhardts typically occur in multicyclic landscapes, that is, in landscapes in which there is evidence in the form of high plains, bevelled crests or distinct shoulders, of more than one phase of planation (Figure 3.9). Like corestone boulders, they are found in a wide range of climatic conditions (King, 1949; Birot, 1958).

Just as gneissic rocks give rise to penitent rocks that are congeners of corestone boulders, so bornhardts shaped in granitic gneiss tend to be ribbed and blocky, as in the Kamiesberg of Namaqualand. The plan shapes of bornhardts are determined by steeply dipping fractures of the orthogonal or rhomboidal systems related to crustal shearing. The convex-upward profiles of such residuals most commonly, though not everywhere, are associated with convex upward sets of sheet fractures, the origin of which is discussed below (Figures 3.2 and 3.3).

Figure 3.9. The bevelled Leeukop, near Vredefort, northern Free State Province, South Africa, developed on a massive compartment of rock and with a rock platform, which is the crest of a nascent bornhardt, exposed in the adjacent plain. The thickness of soil/regolith increases in all directions with distance from the platform.

5.2 Possible origins

Bornhardts are spectacular landforms that have long attracted the interest of geologists, geomorphologists, and travellers alike. They have been explained in several ways.

A few are upfaulted blocks (e.g. Barbier, 1957). Some are intrusive stocks of granite that have resisted weathering and erosion while the country rock has been worn away (Figure 3.10). Some bornhardts are of a rock type or a granite of a composition that is particularly resistant, though many are apparently of the same rock type that underlies the adjacent plains. The suggestion that bornhardts are the last remnants surviving after long-distant scarp retreat associated with valley development (e.g. Holmes, 1918; King, 1949) is not borne out by the field evidence. For example, if bornhardts were of such an origin they ought to be preserved on major divides. They are not, for some are exposed in valley-side slopes (Figure 3.11), others in valley floors (Figure 3.12).

Working in what was then Rhodesia, Mennell (1904) pointed out the importance of fracture systems in shaping granite landscapes and emphasised the effect of contrasted fracture density, with hills developed on massive compartments, whereas well-fractured rocks were weathered and reduced to plains. But he thought in terms of epigene or subaerial weathering. The two-stage mechanism, involving differential subsurface alteration, and comparable to the by then well-established theory of corestone evolution (Figure 3.4), but invoking compartments of rock rather than single blocks, was suggested by Falconer (1911). It found support, first, with the discovery of exposures demonstrating contrasted fracture density between hill and plain (Figure 3.13; Blès, 1986; Twidale, 1987). Second, dome-shaped masses of fresh granite already shaped beneath a regolithic cover were described from a quarry in Cameroon, in West Africa (Boyé and Fritsch, 1973). Later, similar examples of nascent domes, covered by rotted granite *in situ*, showing that they have not been shaped by epigene processes and then buried, were noted in southern Africa and in Australia, where they occur in natural exposures and in road cuttings, as well as quarries (Figures 3.11 and 3.14).

Figure 3.10. Bornhardt formed on a stock of granite intrusive into schist in central Namibia.

Figure 3.11. Nascent bornhardts exposed in flanks of a remnant of the African Surface, near Witrivier, Mpumalanga Province, South Africa. The residual hill is the Bushmans Kop.

Figure 3.12. Quarried crest of granite dome exposed in valley floor, near Malmesbury, Western Cape Province.

Figure 3.13. Shallow quarry with kaolinised rock and corestones indicative of closely spaced fractures located in valley between low bornhardts in massive granite, near Garies, Northern Cape Province.

Figure 3.14. Platform revealed as crest of dome in road-cutting, Midrand, Gauteng Province, South Africa.

Many bornhardts stand in isolation (e.g. Figures 3.1 and 3.6), but others occur in ordered rows, as is to be expected of forms associated with fracture patterns (e.g. Figure 3.2). Also, some few bornhardts are developed on synformal structures, indicating deep erosion and topographic/structural inversion (Lamego, 1938; Twidale *et al*., 1996). Thus, bornhardts have developed for a variety of reasons: they are convergent forms.

Why the rock compartments that became bornhardts were resistant to weathering and erosion can be explained by invoking the same horizontal shearing responsible for the orthogonal fracture systems that have given rise to corestone boulders. Similar shearing of compartments of blocks induces compression along one axis and tension along another aligned normal to the first (Weissenberg, 1947). Compressive stress could account for the resistance of certain granitic compartments, for compression implies tight partings that are virtually impenetrable to water, and as a corollary, little weathering and erosion. This is consistent with what is found in the field, for bornhardts invariably are formed on massive compartments of rock (Figure 3.15).

Figure 3.15. *Rillen* outlined by algal stains on flank of bornhardt developed in massive granite (few open fractures), southern Angola. Note boulder-strewn debris slope at base of sheer rock face.

5.3 Second-stage development

The agency responsible for the second stage of development involving the exposure of the topographically differentiated residual compartments, and including bornhardts, has varied. In some areas wind-driven waves and other marine agencies have uncovered bornhardts. In high latitudes the Laurentian and Fennoscandian cratons have been scoured by ice-sheets, and previously developed regoliths have been stripped to reveal resistant rock masses as inselbergs (e.g. Fogelberg, 1985). But in the low and middle latitudes, as in parts of Africa, South America, and Australia, rivers have been dominant and scarp recession arguably has been the most common mode of exposure of the resistant compartments. Uplift due either to deep crustal events or isostatic responses to erosional offloading, influenced landscape developments by inducing river rejuvenation and the stripping of regolithic veneers. The valleys incised by such streams were bounded by slopes that develop from below (e.g. Penck, 1924; Bremer, 1983; Twidale and Milnes, 1983), because water gravitates to lower slopes and the piedmont, where in consequence, weathering and erosion are most pronounced. This caused the undermining and steepening, regrading and recession of hill slopes, particularly those capped by a duricrust or a resistant stratum, and as appreciated by Holmes (1918), King (1942; 1953), and Fair

(1947; 1948). Thus, though not the reason for the formation of inselbergs, slope retreat probably played a significant role in their exposure.

5.4 Morphological variations

Domical bornhardts are the most common form of granite inselberg (Twidale, 1981b). Variants are derived as a result of the weathering and modification of the basic form.

Figure 3.16. Castle koppie in the Mrewe-Macheke area of central Zimbabwe.

Castle koppies or *kopjes*, known in Britain as tors and in France as *inselbergs de poches* or pocket inselbergs, appear to develop where the crest of a dome is exposed, but intense subsurface weathering of the still-covered flanks continued, converting the gentle slope to a precipitous bluff (Figure 3.16**)**. A similar origin may be proposed for what appears to be the unusually large koppie from the Erongo Mountains of central Namibia (Figure 3.8). It may be attributed to first, the very deep weathering of a dome, the crest of which was exposed, and second, the exploitation of steeply inclined fractures to produce the precipitous flanks.

Nubbins result from the weathering of the outer few shells or sheet structures to produce block- and boulder-strewn hills. In semi-arid Namaqualand and the Limpopo Province of South Africa, most bornhardts carry a scatter of blocks and boulders (Figure 3.7), indicating that some superficial sheet structures have disintegrated, though not enough to warrant them being called nubbins, which are best and most commonly developed in the humid tropics, present or past. They occur also in favourable moist sites, as in the Swakop River Valley of Namibia, where bornhardts occupy the adjacent higher ground. Yet, nubbins are well developed in arid central Namibia (Figure 3.17) mostly in granite but in the Keepmanshoek area in dolerite. Such occurrences could be inherited from humid climates of the past for such conditions obtained in this region during the later Oligocene and early Miocene (e.g. Axelrod and Raven, 1978; Spönemann and Hagedorn, 2000; see also Oberlander, 1972).

Figure 3.17. Nubbin in vicinity of the *Grosse Spitzkoppe*, central Namibia.

5.5 Sheet fractures and structures

Sheet fractures or offloading joints are characteristic of bornhardts (Figures 3.2 and 3.3). Their origin has long been the subject of debate. Several of the suggested explanations clearly are in error. For instance, sheet fractures extend to depths too great to be affected by temperature changes, whether diurnal, seasonal, or secular. Water-driven chemical weathering takes place along pre-existing sheet fractures and is not the reason for their development. Other explanations may apply only in particular instances. For example, the stresses generated late during the emplacement of granite stocks may be pertinent in that circumstance (e.g. Selby, 1982), but are not germane to the understanding of sheet fractures in arenaceous and rudaceous sediments, for example (Bradley, 1963; Bourne and Twidale, 2003; Twidale, 2010).

Two concepts have found favour as of general application. The offloading hypothesis proposed by Gilbert (1904) is plausible for so-called offloading joints are well developed in granitic bodies, which were emplaced deep in the Earth's crust, so that their present exposure implies the erosion of a thick overburden. It is no wonder that the hypothesis was widely accepted for many years (e.g. Ollier and Tuddenham, 1961; Selby, 1977; Sweet and Crick, 1992), for lithostatic loading is an important and ubiquitous factor in landform development (Chapman, 1956). However, the release of such stresses by erosion cannot of itself explain sheet fractures. The hypothesis implies that the form of the land surface (in this instance the domical morphology of the bornhardts) influences the geometry of the resultant fractures and that topography determines sheet fractures.

There are many inconsistencies between theory and reality. For instance, at some sites sheet fractures predate the associated landforms (Twidale, 1971; Gerrard 1974). The development of banks of sheeting slabs on recently eroded cirque walls and incised valley sides, previously touted as compelling evidence of erosional offloading (Lewis, 1954; Kieslinger, 1960; Kiersch, 1964; Gage, 1966) are best explained as due to the readjustment of stress fields and pressure release in parallelism with the fresh surfaces of least principal stress (Müller, 1964). Some bornhardts are developed in rocks with synformal or basin-shaped sets of sheet fractures, which is impossible to explain in terms of erosional unloading, but which is comprehensible by reference to deep erosion and topographic inversion (Twidale *et al.*, 1996).

Sheet fractures have also been explained as associated with shear and torsional stress (Harris, 1888; Merrill, 1897; Dale, 1923) and Gilbert also subscribed to this tectonic interpretation (see Dale, 1923:29). Sheeting fractures demonstrably are planes of dislocation as evidenced by imbrication, slickensides, fault steps, differential movement on opposed faces, and, in conglomerates, sheared cobbles (Twidale *et al.*, 1996; Bourne and Twidale, 2003). Sets of sheet structures are developed within orthogonal compartments that are themselves due to shearing. Several minor forms associated with bornhardts and with sheet structures are known to be generated by such dislocations (e.g. Twidale and Sved, 1978; Twidale and Bourne, 2000; 2009). Sheet structures are confined within compartments of rock defined by steeply dipping components of orthogonal fracture systems. They are readily construed as a response to recurrent shearing along the pre-existing shears. This is consistent with the distribution of sheet fractures within some bornhardts, with the most prominent developments on opposite flanks.

Sheet fractures are caused by applied compressive stress, the resultant strains being manifested as fractures only at shallow depth where dislocation is not inhibited by lithostatic loading. This is confirmed by the contemporary development of sheet fractures (e.g. Twidale and Bourne, 2000). The suggestion that they are tectonic is consistent with the postulated most common origin of bornhardts. It accounts for sheet fractures developed on the various rock types, igneous, sedimentary, and metamorphic, from which they have been reported (e.g. Dale, 1923; Bradley, 1963; Lageat, 1989; Campbell and Twidale, 1991; Twidale, 2010).

The domical form of bornhardts is usually associated with the upward convex geometry of the ubiquitous sheet fractures. However, just as the rounding of boulders is due to preferential weathering of corners and edges, so the domical form of residuals can be attributed to the differential weathering of the compartments on which they are based. At some few sites the domical form is maintained despite the residual being developed on granite subdivided by synformal sets of sheet fractures. This suggests that though the exploitation of sheet fractures by weathering agencies undoubtedly has occurred, preferential weathering is all-important in the shaping of the residuals.

5.6 Stepped morphology and episodic exposure

Given their primal development beneath planation surfaces of low relief it is not surprising that major fields of bornhardts are associated with multicyclic landscapes (Dixey, 1942; King, 1962; Partridge and Maud, 1987; Birkenhauer, 1991; Partridge, 1998). Thus, the bornhardts of the Witrivier District of eastern Mpumalanga Province, South Africa, are exposed beneath a remnant of the Early-Mid-Tertiary African Surface (Figure 3.11) and in the Valley of the Thousand Hills the Umgeni River has cut into the same Surface. Most of the inselbergs of Zimbabwe are located on plains resulting from the dissection of this African Surface also (Lister, 1987). In southeastern Namibia, however, bornhardts of early Cambrian or Precambrian age are exhumed and continue to be re-exposed from beneath a sedimentary cover of Cambrian (Nama) age (Du Toit, 1937; Twidale and Maud, 2012).

A two-stage development may imply different ages of subsurface weathering and erosional exposure, though some bornhardts may have evolved continuously, with exposure keeping pace with weathering below the surface. However, several lines of evidence and argument point to many, perhaps most, bornhardts having emerged episodically as a consequence of multiple alternations of weathering and erosion. Pauses during which subsurface scarp-foot weathering occurred are indicated by breaks of slope, platforms, alcoves and flared slopes at various levels on the residuals (Twidale and Bourne, 1975a; Twidale, 1982; Bourne and Twidale, 2000).

Such episodic exposure and the implied local increase in relief amplitude goes some way to explaining why some bornhardts have persisted in the landscape and are of great age. Hills shed water reducing the possibility of weathering (and hence erosion). The adjacent plains or valleys, on the other hand, receive an excess of water and consequently are weathered. Base-level permitting, they are susceptible to erosion, giving rise to piedmont depressions. Alternatively or additionally, volume decrease resulting from chemical weathering and the flushing of fines consequent on heavy rains and runoff may cause compaction and hence surface subsidence independently of stream behaviour (Ruxton, 1958; Trendall, 1962).

Figure 3.18. Stepped inselberg near Polokwane, Limpopo Province, with dome rising from lower large-radius dome. Granite is exposed in a platform and also in plain in foreground. Note *Rillen* or gutters on steep slope.

In southern Africa episodic exposure has resulted in stepped, or two-tiered or dome-on-dome residuals of which there are many examples in the Polokwane district of the Limpopo Province, in Namaqualand, and in Namibia (Figures 3.18 and 3.19), as well as in Bornhardt's classical Tanzanian landscapes. Particularly prolonged and/or intense subsurface weathering may be invoked in explanation of such two-tiered residuals as the *Grosse Spitzkoppe*, which is indeed claimed to be the largest of its kind in Africa (Migon, 2010). It stands about 600 m above the surrounding plains and consists of a dome with prominent sheet structure surmounted by a bouldery tower (Figure 3.19). Initially, it could have been a massive dome, the summit of which was exposed in an ancient planation surface, such as King's (1962) Gondwanan Surface of Cretaceous age (see also Ollier, 1978; Partridge and Maud, 1987; Partridge, 1998). In terms of episodic exposure it was shaped by subsurface weathering to a depth of say 200 m causing the upper part of the dome to be reduced to a rather stubby bouldery tower. The tower was then exposed by stream erosion, and thus preserved because it was no longer in continuous contact with water. The lower sector was subjected to further or renewed subsurface weathering possibly beneath the earlier Cenozoic African Surface before being revealed in later Cenozoic times.

Figure 3.19. The *Grosse Spitzkoppe*, central Namibia, showing upper tower on domical base.

Stepped bornhardts are widely developed in southern Africa. They have been exposed in stages and have increased in relief amplitude episodically and over time (Jessen, 1936; King, 1949; Twidale and Bourne, 1975a; Bourne and Twidale, 2000; Twidale, 2010). Once formed uplands become self-enhancing and self-perpetuating. Structural advantage leads to reinforcement and to unequal erosion, a sequence of events that has been called concatenation (Twidale, 2007a; see also Behrmann, 1919; Bliss Knopf, 1924; Crickmay, 1976; King, 1970; Twidale *et al.,* 1974; and Brunsden, 1993**).**

An appreciation of such episodic exposure not only explains the great age of some bornhardts but also provides a defence and rebuttal to an adverse criticism of the two-stage concept (Twidale and Bourne, 1975a). King (1966) advocated an origin involving scarp retreat, and pointed out that in any given area the maximum thickness of regolith is less than the height of the local bornhardts, a situation which appears at first sight to be incompatible with two-stage development. However, the argument is negated by the field evidence, both in southern Africa and in Australia where many bornhardts are stepped.

5.7 Contrasts between two-stage bornhardts and corestone boulders

One obvious difference between boulders and bornhardts is one of scale for rather than being developed on an individual block, similar to a corestone, bornhardts are formed on compartments comprising many blocks but with few open fractures. In addition, whereas corestones are weathered all round and thus are detached from the host rock mass, bornhardts appear to remain in physical continuity with, and to be projections from, the main granite body (see Brajnikov 1953). There is no evidence to the contrary for no bornhardt has yet been found that is underlain by weathered granite. Indeed, many granitic residuals resembling those of the Polokwane area clearly extend laterally into rock platforms (Figure 3.18) and the *Grosse Spitzkoppe* also appears to be firmly anchored (Figure 3.19; see also Twidale and Campbell, 1984). In addition, evidence concerning the origin of sheet fractures suggests that they remain rooted in the host mass: they are rounded masses of rock that initially projected into the base of the regolith and were later exposed. Were the bornhardt not physically contiguous with the host mass, applied compressive stresses would not have been transmitted and there would have been no low-angle faults or incipient sheet fractures.

6. Minor features

Many bornhardts demonstrably are etch forms, and some at least of the minor features that decorate the residuals are also of etch origin. Like the host corestone boulders and nascent bornhardts, they have been found already initiated on freshly exposed weathering fronts covered by weathered granite *in situ*, demonstrating that they were not formed on outcrops and then buried. Epigene forms are shaped exclusively after exposure, but some features are developed partly in the shallow subsurface and partly after exposure (e.g. Twidale and Bourne, 1975b). Yet, others are of tectonic origin.

6.1 Tectonic forms

No part of the Earth's crust is tectonically stable. Even the shield lands and cratons, though relatively inactive nevertheless experience seismic events, which disturb and disrupt the land surface and create specific forms. These include fault scarps, some a centimetre or less high, but forming patterns of horsts and graben, and also sheet fractures and structures (Vidal Romani and Twidale, 1998; Clark and Bodorkos, 2004). Vertical and lateral wedges are squeezed out, and slabs and blocks are displaced. Blisters, or arched thin slabs, are further squeezed and converted to A-tents or pop-ups (Twidale and Bourne, 2009). Such tectonic forms continue to be generated (Twidale and Bourne, 2000; 2003).

6.2 Epigene features

A few minor granite forms appear to be wholly of epigene origin. *Tafoni* are hollows or shelters developed in blocks, boulders, and sheet structures (Figure 3.20). The granite is broken down by salt weathering or haloclasty, i.e. the pressure exerted by crystal growth of such salts as halite, as they come out of solution (Winkler and Singer, 1972; Bradley *et al.*, 1978). They are initiated at the base of the host boulder or slab and grow upward and outwards as the result of either granular disintegration or the formation of books of laminae that are found on the interior walls and ceilings of *tafoni*. The laminae fall away under gravity so that the hollows increase in size. Such expansion frequently has caused the enclosing walls or visors to be breached, creating apertures or windows, which in Italian are *tafoni* (singular *tafone*). Many are still forming as, for example, on the granitic coast of the Cape Peninsula.

Haloclasty is responsible for the development of *tafoni* but how salts enter the system remains problematic. On the coast spray carries it on to rock surfaces and inland salt also is carried on the wind not only from the coast but from the surfaces of salinas. In such areas and given time, salts in solution washed out of the sky in rain may have percolated through a rock that is notionally impermeable, yet includes enough tiny fissures and voids to allow the gradual transmission of liquids.

Runoff and overflows from rock basins, particularly those with a vegetated soil fill, erode gutters or *Rillen* that score the slopes of blocks and bornhardts (e.g. Figures 3.15 and 3.18). Some extend a short distance along the weathering front at the base of slopes but many are due to dissection of exposed surfaces. At some sites gutters are inverted and have become ribs as a result of the protection afforded by blue-green algae. As the plants photosynthesise and must therefore have light, both the protective veneer and inverted ribs must be of epigene origin. In places, runoff has scoured the grooves so deeply that the laminar zone has been breached, exposing fresh rock in vertical lenticular depressions that look like button-holes.

A split boulder is one that has separated into two more-or-less equal parts along a clean planar break. Such boulders are not commonplace but neither are they rare. A few blocks and slabs are split by tree roots, but the division of others has variously been attributed to insolation, or to earth tremors. However, gravity is most likely responsible. Many boulders stand on flat surfaces. If the boulder includes a vertical fracture or zone of strain the unsupported hemispheres may have caused tension and eventual separation. Gravity has also caused the slippage of slabs and boulders undermined by basal sapping of slopes or disturbed by earth movement (see Bain, 1926; and Twidale *et al*., 1999).

Figure 3.20. Sheet *tafoni* with, in foreground, granite slope with surficial plates, flakes, scales or laminae, central Namibia.

6.3 Etch or two-stage forms

Newly exposed granite surfaces display pitting, or rough surfaces caused by the preferential weathering by soil moisture of the mica and feldspar, leaving quartz crystals in micro-relief. Where streamlets debouch from granite domes on to the regolith-covered piedmont plain the underlying bedrock surface not only becomes pitted, but the excess available water causes the surface to be lowered, forming shallow channels that are continuations of the gutters on the exposed domes. They do not continue far into the subsurface presumably because the rough surface causes dispersion of flow (as well as turbulence that assists erosion) but at some sites they extend far enough for gutters to converge in the natural subsurface.

Most newly exposed granite platforms display hollows or shallow saucer-like depressions. Many are located along and at the intersections of fractures, though some are due to the exploitation of pods of weak minerals such as mica or feldspar. These saucers are incipient rock basins (or *gnammas*). From time to time they retain water, which not only causes them to be enlarged but also differentiated after exposure (Twidale and Corbin, 1963). The surficial zones of granite are commonly laminated to a depth of several centimetres and saucers in such bedrock extend laterally faster than they penetrate vertically. This has resulted in flat-floored and relatively shallow depressions which Wentworth (1944) termed *pans* (Figure 3.21). Such features have, however, also been referred to as weathering pits (Goudie and

Migon, 1997; Migon, 2010). Where the laminated zone has been stripped away exposing homogenous rock, hemispherical pits are formed. Saucers on slopes (ca. 20º or steeper) become asymmetrical basins or armchair-shaped hollows. A pit may be deepened so much that it penetrates through the uppermost sheet structure, allowing water to plunge through into the sheet fracture, forming a cylindrical hollow.

Figure 3.21. Pans on the crest of Paarlberg, Western Cape Province.

In addition, major fractures are exploited at the weathering front resulting in fracture-controlled clefts or *Kluftkarr*en (Figure 3.22). Zones of strain (as opposed to ruptured rock) also are exploited, because strained crystals are particularly susceptible to moisture attack (e.g. Nabarro, 1967).

Sills of intrusive rocks (quartz, aplite, pegmatite, dolerite) also give rise to shallow linear depressions or ridges, depending on their resistance to moisture attack relative to the host rock. The morphology associated with one pegmatite sill exposed near Usakos, in central Namibia, changes laterally along strike, giving rise to a shallow depression where intrusive into granite but forming a low wall where injected into schist.

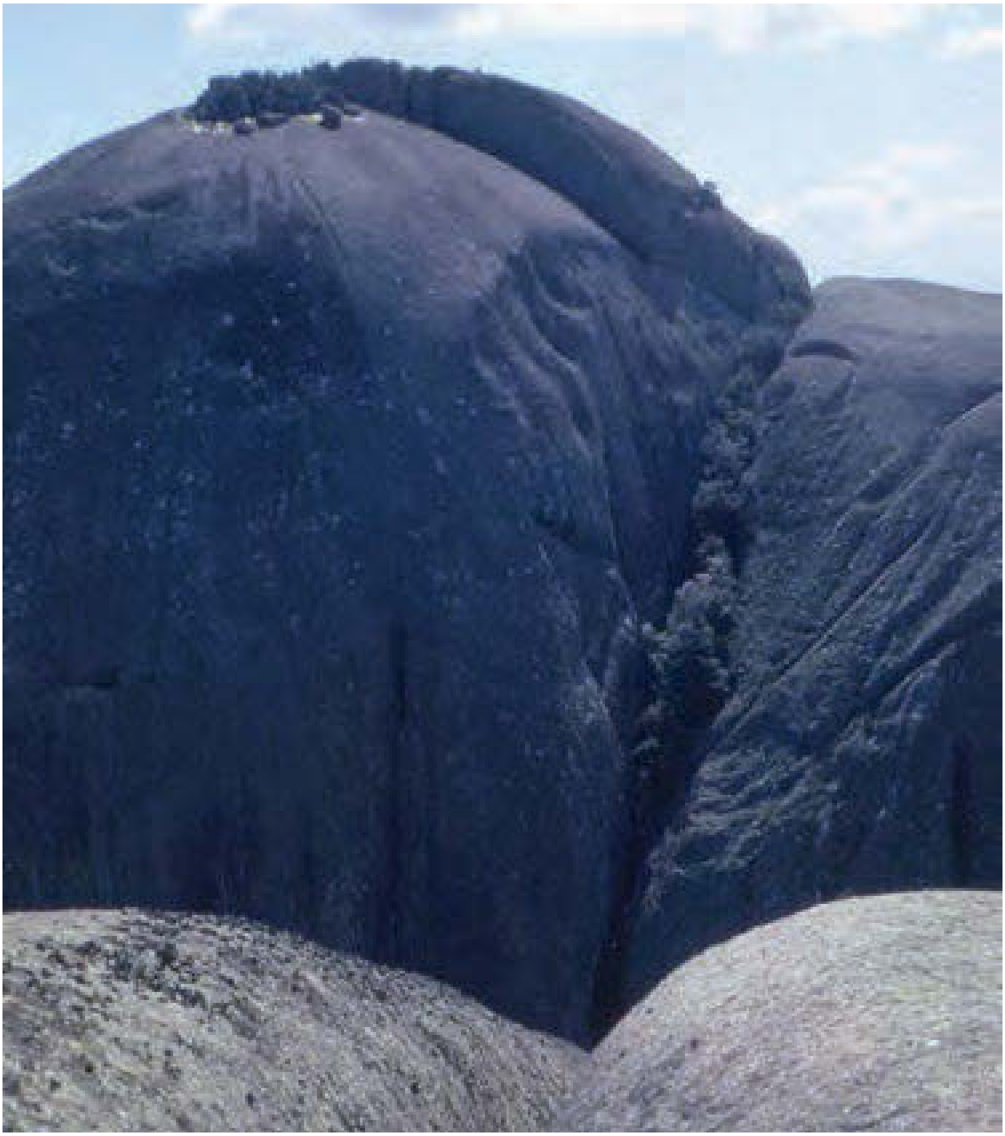

Figure 3.22. *Kluftkarren* or fracture-controlled clefts on Paarlberg, Western Cape Province. Note deep *Rillen* on far dome, and also sheet structure formed on crest of dome.

The bare granite domes and boulders shed water so that the surrounding regolith is frequently saturated. This causes intense and relatively rapid weathering so that the basal rock in contact with soil becomes fretted. In massive granite an annular depression has formed in the bedrock surface around the bases of outcrops (e.g. Clayton, 1956). Moreover, the moisture held in the regolith weathers laterally as well as vertically into the base of the bedrock, forming concavities. After exposure these become flared forms, examples of which have been found already shaped in the shallow subsurface (Twidale, 1962; Figure 3.23). At some sites intense or long-continued scarp-foot weathering has led to flared sidewalls becoming more deeply undercut to form alcoves and cylindrical shelters or footcaves.

Figure 3.23. Flared boulder, Bury Hills, Banket, northern Zimbabwe.

Just as runoff from hills has given rise to basal fretting and flared slopes, so runoff from boulders and blocks creates a zone of intense moisture attack around their bases. In some areas this has resulted in minor depressions. Elsewhere, the protection afforded by the blocks or boulders has led to the formation of plinths and perched or balanced boulders (Figure 3.24).

Figure 3.24. Pedestal rocks with balanced boulders, Domboshawa, central Zimbabwe.

6.4 Part subsurface, part epigene forms

An explanation involving partial etching like that invoked in explanation of castle koppies also offers an explanation of the rock levees bordering gutters in granite that are so well developed on Domboshawa, a notable bornhardt located some 30 km north of Harare in central Zimbabwe (Scott, 1967; Lister, 1973). Immediately adjacent to the gutters, soil water retained in the regolith drains into the channels. This causes weathering of the underlying bedrock to be slower than that beneath the adjacent undrained regolith. This leads to the formation of rims bordering the channel. The rim of the rock levee may even protrude above the regolith and thus be weathered even more slowly so that it becomes more and more elevated relative to the adjacent platform (Twidale, 1993), for the covered area continues to be weathered and worn down while the exposed surface is dry and relatively stable. Thus partial etching has produced the rock levees, though a coating of opaline silica derived from overflow (Whitlow and Shakesby, 1988) may enhance the formation and preservation of the marginal rims. The rock doughnuts (annular rims) and fonts (*Taufbecken* or *benetiers*) found on the coast, as well as inland, and in sandstone as well as granite, can be explained in similar terms (e.g. Coudé-Gaussen, 1981; Twidale, 1993; Twidale and Campbell, 1998).

6.5 Commonplace but enigmatic laminae and plates

It will be recalled that an early stage of granite weathering produces books of laminae. Thin plates are found on exposed corestone boulders and on the slopes of bornhardts (Figures 3.5 and 3.20), where incised gutters reveal books up to 10 cm thick. These plates, a common feature of exposed surfaces might feasibly be explained as desiccated remnants of subsurface laminae, caused by water-related weathering processes or as a reflection of pressure release, as has been proposed in explanation of granular disintegration (Bain, 1931).

Early workers attributed many common granite features to insolation, to heating and cooling, expansion and contraction, and surficial plates and slabs were construed as of a similar origin. Granular disintegration was attributed to contrasted coefficients of heating and expansion of rock-forming minerals of different colours. Certainly the intense but ephemeral heat of bushfires causes flaking, as do the temperatures generated in nuclear explosions (Watanabe *et al*., 1954) but such changes are limited to the surface and the top few millimetres of rock, whereas lamination, for instance, is found at depths several metres below the surface and thus beyond the influence of temperature fluctuations, whether diurnal, annual, or secular.

In addition, the freezing of water produces crystals of ice that in a confined space exert expansive pressure (gelifraction), though this has been questioned, with the shattering instead attributed to hydration (e.g. White, 1976), resulting in lamination and pseudobedding or flaggy structure. As previously stated, the growth of salt crystals (haloclasty) also is sufficient to cause the disintegration of granite or the production of laminae or flakes. Water-related weathering processes active in the subsurface cause alteration of the rock-forming minerals and the formation of laminae. This is a common occurrence. Even in hyper-arid regions, which however receive occasional rains and also dew developed on cold nights, such chemical alteration acts more rapidly than insolation (Barton, 1916).

Weathering causes the release of chemicals that are circulated in solution in groundwater. In arid and semi-arid lands, because water is evaporated at and near the surface, minerals precipitate out and remain near the land surface. Also, the action of lichens and mosses on exposed granite surfaces forms crusts of iron, silica and manganese salts and are known as case hardening (e.g. Figure 3.25).

Figure 3.25. Boulders with remnants of case-hardened crust near Augrabies Falls, Northern Cape Province. Some of the crust has fallen away, revealing the rotted granite beneath.

The weathering front, of which corestone surfaces are discrete parts, commonly displays a zone rich in iron oxides, sourced in groundwater. Such ferruginous zones occur also at or near the base of regoliths, though in places they are developed a few millimetres or centimetres within the fresh bedrock and beneath the weathering front (Twidale and Vidal Romani, 2002). In places a ferruginous discoloration persists as the circular reddish former outlines of corestones even where the host kernels have been rotted.

At some sites the case-hardened granite surfaces are subdivided into polygonal plates up to three centimetres thick. Some plates are of coherent granite but in some, especially those rich in ferruginous oxides, the granite is rotted and friable. Some, tending to ordered orthogonal shapes, and resembling blocks of chocolate, are formed on what obviously are fractures and may plausibly be attributed to shearing. The polygonal patterns, however, occur on corestone boulders shaped by weathering, showing that the cracking is not a primary petrological feature. During or following exposure the iron-rich zones originating around corestones or on other weathering fronts became desiccated. On exposure they contracted and cracked, causing some of the individual plates to arch slightly, as is observed in the field, but the origin and provenance of some polygonal cracking remains uncertain.

Thus, these commonplace minor features reveal many doubts, uncertainties, and contradictions, partly because many laminae, slabs and splinters are of convergent type and have formed in different ways. Like several other *minor* features, however, their possible origin influences the interpretation of the hosts, or major granite landforms.

7. Conclusion

The study of granite landscapes highlights the importance of several features and factors that are of significance to general theory. The Earth's crust is not passive but is still tectonically active, and this is reflected even on the supposedly rigid and stable shields and cratons. Tectonic forms continue to develop in brittle granite. Crustal stresses have created the fractures, the pattern and condition of which are vital to any understanding of granite forms and landscapes.

For this reason and because shallow groundwater are ubiquitous many granite forms are azonal. Bornhardts, for example, are found in southern Africa in the Namib Desert, the moderate (winter rainfall) Western Cape, and sub-humid KwaZulu-Natal. Boulders and rock basins are found wherever granite is exposed. However, nubbins are favoured by humid tropical conditions and *tafoni* by aridity.

Once standing in positive relief, granite uplands shed rain and runoff to the adjacent plains, and as a result of such unequal weathering and erosion tend to be resistant and remain upstanding. Some are sufficiently stable and unchanging to form refuges in which are found representatives of ancient plant species (e.g. Hopper *et al.*, 1996) and some of the oldest forms and surfaces yet recognised are preserved in granitic rocks (e.g. Twidale, 2007b).

Thus, understanding granitic landforms offers possible explanations for the many landscapes developed on petrologically different but physically similar and widely exposed rocks that occur in many parts of southern Africa. Indeed, the analysis of granite landforms and landscapes demonstrates not only the significance of structure, and particularly fracture geometry and density, but also subsurface etching. In addition, the flow-on effects of these two factors are clear, for the exploitation of structural weaknesses frequently induces positive feedback or reinforcement mechanisms resulting in the unequal distribution of weathering and erosion. Together with the relative stability of granite in dry sites and environments, such concatenation accounts for the persistence of granite forms, major and minor, through time.

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4

Weathering

Weathering

Paul D Sumner, Kevin J Hall, K Ian Meiklejohn and Werner Nel

1. Introduction

In the first detailed review of weathering studies in southern Africa (Hall, 1988a) the most significant achievement identified was the recognition of climatic control on dolerite weathering across the sub-continent (Weinert, 1961; 1965; Figure 4.1), which largely reflects a diversity of precipitation and moisture conditions. Other research focused on the use of weathering products in palaeoclimatic interpretation and weathering properties in rock mass strength assessments. However, Hall (1988a) noted a lack of field data reflecting microclimatic conditions that could be used to establish weathering processes and called for collection of base-line measurements of rock properties, rock moisture content and chemistry. Laboratory simulations based on field data were identified as a further priority to isolate active weathering processes. Criticisms were also directed at the use of weathering products as indicative of processes, particularly with respect to mechanical breakdown where process-product relationships are unclear. In closing Hall (1988a:25) notes that "in many ways, weathering studies have yet to begin in southern Africa and there is a wealth of information waiting to be gathered."

Figure 4.1. Deep weathering on dolerite exposed in a road cutting in the west-central Free State Province, South Africa.

Here we review the subsequent weathering studies in southern Africa. While there have been advances on several fronts that directly address the issues noted above, such as in field monitoring of the contemporary weathering environment and in palaeoclimatic interpretation, there have also been several new areas where weathering research has been undertaken. These include laboratory testing and simulations, weathering studies in the sub-Antarctic and in application of weathering to the deterioration of southern African rock art. By necessity the geographical focus in this chapter lies in the eastern mountains of southern Africa, where much of the research has been undertaken, but we include studies from the Western Cape mountains, Namibia and on Marion Island in the sub-Antarctic. Emphasis in the review is on surface weathering phenomena and their application to landforms and rock art, while soil formation and karst landforms are not included.

2. Weathering studies

A summary of weathering studies in southern Africa and on sub-Antarctic Marion Island is presented in Table 4.1. Three general themes emerge: contemporary field monitoring of the weathering microclimate; weathering in association with landforms, notably in palaeoenvironmental interpretation; and laboratory studies. Sub-Antarctic Marion Island, annexed by South Africa in 1948, provides a stark contrast to the environment of the African sub-Continent and studies there span the first two themes.

Table 4.1. Geomorphic studies that incorporate aspects of rock weathering in southern Africa since 1990 (studies specific to the weathering of San rock art are listed in Table 4.2). Basalt refers to Drakensberg Group basalts unless specified as MI (Marion Island basalt).

AUTHOR(S)	TYPE OF PAPER OR STUDY LOCATION	ROCK TYPE(S)	PRIMARY FOCUS
Cooks and Otto (1990)	SEM analysis Samples from Magaliesberg	Quartzite	Weathering effects of lichen
Hall (1991)	Discussion paper with SA focus	General	Freeze-thaw and geocryological studies in SA
Hall (1992)	Discussion paper with SA focus	General	Cryogenic studies in SA incorporating weathering
Boelhouwers (1991)	Landform study Western Cape mountains	Quartzite, shale	Periglacial processes and landforms incorporating weathering
Boelhouwers (1993)	Discussion paper Western Cape mountains	Quartzite, shale	Weathering and relict cryoclastic debris production
Meiklejohn (1994)	Landform study Drakensberg and Lesotho Highlands	Basalt	Valley asymmetry resulting from periglacial processes
Wessels and Wessels (1995)	Field experiment Golden Gate, Eastern Free State Province, SA	Sandstone	Strain analysis and mechanical weathering by lichens
Wessels *et al*. (1995)	Field experiment Golden Gate, Eastern Free State Province, SA	Sandstone	Strain analysis and moisture uptake by lichens
Hall and Hall (1996)	Laboratory study on SA rock types	Sandstone, dolerite	Experimental wetting and drying

AUTHOR(S)	TYPE OF PAPER OR STUDY LOCATION	ROCK TYPE(S)	PRIMARY FOCUS
Goudie and Migón (1997)	Landform study Central Namib Desert	Granite	Weathering pits
Goudie *et al*. (1997)	Process study (salt weathering) Central Namib Desert	Limestone	Salt weathering rates
Boelhouwers (1999)	Landform study Hex River Mountains	Quartzite	Interpretation of relict periglacial slope deposits
Boelhouwers *et al*. (1999)	Cederberg Mountains	Quartzite	Relative-age dating openwork deposits incorporating weathering aspects
Grab (1999)	Landform study Lesotho Highlands	Basalt	Interpretation of relic deposits
Migón and Goudie (2000)	Landform study Central Namib	Granite	Landforms and associated weathering attributes
Boelhouwers *et al*. (2002)	Landform study Lesotho Highlands	Basalt	Interpretation of relict blockstream
Sumner and De Villiers (2002)	Landform study Amatola Mountains, Eastern Cape Province, SA	Dolerite	Interpretation of relict screes
Sumner and Nel (2002)	Laboratory study on SA and Marion Island rock types	Basalt, MI basalt, dolerite, sandstone, quartsite	Rock strength reduction on moisture application
Sumner *et al*. (2002)	Landform Study Marion Island	MI basalt	Relative age dating rock surfaces incorporating aspects of weathering
Boelhouwers (2003)	Review and landscape model Lesotho Highlands	MI basalt	Quaternary slope development and associated weathering
Boelhouwers and Sumner (2003)	Landform study Western Cape Province, SA, Lesotho Namibia, Karoo	Quartzite, basalt, dolerite	Significance of relict blockfields and blockstreams
Boelhouwers *et al*. (2003)	Field monitoring Marion Island	Basalt	Rock surface temperatures, rock moisture monitoring
Nel *et al*. (2003)	Landform study Marion Island	MI basalt	Mechanical disintegration and scree production
Sumner (2004a)	Field monitoring Marion Island	MI basalt	Mass loss from small clasts
Sumner (2004b)	Landform study Lesotho Highlands	Basalt	Interpretation of relict block accumulations
Sumner and Meiklejohn (2004)	Landform study Marion Island	MI basalt	Interpretation of autochthonous blockfields
Sumner *et al*. (2004)	Field monitoring and review: Namibia, Drakensberg, Marion Island, Antarctica	Granite gneiss, basalt (SA and MI), sandstone (Antarctica)	Rock surface temperatures

AUTHOR(S)	TYPE OF PAPER OR STUDY LOCATION	ROCK TYPE(S)	PRIMARY FOCUS
Grab and Mulder (2005)	Landform study Lesotho Highlands	Basalt	Bedrock terrace formation
Viles (2005)	Field monitoring Central Namib Desert	Granite, marble	Rock surface temperatures, surface moisture, microclimate, surface deterioration using SEM
Sumner and Nel (2006)	Field monitoring Drakensberg	Basalt, Sandstone	Rock surface temperature monitoring
Grab (2007a)	Field monitoring High Drakensberg	Basalt	Rock temperature monitoring
Grab (2007b)	Field monitoring High Drakensberg	Basalt	Rock surface temperature monitoring
Hedding *et al*. (2007)	Landform study Marion Island	Basalt	Relative-age dating of pronival rampart debris incorporating aspects of weathering.
McKechnie *et al*. (2007)	SEM analysis Samples from Mpumalanga Province, SA	Quartzite	Mechanical weathering by lichens
Sumner *et al*. (2007)	Field monitoring Southern Namibia	Granite gneiss	Rock surface temperature monitoring
Viles and Goudie (2007)	Field monitoring Coastal Namib Desert	Marble, granite (and Bath Stone)	Salt weathering and rock breakdown rates
Boelhouwers *et al*. (2008)	Review of periglacial processes on Marion Island	MI basalt	Includes rock temperatures, moisture conditions
Sumner and Loubser (2008)	Laboratory study on SA rock types	Sandstones	Experimental wetting and drying
Edwards (2009)	MA dissertation on a wetland in the KwaZulu-Natal Midlands	Dolerite	Chemical weathering of dolerites and the development of wetlands
Sumner *et al*. (2009)	Review Drakensberg-Lesotho	Basalt, sandstone	Basalt and sandstone weathering and rock art deterioration
Grab and Svensen (2011)	Landform study Golden Gate	Sandstone	Rock doughnut and pothole structures
Viles and Goudie (2012)	Review Central Namib Desert	Various Namib rock types	Review of weathering controls, processes and associated landforms

2.1 Contemporary weathering environment

Southern Africa has a wide range of climates and rock types. Such diversity was recognised early and Weinert (1965) summarised the effects of climate on weathering across the sub-continent. While rock properties can play a fundamental role in weathering, important drivers monitored typically include microclimates, rock temperature and rock moisture, as opposed to general climatic conditions. Several studies have attempted to provide such data in southern Africa. The most detailed data stem from studies on the sandstones where rock art is found and these are described below. Attention has also been directed toward monitoring microclimatic conditions in the eastern mountains and the desert in the west. Although the studies are few, given the range of potential microclimatic conditions and the diverse geology across the sub-continent, they may provide a broad regional comparison.

In the eastern mountains of Lesotho and the uKahlamba-Drakensberg rock temperatures have been measured in the basalts above and below the escarpment. At 1 920 m, at the lower altitudinal limit of the basalts, Sumner and Nel (2006) recorded a minimum rock surface temperature of –6.7 °C and a maximum of 56.0 °C; ranges which far exceed those found for air and the soil surface. Although no rock moisture data were collected, Sumner and Nel (2006) suggest a potential for frequent wetting and drying cycles based on summer rainfall patterns, but they noted that the low occurrence of precipitation in winter probably precludes effective frost action. Above the escarpment Grab (2007a, b) recorded basalt temperatures at a variety of locations and at different rock depths. Rock surface temperatures dropped as low as –13.6 °C and ranged up to 44.3 °C, but varied substantially with topographic setting, notably due to aspect. Again, no rock moisture data were presented but Grab (2007a) suggests that cryogenic and thermal stress weathering are currently active.

Salt weathering is suggested to be highly effective on the Namib Desert coast (Goudie *et al.*, 1997) and later studies provide interesting environmental data from the arid west for contrast with the mountains in the east of southern Africa (e.g. Sumner *et al.*, 2009). Viles (2005) collected a range of microclimatic data from four sites in the central Namib Desert where granite and marble samples were monitored in the field. Rock temperatures reached 61.5 °C in the summer and dropped to –0.3 °C in the winter. Although complex temporal patterns of temperature were noted, the four sites showed very similar overall trends and Viles (2005) suggests some degree of spatial homogeneity in the area. Moisture conditions, monitored with surface wetness sensors, were found to vary between sites and linked to fog frequency. Over a two-year period of exposure, blocks of cut granite showed no evidence for surface weathering although marble samples were affected. In southern Namibia, near Aus, Sumner *et al.* (2007) measured temperatures of different surfaces of a block of granite gneiss. Temperatures reached 62.8 °C in summer and dropped to –1.6 °C in winter. A high potential for thermal stress fatigue was noted, supported by observations of fracture patterns on boulders. Short-interval surface temperature data confirmed frequent oscillations exceeding 1 °C/min and greater than 2 °C/min on cooling. Similarly, Meiklejohn *et al.* (2009) found rock temperature oscillations in Clarens Formation sandstones to exceed 2 °C/min.

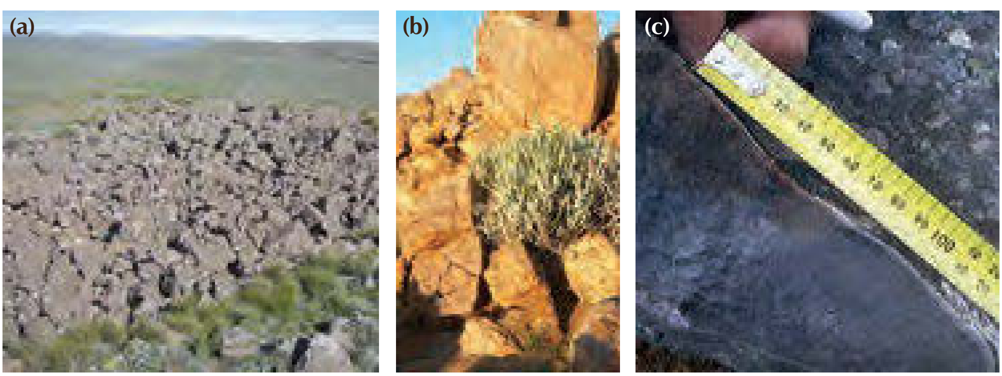

Figure 4.2. (a) Relict blockfield in eastern Lesotho, blocks typically measure 0.5-1.0 m long-axis size and are considered a product of weathering and slope emplacement under a colder climate; (b) 0.1 to 0.2 m in diameter polygonal fracture patterns on granite gneiss in southern Namibia, note encroaching afternoon shadow; (c) thin chemically weathered rind on fractured Marion Island basalt where weathering rates are considered slow.

Although it is interesting to compare the eastern mountains and western desert of southern Africa, few regional conclusions on the weathering environment can be drawn by comparing the published field data. While the studies in themselves are useful, and provide detailed analyses of sites, the overall area is vast and monitoring undertaken to date is spatially and temporally limited. There are also site-specific attributes that influence outcomes, such as rock type, local topography, aspect and sensor setting, as well as instrumentation differences that make inter-site comparisons difficult, as does any attempt to draw regional inferences. Overall temperature ranges are perhaps not as markedly different from east to west as anticipated (see also Sumner *et al.*, 2004), except for the data from above 3 000 m. What is still apparent though is the general absence of actual rock moisture monitoring thus confining any interpretation to *potential* processes rather than more clearly defined conclusions. Also, few process-landform conclusions can be drawn from the field data presented.

2.2 Weathering and landforms

A few landform studies ascribe direct or indirect associations with weathering processes. Rock doughnut and pothole structures are described in the sandstones at Golden Gate (Grab and Svensen, 2011) and granite weathering pits in central Namibia (Goudie and Migón, 1997). Detailed descriptions of landforms and weathering in Namibia are given by Migón and Goudie (2000), as well as Viles and Goudie (2012); the latter notable for cross-scale connections between weathering and landscape evolution. Boelhouwers (1991; 1993) describes contemporary weathering processes associated with periglacial activity in the Western Cape mountains, however, much of the research linking weathering and landforms has been undertaken on relict landforms and associated palaeoenvironmental interpretation.

Surface weathering is used in relative-age dating of debris flows in the Cederberg (Boelhouwers *et al.*, 1999). Boelhouwers (1993; 1999) also described the association of weathering patterns and the formation of relict block accumulations, including blockstreams at altitudes above approximately 1 600 m in the Hex River Mountains in the Western Cape Province of South Africa. Mechanical fractures superimposed on Neogene chemical weathering forms are described, which suggest a period of enhanced mechanical weathering and block production during colder periods of the Late Pleistocene, with subsequent block accumulations mobilised under periglacial conditions. In the Eastern Cape mountains, Sumner and De Villiers (2002) also argue for a former period dominated by mechanical weathering, rather than chemical weathering, which contributed to the formation of the (now relict) dolerite screes above 1 550 m, first reported on by Marker (1986). It appears likely that the Eastern and Western Cape Provinces of South Africa, accumulations are associated with the colder conditions around the Last Glacial.

Similarly, a former mechanical weathering-dominant phase post-dating a chemical weathering period is proposed for the eastern Lesotho mountains. Above approximately 3 000 m, blocky slope and valley-floor accumulations in the form of relict blockfields and blockstreams are superimposed over coarse-matrix colluvium (Grab, 1999; Boelhouwers *et al.*, 2002). Such blocky accumulations are ubiquitous, but predominate on south-facing slopes (Sumner, 2004b) that are typically steeper than the north-facing slopes. The slope asymmetry is in part explained by enhanced chemical weathering on the north-facing slopes with a former periglacial overprint on the cooler opposite slopes where sediments are preserved (Meiklejohn, 1994; Boelhouwers, 2003). Grab *et al.* (2005) also show aspect-related differential weathering patterns on scarps. Mechanical breakdown of basalt scarps and the mobilisation of blocks over and within slope mantles during a colder periglacial period (Boelhouwers *et al.*, 2002; Sumner, 2004b) follow the mode of development described for the Hex River Mountains. Periglacial landforms are described elsewhere in this book, but based in part on weathering attributes it appears that there is sufficient evidence for linking the Late Pleistocene mechanical weathering of the Cape to that of the highlands of Lesotho. A non-periglacial origin of blockfields caused by *in situ* weathering followed by the washing out of interstitial fines is also described for sites in Namibia and the Karoo (see Boelhouwers and Sumner, 2003).

Figure 4.3. Tafoni weathering patterns in Clarens Formation sandstone, Golden Gate Highlands National Park, Free State Province.

2.3 Laboratory studies

Durability testing of dolerites basalts and sandstones has been undertaken for engineering purposes (see below), but laboratory-based studies in southern Africa that confront geomorphological issues directly remain scarce. Sumner and Nel (2002) tested the effect of rock moisture on the hardness, and by implication the strength, of different rock types from southern Africa and Marion Island. A decrease in rock strength, in particular for the sandstone, was noted under high rock moisture contents. In a study on San rock art weathering, field thermal profiles were replicated in climatic cabinets to establish changes in porosity and microporosity of sandstone clasts (Meiklejohn, 1995; Meiklejohn *et al.*, 2009) and showed that over time porosity increased and microporosity decreased. The enlargement of pores and bedding planes was noted to be, in part, responsible for the deterioration of the paintings

Other laboratory testing focused on the effects of cyclical wetting and drying on rock properties and consequent mass loss. Several South African rock types have been tested using different moisture applications (Hall and Hall, 1996; Sumner and Loubser, 2008; Loubser, 2010) and show both mass loss and physical changes induced by moisture variability under stable temperatures. In a study related to salt weathering, Viles and Goudie (2007) tested Namibian rock samples in the laboratory for mass loss, soluble salt content and rock strength. A scanning electron microscope (SEM) was used to determining surface effects. In lichen studies the SEM has also proved valuable in determining rock surface lichen morphometry and associated geomorphic imprints (Wessels and Schoeman, 1988; Cooks and Otto, 1990; McKechnie *et al.*, 2007).

2.4 Weathering on sub-Antarctic Marion Island

The weathering environment on the island is characterised by frequent rock wetting and drying cycles (Boelhouwers *et al.*, 2003) associated with the passage of rain-bearing warm and cold fronts. Drying cycles can be driven simply by wind, as opposed to insolation, under cloudy conditions. Rock temperatures seldom drop below 0 °C at sea level and ranges generally follow that of air temperature due to the enduring cloud cover. At higher altitudes rock temperatures fluctuate more, probably due to higher insolation receipts and lower air temperatures (Boelhouwers *et al.*, 2003) with rates of temperature change occasionally exceeding 2 °C/min at 1 000 m a.s.l. (Sumner *et al.*, 2004). A direct association between the weathering environment and the mechanically-derived products evident on the island has, however, yet to be derived. Weathering rates of exposed surfaces are largely unknown and complicated by the cyclicity of glaciation and deglaciation in the past. Much of the bedrock fracturing and block production is probably due to dilatation on deglaciation and fractures tend to follow cooling joint structures and bonded discontinuities in the lava flows (see Boelhowers *et al.*, 2003). Although scoria fragments can lose up 1.2% mass annually (Sumner, 2004a), post-glacial bedrock surfaces still exhibit striations and this points to slow rates of basalt surface recession, at least spanning the Holocene. Lichens are ubiquitous on the island but their weathering effect has yet to be studied.

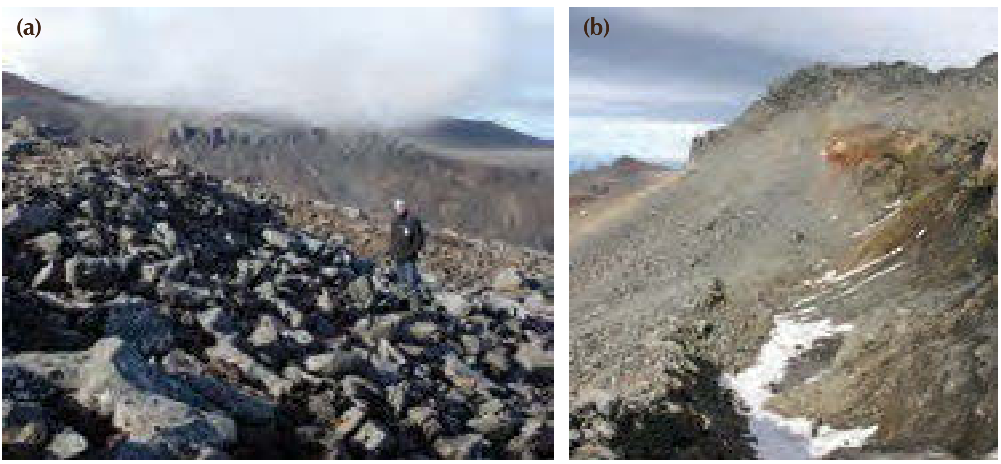

Figure 4.4. (a) Blocky accumulation of basalt *grey lava* on sub-Antarctic Marion Island with basalt screes off the Feldmark Plateau in background and red scoria cone, Beret Kop, in upper left; (b) Pronival rampart on Marion Island with minimal summer snow infill of the trough (note person lower centre for scale): relative-age dating of the downslope (distal) deposits incorporated aspects of surface weathering (Hedding *et al.*, 2007).

Weathering-associated landforms include blockfields and screes. Autochthonous blockfields are found below 750 m a.s.l. but consist mostly of a single layer of disintegrated bedrock with little chemically derived product present. Sumner and Meiklejohn (2004) conclude that weathering rates are slow under the island's periglacial climate, and that block production was probably limited to pressure release immediately following deglaciation. Screes are widespread below 700 m on the island where steep slopes exhibit mechanical disintegration and subsequent block production (Nel *et al.*, 2003) although this is likely related to glacial over-steepening or faulting on deglaciation. Hedding *et al.* (2007) report on a pronival rampart at 900 m a.s.l. where late-lying snow underlies a structurally-derived rock scarp. Relative-age dating of the rampart, using lichen and moss growth, rock surface hardness, clast angularity and weathering rind thickness reveal a retrogressive development mode that contrasts with the normal mode of rampart formation. Relative-age dating using the above weathering parameters was earlier successfully implemented on a range of basalt glacial and periglacial landforms, including screes, moraines, and stone banked lobes (Sumner *et al.*, 2002). While a temporal resolution could be resolved that spans the post-glacial time period, weathering rates appear nonetheless to be slow.

3. Applied weathering

It is somewhat enlightening that a search of Google Scholar® on applied weathering studies in South Africa primarily returns texts that pre-date 1988. There are, of course, exceptions, and not all of these are distinctly geomorphological. For example, Laubscher (1990) discusses the impact of weathering on a geomechanics classification system, noting particularly the impact of chemical weathering on decreasing the strength of Kimberlites. A recent approach is the application of geochemical information to reconstruction of palaeoenvironments and to relict sediment sourcing. For example, Dunajko and Bateman (2010) used, amongst other analyses, weathering attributes to discern provenance of sediments in the Wilderness Barrier Dunes. Young *et al.* (1998) utilise geochemical evidence, or the lack thereof, to identify the environment to help validate the Earth's oldest reported glaciation (~2.9 Ga), while Retallack *et al.* (2003) adopt a similar approach, using weathering evidence, to identify the Permian-Triassic vertebrate extinction boundary. Retallack *et al.* (2003:1133) were able to suggest a shift from an arid and highly seasonal Permian palaeoclimate to that of a (Triassic) warmer and wetter environment as shown by an increase in chemical weathering. In much the same way, Grandstaff *et al.* (1986) used weathering zones between pre-Cambrian sedimentary rocks to help reconstruct the environment at 2.8-3.0 Ga ago, including evidence for past partial pressures of oxygen and carbon dioxide gases in the palaeo-atmosphere. This is an exciting and novel application of weathering, and one that can be expected to be used more in the future.

Figure 4.5. Lesotho Highlands Water Project monument at the Ash River outfall: a concrete lining with coarse dolerite aggregate was used to stabilise the sandstone and basalt transfer tunnels that weathered rapidly on exposure after drilling.

Other geotechnical considerations, notably on durability of rock, incorporate aspects of weathering. Bell and Jermy (2000) identified *rapid weathering dolerites*, as used for construction purposes, through strength and durability testing. Haskins (2006) developed weathering indices for weathered granites underlying the Injaka Dam construction and relates these to engineering performance. However, much attention has been focussed on the lithologies underlying with the Lesotho Highlands Water Project (LHWP): the Drakensberg basalts and the underlying sandstones, specifically the Clarens Formation. Geotechnical investigations found that the unconfined compressive strength of the Clarens Formation varied appreciably, from moderately strong to extremely strong (Castro and Bell, 1995) and proved highly resistant to laboratory durability testing (Van Rooy and Van Schalkwyk, 1993). Durability of the basalt was tested extensively (Van Rooy, 1992; Van Rooy and Van Schalkwyk, 1993; Haskins and Bell, 1995), but rapid weathering on exposure of the basalts during construction, as well as the complexity of jointing, led to an increase in the foundation depth of the Katse Dam and the lining of underground water transfer tunnels (Bell and Haskins, 1997). Sumner *et al.* (2009) summarise the nature of weathering of the basalts and sandstone in the LHWP.

3.1 Weathering of cultural stone

Recent research directed towards the conservation of indigenous rock art has provided the most substantial and advanced work on weathering in southern Africa during the past twenty years. The San people inhabited the interior of southern Africa for at least 25 000 years (until as recently as 1870 in the Drakensberg) and the sandstones of the Clarens Formation provided a *canvas* to paint images that reflected both the San's history and their spirituality (Meiklejohn *et al.*, 2009). This heritage is globally acknowledged and was one of the primary reasons for the uKahlamba-Drakensberg Park being declared a World Heritage Site (UNESCO, 2000). Deterioration of the paintings, the age of which is mostly unknown, has long been recognised and could adversely affect the international status of the region (Hall *et al.*, 2007a). Nearly two decades of research has focused on understanding the processes responsible for deterioration, primarily through weathering, in an attempt to preserve the art for future generations.

Research into the deterioration of rock art has had several points of foci (Table 4.2) that relate directly to weathering at rock art sites (e.g. Meiklejohn *et al.*, 2009) or to the sandstone in general (e.g. Mol and Viles, 2010), microclimatic monitoring (as detailed below) and long-term deterioration assessments. Examples of the latter include the comparison of photographic evidence, which show an indication of change over the historical record (Ward and Maggs, 1994; Hœrlé, 2005; Leuta, 2009), including granular surface disintegration and flaking due to weathering. From an understanding of weathering, however, this approach has limitations as processes are not possible to discern, rates are unlikely to be linear and environmental variations have probably occurred through time (see Sumner *et al.*, 2009). Thus, detailed long-term data regarding environmental conditions on rock surfaces within the shelters is critical.

Table 4.2. Research on rock art weathering in the Drakensberg since 1990 including MA and PhD theses.

AUTHOR(S)	PRIMARY LOCATION(S)	OUTPUT-TYPE AND FOCUS	DATA/MEASUREMENTS
Ward and Maggs (1994)	Main Caves, Giant's Castle Game Reserve	Journal article: comparison of deterioration from photographs	Visual comparison of deterioration based on early photographs
Meiklejohn (1995)	Main Caves and Battle Cave, Giant's	PhD: weathering processes associated with two rock shelters	Microclimate, rock temperatures, rock moisture fluctuations, rock properties

AUTHOR(S)	PRIMARY LOCATION(S)	OUTPUT-TYPE AND FOCUS	DATA/MEASUREMENTS
Meiklejohn (1997)	Main Caves and Battle Cave, Giant's	Journal article: rock moisture and weathering	Chemical weathering and moisture movement
Ward (1997)	General	Journal article	Review; Drakensberg
Hoerle and Salmon (2004)	Game Pass Shelter, Kamberg	Journal article: Microclimate and weathering	Shelter microclimate
Hoerle (2005)	Game Pass Shelter	Journal article: weathering activity	Environmental conditions, archival documents, observations
Hoerle (2006)	Game Pass Shelter	Journal article: rock temperatures and processes	Rock temperatures simulation measurements
Hoerle *et al.* (2007)	Game Pass Shelter	Journal article: rock strength assessment	Ground-penetrating radar on shelter backwall
Hall *et al.* (2007a)	General	Journal article	Discussion paper on new findings and new challenges
Hall *et al.* (2007b)	Main Caves	Journal article: surface temperatures.	Thermal responses of rock art pigments
Arocena *et al.* (2008)	Giant's Castle sample	Journal article: pigment mineral composition	X-ray diffraction
Prinsloo *et al.* (2008)	Giant's Castle sample	Journal article: pigment and salt mineral composition	Raman spectroscopy
Hall (2009)	General	Journal article	Discussion paper on weathering and climate change
Meiklejohn *et al.* (2009)	Main Caves and Battle Cave	Journal article: Microclimate and weathering	Microclimates, rock temperatures, rock moisture, rock chemistry and rock properties
Leuta (2009)	Main Caves and Battle Cave	MSc dissertation	Photographic comparison of deterioration based on photographs
Sumner *et al.* (2009)	General	Journal paper	Review, Drakensberg and Lesotho
Loubser (2010)	Samples from adjacent to Main Caves	MSc dissertation: experimental wetting and drying (see also Sumner and Loubser, 2008)	Laboratory analysis of cyclical wetting and drying including sandstone samples
Hall *et al.* (2010)	Main caves	Journal article: light penetration into sandstone	Pyranometer response beneath rock surface
Mol and Viles (2010)	Golden Gate	Journal article: moisture regime of sandstone	Electric resistivity tomography, protimeter and Equotip

Detailed field investigations that focused primarily, but not exclusively, on environmental monitoring at Main Caves and Battle Cave began in the early 1990's (Meiklejohn, 1995). Monitoring continued until recently at both caves (Meiklejohn *et al.*, 2009) and there has been a resurgence in interest over the past decade at these and other sites, such as Game Pass shelter (Hœrlé, 2004), as well as with laboratory testing (see Laboratory studies above). However, environmental monitoring only has value where, first, the data pertain specifically to the rock surface upon which the art is painted and second, where data span sufficient time periods to reflect overall conditions to which the art is subjected. Hœrlé and Salmon (2004) provided a short record of general climatic conditions in Game Pass Shelter that appears comparable to general weather records for the area. Subsequent rock surface temperatures at the same site, obtained from cut blocks of sandstone, show few cycles below 0 °C in winter and low rates of temperature change but, as these records are restricted to a six week period (Hœrlé, 2006), they are highly constrained temporally. Longer duration records that include surface and sub-surface rock temperatures and rock moisture conditions from the shelter surfaces are presented by Meiklejohn (1995) and Meiklejohn *et al.* (2009). Moisture and temperature changes at granular and pore-scale are suggested, through a suite of weathering processes, to enlarge pores and bedding planes near the rock surface thereby causing deterioration of the art (Meiklejohn *et al.*, 2009). Particularly noticeable is the associated impact of salt weathering, where the crystallisation of salts and possibly also their volumetric changes from temperature and moisture variations has resulted in the deterioration of the art (Meiklejohn, 1995).

New instrumentation and techniques have recently allowed for downscaling of surface monitoring. A key attribute defining the scale at which to assess processes is that the paint acts as a surface modifier. While rock surfaces adjacent to the paintings will weather according to *natural* physical, chemical and biological processes, the pigments alter the albedo, porosity, chemistry and thermal properties of the surface (Hall *et al.*, 2007b). Thus, there is a need to evaluate art weathering both at the environmental scale as well as at the resolution of pigment-surface interaction. Environmental data can still provide important information pertaining to changing conditions, particularly where the immediate environment is altered by providing access to tourists, or through vegetation fires that can impact on subsequent shading of rock surfaces. The real value in understanding processes, however, lies in detailed surface monitoring in direct association with the art. Thermal infrared data collected at Main Caves show that white and red pigments, as well as the rock surface, have different responses to solar radiation (Hall *et al.*, 2007a, b). This can lead to pigment-pigment and well as pigment-rock stresses under different thermal and moisture conditions allowing for fracturing and the ingress of microbial activity (Arocena *et al.*, 2008).

Interestingly, work elsewhere on Clarens Formation sandstone has shown (Büdel *et al.*, 2004) that cryptoendolithic cyanobacteria can cause flaking but, so far, no such evidence of weathering by biological agencies has been found at the rock art sites (see Arocena *et al.*, 2008). This (so far) absence of cryptoendolithic activity is intriguing for, where the sandstone is light transmissive, the light penetration facilitates colonisation by endolithic organisms (Hall *et al.*, 2010). Hence the questions arise: If there is no colonisation, why? Is there no colonisation, or is it simply that none has yet been found? If the latter, then it is equally clear that having examined many sites it is, nevertheless, not as common as might be supposed and thus the question still remains as to why this is so. Transmission of light into the sandstone also impacts thermal gradients and these differ between the art (where the pigments act as a barrier to light penetration) and adjacent to the paintings where light can penetrate freely. Such situations may lead to contrasting thermal stresses at the pigment boundaries. Understanding surface interactions is thus complex and is further compounded by the recent recognition that not all of the art is painted on naturally-weathered sandstone, but that some surfaces have been prepared by the artist by grinding or through the application of a clay base (Hall *et al.*, 2007a). This particular aspect, and its impact on weathering processes and rock art deterioration, requires further investigation.

Although in a local context a (relatively) substantial amount of research has been undertaken on rock art deterioration (Table 4.2; and readers are directed to recent summaries by Meiklejohn *et al.* (2009) and Sumner *et al.* (2009) for further detail) much still remains to be done. Hall *et al.* (2007a) and Sumner *et al.* (2009) note that this includes better determining paint composition, thickness and properties, ongoing monitoring to determine stresses at the pigment-pigment and pigment-rock interfaces, and how these factors impact on state of the art. Further, it is necessary to continue the investigation of environmental changes at rock art sites to better understand how weathering may have changed. Ultimately, we need to accept that weathering is inevitable, and that some deterioration is expected through time. However, further research will, hopefully, facilitate methodologies to arrest process rates, and thus help conserve the art for the appreciation of future generations.

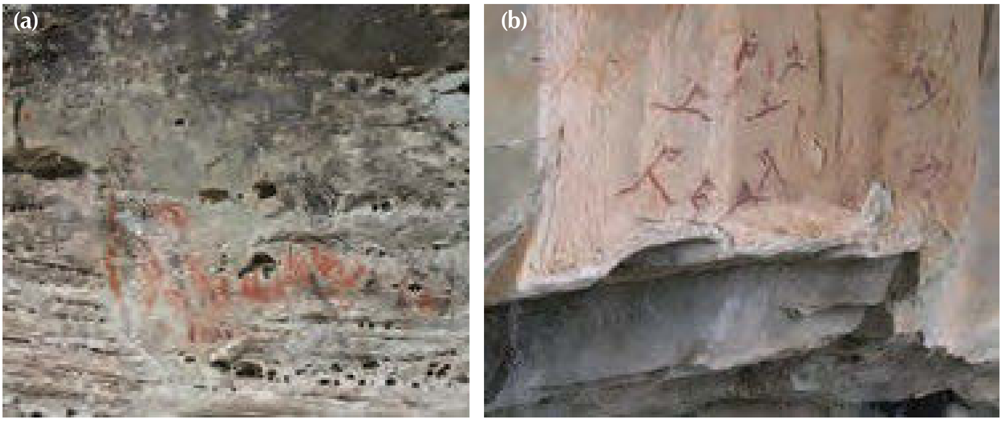

Figure 4.6. (a) Weathered San rock art on sandstone, note the preferential weathering along bedding planes adjacent to central smoothed surface; (b) Well preserved *Battle scene* painted onto gypsum precipitates possibly indicative of former wetter period.

4. Discussion

After almost a quarter of a century since the review of Hall (1988a) has there been any advance regarding weathering studies in southern Africa? While the answer must, of course, be *yes*, the question also arises as to what is the nature of those advances, what gaps remain, and, on the global stage, what is the impact of work from this region?

Perhaps central to weathering studies is the understanding of process and the outcome at a landform or landscape scale (Hall *et al.*, 2012). In this regard we still understand little in the local context that can contribute to the broader geomorphic community. Studies in Namibia (see Viles and Goudie, 2012) best attempt to bridge this scalar divide, but much emphasis in local studies relates either to microclimatic monitoring or to the interpretation of relict landforms associated with former cold climates without providing detailed or definitive process-landform linkages. An ongoing focus with the Last Glacial and its imprint in Lesotho, in particular, has directed much attention to the basalts and the relict landforms. Unfortunately, little is known of the contemporary weathering environment and the susceptibility of Drakensberg basalt to, say frost action, as a basis for extrapolation to weathering processes in the past. This continues to be a challenge, not only in the southern African context (Hall and Thorn, 2011), as these linkages remain tenuous, and weathering-landform associations still remain speculative some two decades later (see Hall, 1988b; 1991; 1992).

Several advances have been made in clarifying the contemporary weathering environment in southern Africa through rock surface and microclimatic monitoring. The region is vast and diverse with respect to topographic regimes and rock types and studies have focussed on the Drakensberg Mountains in the east and the Namib Desert in the west. Intriguingly, no data are forthcoming from rocky coastal areas. Temperature has been the primary focus in most studies but the similarity in thermal regimes between sites in the east and west of the sub-continent is remarkable (Sumner *et al.*, 2004). Few studies, however, report on rock moisture conditions (Meiklejohn, 1995; Mol and Viles, 2010), or use surrogates for surface wetness (e.g. Viles, 2005). Most data are also temporally limited; typically monitoring is shorter than three years, which poses limitations in understanding climatic fluctuations and the effect on rock weathering. In addition, still little is known on rates of weathering and, while some laboratory studies attempt to address this, relationships between weathering processes and rock properties is relatively understudied.

Studies on Marion Island by southern-African based geomorphologists over the last two decades have provided insights into the geomorphology of a distinct environmental setting. Few landmasses exist in the maritime sub-Antarctic region and the extreme ocean setting of Marion Island distinguishes the island as an active, wet periglacial environment with frequent frost cycles (Boelhouwers *et al.*, 2003). Frost cycles are also frequent in rock, particularly at high altitude and the abundance of rock wetting and drying cycles could, outside of the coastal zones of the World, possibly be unique. Notwithstanding the dynamic environmental setting, weathering rates appear to be slow for the basalt (although rapid in scoria) with thin weathering rinds and preserved pre-Holocene glacial striations on bedrock surfaces. The susceptibility of the basalt to breakdown by wetting and drying is also questioned in laboratory simulation (Loubser, 2010) while Sumner and Meiklejohn (2004) speculate on limited weathering and ascribe blockfield formation to dilatation immediately post the Last Glacial event. Thus, although apparently a very dynamic weathering environment, particularly for mechanical weathering, questions remain around the actual weathering process and their efficiency in such a setting. This also highlights the need for detailed information on the effect of rock properties as a control on weathering type and rate, a point that is emphasised below.

Figure 4.7. Rock art at Game Pass shelter, Drakensberg, painted on a sandstone rock surface that was pre-prepared by the artist by smoothing the surface before paint application.

Several new areas of focus also need attention in future studies. The role of biological weathering in southern Africa (see e.g. McKechnie *et al.*, 2007; Hall *et al.*, 2010) remains largely unknown, and may play a significant role particularly where environmental conditions change. The impacts of climate change on the weathering environment, both from a theoretical perspective and where applied to cultural stone conservation (Hall, 2009), are also open to speculation and may develop as an important international focus. A further consideration is that the downscaling of monitoring, such as in recent studies pertaining to rock art, creates an even greater challenge in attempts to upscale the contribution of weathering to landscapes. It is significant to note that where rock art surfaces in cliff overhangs may have remained relatively static over periods of thousands of years (i.e. apparently slow weathering rates), these are precisely the sites where weathering might be expected to be *enhanced* given that they are recessed from the adjacent cliff-lines. The role of surface weathering in this context begs to be explored and the southern African setting may provide such a platform.

Studies into the weathering associated with San art have arguably provided the most important contributions to our understanding of rock weathering. Environmental monitoring continued for 15 years (Meiklejohn *et al.*, 2009) thus for the first time providing a platform for incorporating changing environmental or climatic conditions. More recently, downscaling objectives facilitated by advances in monitoring equipment have allowed for finer and finer resolution monitoring, including light and moisture penetration and grain-scale thermal interactions of pigments and the rock surface (Hall *et al.*, 2007b; 2010). While this downscaling is inevitable as we pursue greater detail on the mechanisms of rock weathering there is an unavoidable consequence of disassociating further from what is surely the ultimate objective of weathering studies, that is to describe the landscape around us (Hall *et al.*, 2012). Another consequence of such a reductionist approach that is further compounded by laboratory process simulations is a continual dissociation with the underlying tenant of *weathering*: that the environmental conditions are assumed to play the dominant part in rock breakdown. In recognition of this, and partly as a consequence of the recent approach in rock art studies, Hall *et al.* (2012) challenge the traditional top-down approach and propose that rock type or the properties of the rock are dominant in determining process type and rate. The weathering (or rock decay) is argued as a function of rock-type interacting with the energy transferred to, or within, the rock. Although Hall *et al.* (2012) contrast with the traditional model of weathering, such a conceptual approach may be actively pursued in settings such as Marion Island where the dynamic environment does not appear to align with the observed weathering rates. The approach also shifts the emphasis away from purely environmental monitoring, which does not necessary explain process, rates and landform linkages, to the attributes of the rock itself thus addressing the shortcoming in this regard in southern African weathering studies.

5. Conclusion

Hall (1988a:24) noted in conclusion of the earlier review that "weathering studies in southern Africa are somewhat rare and … considering the diversity of the country, they are limited in scope." This has clearly changed, but perhaps not as much as one might have hoped. Advances are still somewhat constrained and focused in certain areas, with some perpetuation of the conservative thinking of old. Researchers are still few, but increasing international collaboration that also facilitates technological or methodological advances, with a view to confronting issues in weathering both in a local and broader context is a positive move forward. Clearly both the work on reconstructing ancient environments and the weathering of rock art have international relevance. Although both approaches have been specific to the South African situation, there is nothing that is not applicable elsewhere in terms of approaches and the findings are of universal, rather than just local, significance. The advances and enthusiasm of those focused on weathering suggest a continuing and significant growth but, echoing Hall (1988a:25), it still remains that "the potential for (new) work is almost boundless" and that "there is a wealth of information (still) waiting to be gathered."

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5

Fluvial Geomorphology

Fluvial Geomorphology

Kate M Rowntree

1. Introduction

Fluvial geomorphology is concerned with the study of landforms and landscapes produced primarily as the result of flowing water. Its scope is wide in both space and time. At the largest scale, fluvial geomorphologists are concerned with whole river systems and the landform that contains them, the drainage basin. As drainage basins dominate most continental landscapes, fluvial geomorphologists have a major role to play in explaining long-term evolution at the continental scale. Within a drainage basin, the network of river channels may take up only some 2% of the space of the basin, but they play a crucial role in transporting water, sediment, nutrients and dissolved chemicals through and out of the basin. Local deposition of the sediment that is carried by the river gives rise to numerous local landforms on the valley floor and within the channel itself. These may be at the scale of vast floodplains that form and persist over thousands of years or small sand dunes or pebble clusters on the channel bed that form, deform and reform with every flood that passes down the river.

Many fluvial geomorphologists have explored the links between time and space in river systems. Schumm and Lichty (1965) produced a seminal paper that examined how the relationship between spatially determined dependent and independent variables changes at different timescales. They presented a three-fold subdivision of time into cyclic time, graded time and steady state time (Figure 5.1). Cyclic time was equated to geological time periods over which landscape evolution takes place at the regional and drainage basin scale; this timescale can encompass cycles of erosion as proposed by Davis (1889) or King (1942). Schumm and Lichty (1965) equated cyclical time to geological time that could, for example, last for the length of the Pleistocene epoch, or longer. Time, initial relief and climate are the independent variables (the drivers), catchment and drainage network morphometry the dependent variables (the response). Graded time, or modern time (Schumm and Lichty, 1965) is the timescale over which a dynamic channel equilibrium can be achieved at the reach scale, with channel form being maintained by a balance between erosion and deposition within the reach over the timescale of decades to centuries. At this timescale the independent variables are catchment-scale variables that determine catchment hydrology and catchment sediment yield, while channel form is the dependent variable. Steady time refers to the short term or present day during which the system can be considered to be constant, with equilibrium between independent system variables of flow discharge, channel morphology and the dependent variables of flow hydraulics and sediment transport. This would be a timescale of one year or less (Schumm and Lichty, 1965).

These different timescales have influenced the approaches taken by fluvial geomorphologists. Two eminent fluvial geomorphologists, Slaymaker (2009) and Church (2010), have presented essays on their views of the historical development of geomorphology and its future trajectory. Their ideas are summarised here as a backdrop to developments in southern African research. As explained by Church (2010), prior to the 1950s geomorphologists were largely concerned with regional scale landscape evolution through cyclical time and took a historical, interpretive approach, based on qualitative observation and comparison of landscape features. After 1950, access to changing technologies (aerial photographs, improved survey techniques, the advent of dating techniques, satellite imagery and, more recently, vastly improved computing power) and widespread implementation of, for example, hydrometric monitoring was coincident with a shift to process response studies grounded in a paradigm of Newtonian mechanics and aligned with respect to graded or steady state time. Fluvial geomorphologists borrowed steady state mechanistic theory from hydraulic engineers to explain sediment transport in steady state time and channel form relationships in graded time using concepts such as dominant discharge (Church, 2010). The major paradigm was one that encompassed ideas of equilibrium, linearity and predictability.

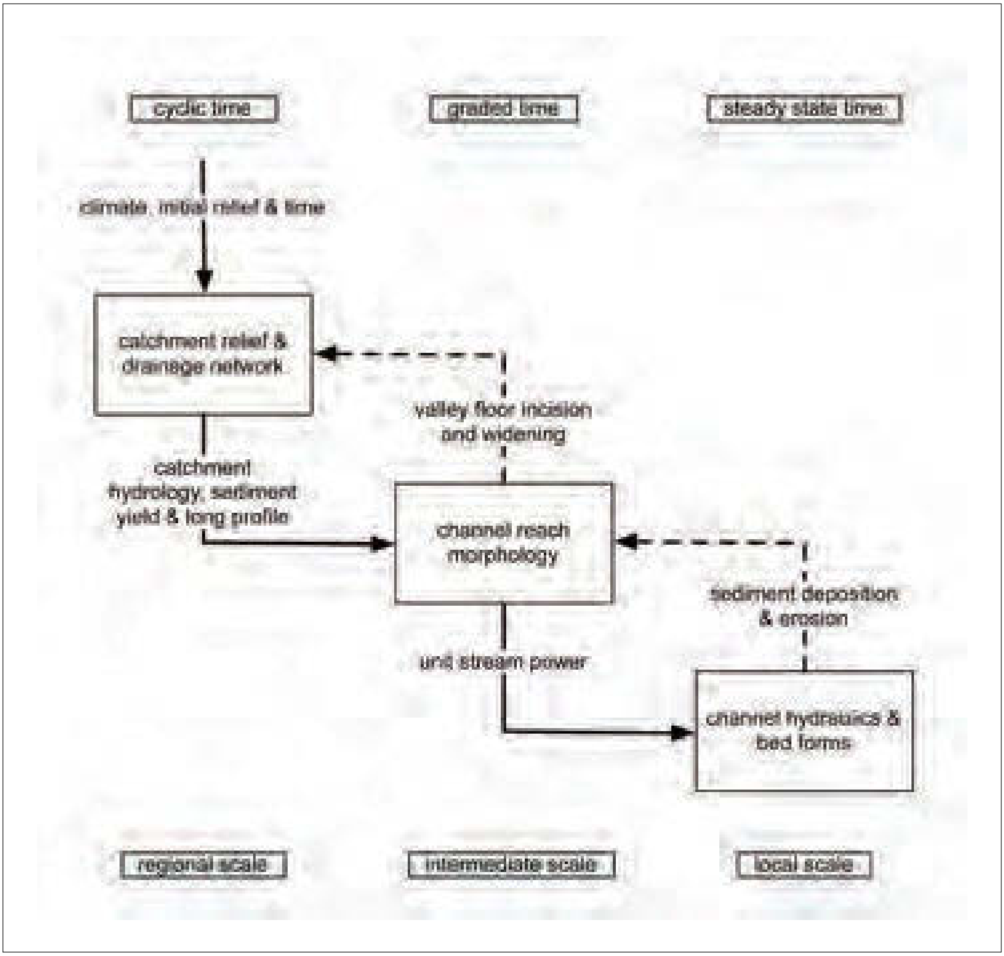

Figure 5.1. Time and space scales (after Schumm and Lichty, 1965). The solid arrows indicate independent variables acting through time and space, from higher order to lower order scales. The broken arrows indicate feedback in the form of morphological change working from lower order to higher order scales.

Unfortunately, the common experience was that although some generalisations can be made, rivers and their landforms are not predictable, especially as time and space scales increase. From the

1990s there has been a shift towards a paradigm that treats fluvial landscapes as complex systems characterised as much by disorder, irregularity, instability, unpredictability and non-linearity as they are by the inverse – order, stability, regularity, predictability and linearity (Slaymaker 2009; Church, 2010). Geomorphological systems exhibit self-regulation and develop emergent properties (Church, 2010). Church (2010: 274) promotes modern geomorphology as a system science where "System sciences are ones that seek explanation by integrating the effects of many elements and processes. Geomorphology is quintessentially a system science." Church (2010) identifies five important features of geomorphic systems: (1) the juxtaposition of different physical processes at the same place and time; (2) the importance of temporal and spatial scales; (3) the significance of sediment storage and, as a result, of prior history; (4) the occurrence of thresholds and limits that can lead to abrupt change; (5) the occurrence of variable system forcing over all timescales.

Both Slaymaker (2009) and Church (2010) point to a failure to fully integrate geomorphology with ecology and human agency, two major factors that determine the nature of our landscapes. The two are related because humans use and abuse ecological productivity, while ecological processes are an integral part of the landscape system. Given the increasingly significant impact of humans on environmental processes, Slaymaker (2009) argues for geomorphologists to adopt a critical realist approach in empirical observations and theory development; that is, an approach tempered by the realisation that science is not infallible and that human agency is based at least in part on value judgment, not rationality. This has important implications for river restoration projects in which the desire is to restore the natural system, but what is perceived as natural is based on value systems (Wohl and Merritts, 2007).

Fluvial geomorphology in South Africa has followed the trajectories outlined by Slaymaker (2009) and Church (2010), but the path has not been a continuous one. King (1942) presented a major work on the South African landscape that fits well within the historical interpretive paradigm of cyclic time. In his book on South African Scenery (King, 1942), and subsequent works (King 1953; 1962; 1976), he challenged Davis's (1889) cycle of erosion, promoting the idea of scarp retreat as being more relevant to landscapes such as are found in southern Africa. King's ideas are comprehensively summarised by Twidale (1992). Significant sedimentological work on palaeoriver systems was carried out by geologists interested in gold and diamond bearing alluvial gravels (as exemplified by Buck and Minter (1985), Marshall and Baxter-Brown (1995) and Spaggiari *et al.* (1999) archaeological or palaeoenvironmental studies have also contributed to our knowledge of Pleistocene sediment formations (see Butzer *et al.*, (1973) for the Vaal River in South Africa and Smith et al. (1993), Heine and Heine (2002), Bourke et al. (2003), Srivastava et al. (2006) for the Homeb Silts of the Kuiseb River in Namibia).

Previously, there was less interest in studying channel form-process relationships for their own sake. This began to change from the 1990s onwards for two reasons. The first was the opening up of South Africa to international scientists after the change in government in 1994. The second was the challenge posed to geomorphologists by ecologists, and new water legislation that called for the protection of water resources, including the river ecology and the river channels that provide habitat to river organisms. South African fluvial geomorphologists over the last twenty years have responded to the needs of river managers; international researchers have contributed to this call (e.g. Heritage *et al.*, 1997; Heritage *et al.*, 2001), but they have also focused on more fundamental research on longer-term landscape development (e.g. Tooth, 2004). Overseas researchers have not only brought their international experience, but have opened up access to laboratory facilities not available previously in South Africa. These include facilities for dating using Optical Stimulated Luminescence (OSL) (Bourke *et al.*, 2003; Rodnight *et al.*, 2005; Brook *et al.*, 2006), long-lived cosmogenic nuclides such as ^{10}Be, ^{26}Al, ^{3}He and ^{21}Ne (Granger and Muzikar, 2001; Kounov *et al.*, 2007) and short-lived radionuclides such as ^{137}Cs and ^{210}Pb (Humphries *et al.*, 2010; Rowntree and Foster, 2012; Foster *et al.*, in press). Laboratory facilities for sediment tracing include geochemical analysis of sediments (Manjoro *et al.*, 2012) and environmental magnetism (Rowntree and Foster, 2012; Foster *et al.*, 2012).

The following sections of this chapter are organised as follows. Section 2 presents a brief overview of the regional drivers of river systems within South Africa. The development of conceptual thinking around time and space scales, especially as conceived by southern African geomorphologists (Section 3) is then examined. Section 4 looks at the abiotic-biotic response as exemplified by riparian vegetation. Landscape connectivity is presented as a conceptual framework for understanding fluvial systems in Section 5. Geomorphologists are playing an increasing role in developing tools and guidelines for the management and conservation of South African rivers; some of these developments are reviewed in Section 6. Finally, gaps in our knowledge are identified and future research questions proposed.

2. Regional drivers of river systems

River character is determined by both regional and local factors. At a regional scale, catchment runoff and sediment yield provide the main inputs into the channel network while the channel longitudinal profile determines the distribution of potential energy available for erosion and transportation of sediment. At the local scale, the channel form adjusts to the downstream manifestation of these regional factors as well as to local constraints such as underlying geology and riparian vegetation. This section examines the translation of regional scale factors into the flow regime, sediment yield and downstream changes associated with the longitudinal profile within the southern African context.

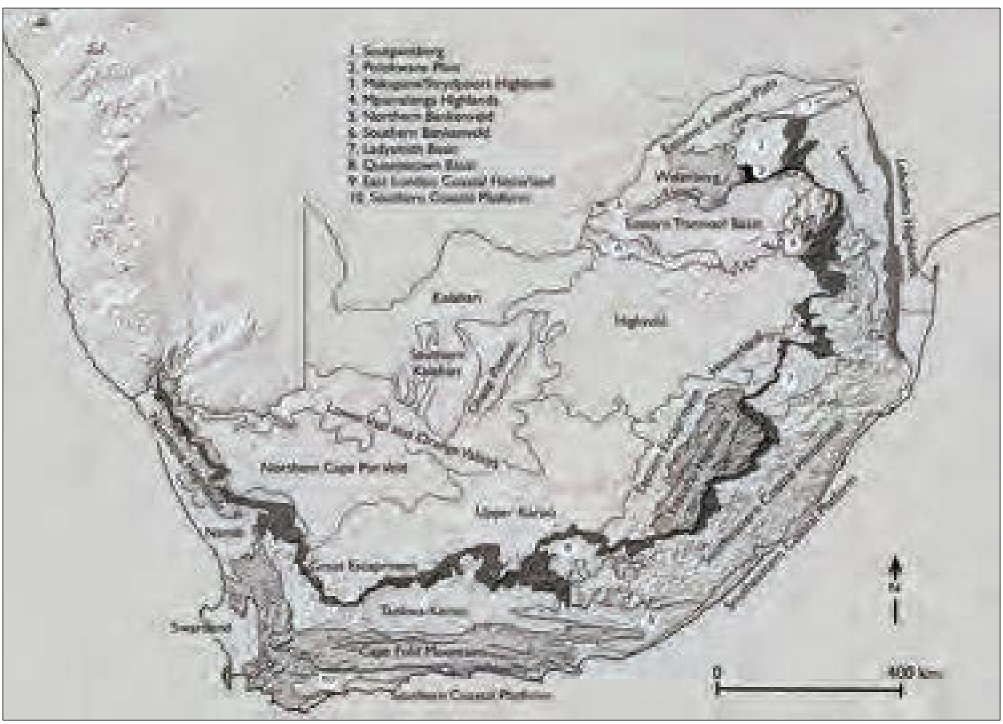

Figure 5.2. Geomorphological provinces (after Partridge *et al.*, 2010). Based on GIS cover provided by J. Moolman, Department of Water Affairs and digital elevation model from Natural Earth. Free vector and raster map data @ naturalearthdata.com.

The macro-geomorphology of the region is described by Partridge and Maud (1987) and Maud (this volume). Dollar (1998) provides a comprehensive review of the evolution of southern Africa's river systems. Other authors who have contributed to our understanding of the evolution of southern Africa over the geological timescale include Hattingh (1994), Hattingh and Goedhart (1997) and Hattingh and Rust (1999) at the local scale of the Sundays River in the Eastern Cape Province of South Africa, and Moore *et al.* (2007) for drainage systems of south-central Africa. Partridge *et al.* (2010) have defined and mapped South Africa's geomorphological provinces (Figure 5.2), where a province is a contiguous area with similar geomorphological characteristics in terms of terrain, altitude, annual rainfall characteristics and so on. Detailed descriptions of each of the 34 provinces and 12 sub-provinces are given by Partridge *et al.* (2010).

Although the provinces of Partridge *et al.* (2010) were developed only for South Africa, Lesotho and Swaziland, some extend into adjacent countries within southern Africa. The Highveld province continues north into Zimbabwe and Zambia, while the Great Escarpment extends into Namibia as a topographic feature, although rainfall decreases significantly. A distinctive feature of Mozambique not found in South Africa is the extensive coastal plain, which comprises the deltas of the modern and proto-Zambezi (Moore *et al.*, 2007). The southern end of the African Rift Valley extends through Malawi, Zimbabwe, northern Botswana and Namibia where it plays a key role in guiding the course of modern rivers (Figure 5.3).

Figure 5.3. High water levels during seasonal floods inundate the Zambezi floodplain in Caprivi. Floods such as this one are important ecosystem drivers. They provide important areas for fish spawning, away from the high velocity flows in the main channel, as well as recharging local groundwater aquifers that maintain dry season flow and provide water to the floodplain vegetation. Note the clear water in a river of which the bulk of its sediment load is sand.

Three types of river system have been identified by Tooth and Nansen (2011) in the southern African region: moderate sized, moderate gradient rivers that drain seaward from the Great Escarpment to the coast, or from parallel ranges such as the Cape Fold Mountains; larger, moderate gradient rivers associated with the Orange River system or Highveld rivers such as the upper Olifants River, and moderate to low gradient rivers of the endoreic Kalahari Basin (e.g. the Okavango). To this must be added the vast 1 390 000 km^2 basin of the Zambezi River.

The Zambezi River has been subdivided into three sections (Moore *et al*., 2007), the characteristics of which are summarised below. The upper reaches are incised into Precambrian basement rocks, but for most of its upper course the Zambezi is a low gradient (0.00024) river flowing over unconsolidated Kalahari sands, characterised by dambos at the head of tributary streams and wide floodplains along main channels. Moore *et al*. (2007) describe a 'dambo' as a shallow grassy valley with a persistently high water table. Dambos constitute important sponges that feed the tributary streams.

The river steepens below the Ngonye Falls as it incises into Karoo age basalts and sub-Kalahari bedrock, before falling some 100 m over the 1 700 m wide Victoria Falls. Downstream of the falls the river flows through a series of deep gorges, including the 100 km long Batoka Gorge, with an average gradient of 0.0026. These gorges provide dam sites for Kariba and Cahora Bassa. Between the Kariba and Cahora Bassa Gorges, the river flows through a wide floodplain between steep escarpments that hosts many of the region's wilderness areas.

The lower section of the river, below Cahora Bassa, forms a 100 km long floodplain delta system of oxbows, swamps and multichannel meanders. This alluvial section of the river is particularly vulnerable to the impact of upstream dams and river engineering works on the floodplain.

Table 5.1. Major river basins of southern Africa.

	LENGTH KM	BASIN AREA KM2	MEAN ANNUAL RUNOFF MILLION M3/A
Zambezi	2,650	1,390,000	94
Orange	2,300	721,000	11.5
Okavango	1,100	530,000	11
Limpopo	1,750	408,000	5.5
Kunene	1,050	106,500	5.5
Incomati	480	49,965	3.5
Source: **Limpopo River Awareness Kit (2010)**			

2.1 Flow regime

A summary of characteristics for the six largest river basins in southern Africa is given in Table 5.1. The mean annual runoff is a good measure of the magnitude of the river, but masks the variability inherent in the flow regime. Southern African rivers are known for their high variability both in terms of inter-annual variability of annual runoff and flood events (McMahon *et al*., 1987), an expression of the seasonal and inter-annual variability of rainfall. Tyson (1986) and Mason and Jury (1997) have identified an approximate 18-year rainfall cycle in the summer rainfall region that extends into Zimbabwe and Botswana. Shorter-term rainfall variability in the range of three to six years has been linked to El Niño Southern Oscillation – ENSO – that brings wet and dry periods, the wet years often being accompanied by extreme floods, the dry years by drought (Vogel, 2000). An additional 80-year cycle has been tentatively identified for flows in the Zambezi (Moore *et al.*, 2007).

Inter-annual variability of flow is inversely related to the mean annual runoff (McMahon *et al.*, 2007). Flow variability is further increased by high evaporation rates that exacerbate the high rainfall variability. Southern African rivers also exhibit a high seasonal variability. In South Africa the rainfall is highest in the summer rainfall region to the east and along the southern coast into the winter rainfall region of the Cape Peninsular and Cape Fold Mountains. The rest of the western half of South Africa, Namibia, and Botswana are semi-arid to arid. The Zambezi Basin is subject to tropical influences, with summer rainfall, and can be described as sub-humid to humid.

In the winter rainfall area of the Western Cape Province of South Africa, high winter flows contrast with extremely low summer flows (except where irrigation releases raise natural flow levels) whereas in the summer rainfall regions the inverse is true. Both the Okavango and Zambezi have a distinct seasonal lag between peak rainfall and peak discharge that increases down river. Moore *et al.* (2007) present data for the Zambezi demonstrating that at Chavumba (catchment area 75,500 km^2) discharge peaks in March whilst at Victoria Falls (catchment area 507 200 km^2) discharge peaks in April (Figure 5.4).

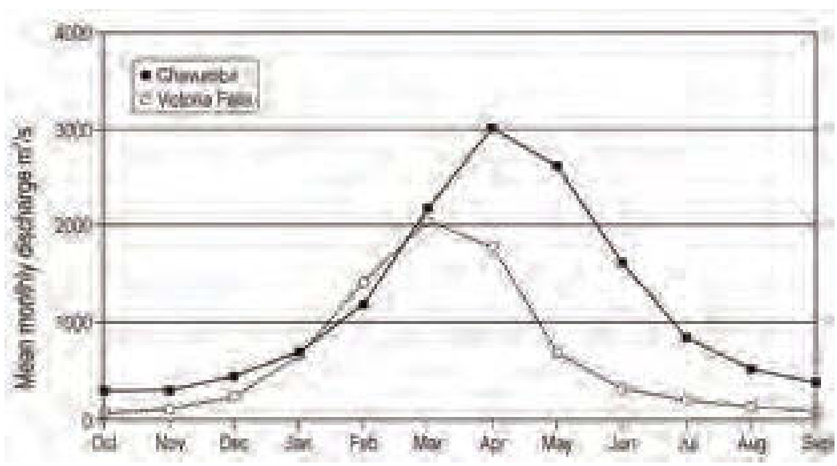

Figure 5.4. Mean monthly flow discharge for the Zambezi River at Chavumba and Victoria Falls. (Drawn from data available in Moore *et al.*, 2007.)

A simple classification of southern African rivers divides them into perennial, intermittent (seasonal) and ephemeral. Dryland rivers can be further subdivided into endogenous and exogenous, where endogenous rivers have their headwaters in humid mountain areas and therefore tend to be perennial, though often strongly seasonal (e.g. the Orange and Sabie Rivers). Exogenous rivers have semi-arid or arid headwaters and therefore tend to be ephemeral, their catchments being too dry to produce runoff other than during storm events. Most Karoo rivers fall into this category, including the major basins of the Great Fish, Sundays and Gouritz Rivers. River basins can also be exoreic, with an outlet to the ocean, or endoreic, ending in an internal basin with no outlet (e.g. the Okavango). A natural transmission loss into the bed and banks of dryland rivers reduces the downstream flow volume (see

Hughes and Sami (1992) for South Africa, and Morin *et al.* (2009) for the Kuiseb in Namibia), a trend that is increased by water abstraction for irrigation and other use.

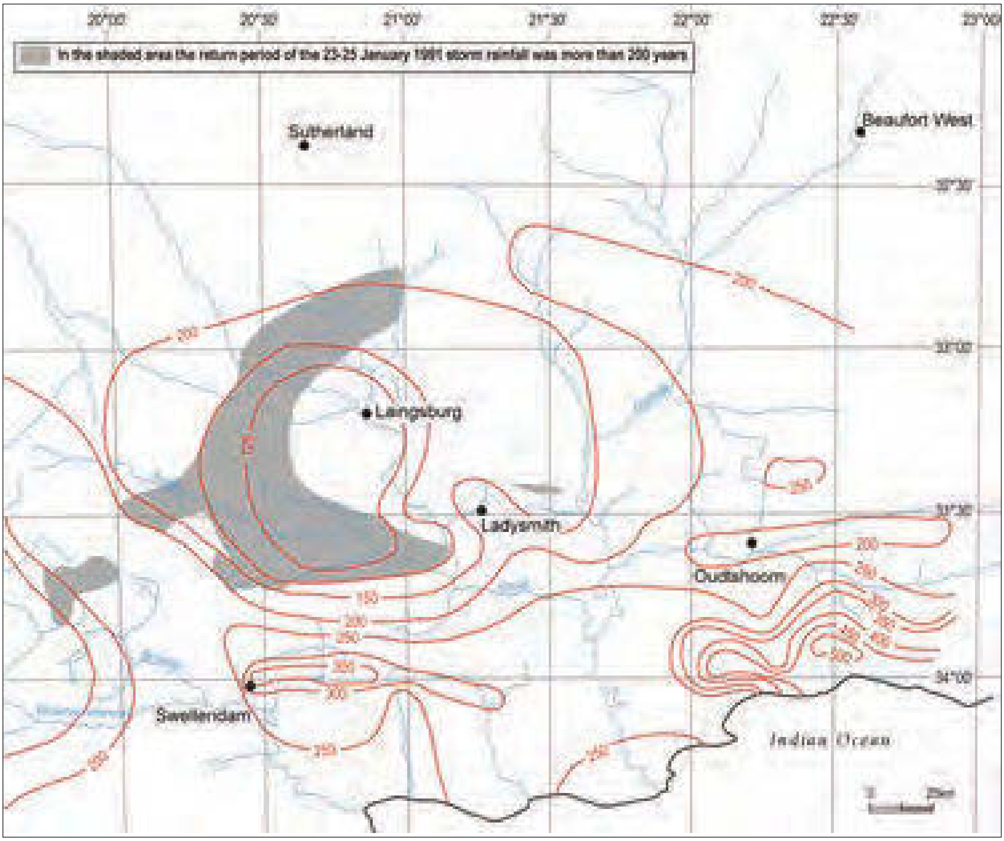

Figure 5.5(a). The flood at Laingsburg, January 1981. Rainfall isohyets 23-25 January. Shaded area represents the area for which the 3-day rainfall exceeded the 200-year return period.

The smaller runoff producing rainfall events in semi-arid areas tend to be isolated in time and space, which means that only small areas of a catchment are effective (produce runoff) during any one storm. Rare events, however, may extend over much larger areas, producing large amounts of runoff from the whole catchment. For example, the Laingsburg flood of January 1981 was caused by an upper-air cut-off low, the Kruger Park floods of 2000 by the tropical cyclone, Eline that affected extensive areas of southern Africa. In the case of the Laingsburg flood an average three-day rainfall of 142 mm fell over an area of 4 000 km^2 (Kovacs, 1982) (Figure 5.5a). Kovacs (1982) gave the peak flow through Laingsburg as 5 740 m^3/sec and estimated a return period of 500 years. He estimated a mean velocity of over 3 m s^{-1} at the flood peak, which carried huge quantities of sediment that covered many buildings in the flooded areas (Figure 5.5b). Kovaks (1982) recorded a mean sediment concentration of 9%. Heritage et al. (2004) estimated the flood in the lower Sabie River to be 6 000 to 7 000 m^3/s, with a return period exceeding 200 years, but in the middle and upper catchment the return period was only 50 years. This reflects the widespread distribution of rainfall that produced runoff from the entire catchment, including the semi-arid lower areas.

These two examples of extreme floods support the findings of McMahon *et al.* (1987) who found that in southern Africa and Australia there was a positive relationship between the ratio of the 100-year flood to the mean annual flood (Q_{100}/Q_{maf}) and catchment area, indicating that as area increases the difference between the biggest and smallest floods also increases.

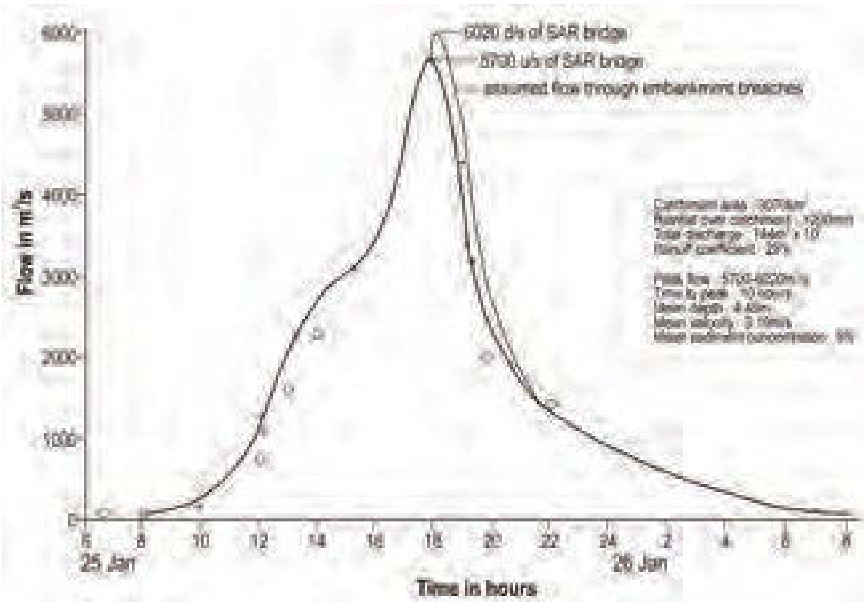

Figure 5.5(b). The flood at Laingsburg, January 1981. The flood hydrograph at the SAR bridge in Laingsburg (redrawn from Kovacs, 1982).

This brings us to a discussion of the concept of magnitude and frequency of channel forming flows (the dominant discharge debate) and its application to southern Africa rivers. Many authors have discussed this concept; reviews by South African authors Dollar and Rowntree (2003) and Beck and Basson (2003) inform the following discussion.

It is widely believed that the flow magnitude with a recurrence interval of between one and two years is the flow responsible for the majority of sediment transport and therefore also for shaping the channel. This flow has been equated to the bankfull discharge; that is, the discharge that just overtops the channel bank and spreads out onto an adjacent floodplain. Flows above bankfull provide lateral connectivity, recharging the floodplain with water, sediments and nutrients.

The concept of hydraulic geometry that expresses the relationship between channel form and flow discharge was developed by Leopold and Maddock (1953). Their research showed that downstream changes in bankfull width, depth and velocity were related to a discharge variable such as the flood with a recurrence interval of 1.5 years (i.e. the dominant discharge). Beck and Basson (2003) reviewed previous findings from the international literature, summarised in Table 5.2. They also carried out an extensive study of the hydraulic geometry of South African rivers using cross-sections surveyed by the Department of Water Affairs and Forestry (DWAF) at sites that were to be dammed (Table 5.3). They introduced a second variable in addition to discharge – water surface slope – to incorporate the concept of stream power (see Box 1, page 110). They found the 10-year flood to give the best coefficient of determination when compared to the 1 in 2-, 5- and 20-year flood. Using this flood as the driving discharge variable they derived the equations given in Table 5.3. It can be seen that their exponent values for width were slightly lower than those given in Table 5.2, but the depth exponent was significantly higher. This suggests that the channel formed by the 10-year flood tends to be deeper and narrower than has been found in other environments. The corresponding exponent for velocity would be 0.053, considerably lower than the international 'norm' (Table 5.2). The incorporation of water surface slope improved the coefficient of determination for both width and depth.

Table 5.2. Width and depth exponents for downstream hydraulic geometry equations (taken from summary of international research compiled by Beck and Basson (2003). Velocity was not included as a variable in their study, but the velocity exponent normally varies between 0.09 and 0.16 (after Knighton, 1998).

	WIDTH EXPONENT (b)	DEPTH EXPONENT (f)
$W = A Q^B$ $D = C Q^F$	0.5 – 0.53	0.30 – 0.34
	width increases fastest	depth increases more slowly
W = width, D = depth, a & c constants, b & f exponents		

Table 5.3. Regime equations derived for 59 reaches on South African rivers (after Beck and Basson, 2003).

DEPENDENT VARIABLE	EQUATION	CONSTANT CW / CD	EXPONENT Q (a)	EXPONENT S (b)	R^2
Width (W)	$W = CwQ^a$	4.417	0.485		0.51
	$W = CwQ^aS^b$	2.488	0.357	-0.230	0.66
Depth (D)	$D = CdQ^a$	0.125	0.462		0.72
	$D = CdQ^aS^b$	0.085	0.377	-0.153	0.82
W = width, D = depth, S = water surface slope, Cw and Cd are width and depth constants respectively, a & b exponents					

Hydraulic geometry relationships depend on two assumptions. First, the dominant discharge concept holds and, second, the channel in question responds in a manner that leads to an equilibrium condition. Knighton and Nanson (1997) question the extent of equilibrium conditions in dryland rivers with highly variable flow regimes, while Nanson *et al.* (2002) and Tooth and Nanson (2011) argue that both equilibrium and non-equilibrium forms can occur in dryland rivers, sometimes co-existing in one channel (Nanson *et al.*, 2002). These authors discuss the importance of high magnitude, low frequency events in controlling channel form in dryland rivers, which are widespread over southern Africa, and show how the effects of individual events may be preserved for longer in the landscape than in humid counterparts. Some dryland rivers, therefore, can display transient states rather than equilibrium forms. This was in accordance with results found by Heritage and Van Niekerk (1995), Rountree *et al.* (2001), and Heritage *et al.* (2004) for the Kruger Park rivers. Heritage and Van Niekerk (1995) observed a build-up of sediment during drought periods. Rountree *et al.* (2001) found that widespread stripping of sediment during an extreme flood in 1996 (a non-equilibrium response) followed a long period of sediment accumulation towards a more equilibrium type channel form. Heritage *et al.* (2004) showed how the 2000 flood event led to a highly variable response, depending on channel type. While some areas were subject to stripping, sediment deposition occurred elsewhere.

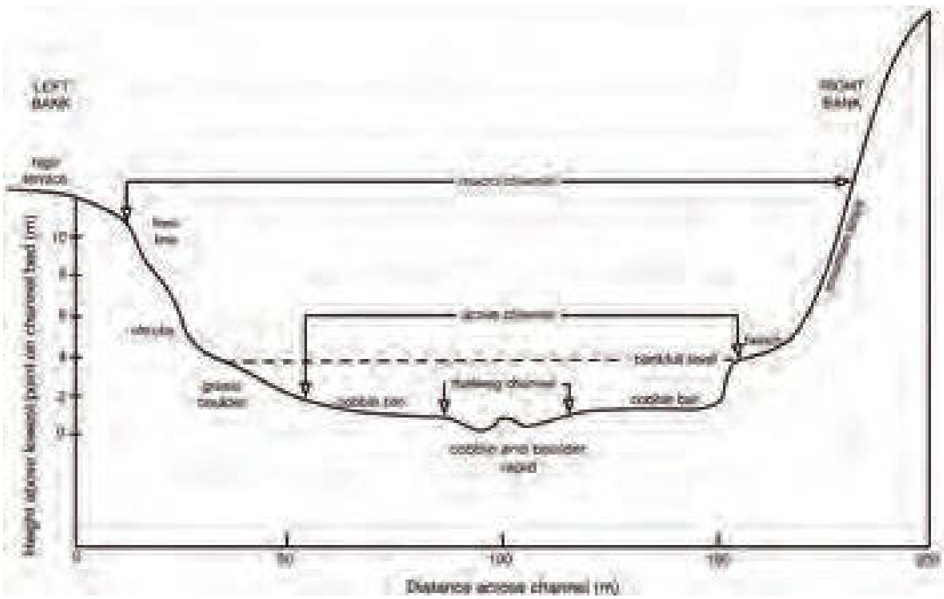

Figure 5.6. Cross-section of the Tugela River showing multiple bench and terrace levels.

In South Africa the situation is complicated further by the complex channel morphology and a strong bedrock influence that constrains channel adjustment in many river systems (Dollar and Rowntree, 2003; Heritage *et al.*, 2003). The complex channel form is exemplified by rivers draining the southeastern coastal hinterland (Figure 5.2) that typically have channels with a compound form, the active channel being incised below a higher river terrace. One or more inset flood benches within a macro-channel replace the floodplain, as indicated in Figure 5.6. Sheldon and Thoms (2006) note that composite river forms with inset benches are also common in many Australian rivers characterised by highly variable flow. The relationship between flood magnitude-frequency and channel form in these compound channels has been researched for the semi-arid Kruger Park rivers by Heritage *et al.* (2001) and for the perennial Mkomazi, Mhlatuze and Olifants Rivers of Limpopo Province of South Africa by Dollar and Rowntree (2003). Both groups of workers found that inset benches adjacent to an active channel could be related to frequent floods of one to two year recurrence interval, but the macro-channel was related to floods with a ten- to twenty-year recurrence interval. It is possible that Beck and Basson (2003) used the macro-channel when developing their hydraulic geometry relationships.

Floods are clearly a crucial component of the flow regime that determines sediment transport and channel morphology. Unfortunately many rivers in southern Africa are ungauged, and big floods often exceed the measuring limits where gauges do exist. Runoff has been modelled for all quaternary catchments as the basis of water resource assessment (Middleton and Bailey, 2009), but these simulations do not provide flood statistics. Qquaternary catchments are the smallest catchment subdivision used by the DWAF for water resource assessments purposes; they are sub-catchments that are considered to be homogenous in terms of their hydrological response. There have been attempts to develop regional flood frequency curves that estimate floods of different return periods from catchment area (e.g. Farquharson *et al.* (1992) for southern Africa, Kjeldse *et al.* (2002) for KwaZulu-Natal Province of South Africa), while the South African National Roads Agency (2006) provides a method for estimating flood peaks from ungauged catchments. There is scope for extending the geographic range and accuracy of these studies.

2.2 Sediment yield

Sediment yields for southern African catchments have been estimated in three ways: by monitoring the concentration of sediment carried by flow (Rooseboom and Harmse, 1979; McCarthy *et al.*, 1991; Jacobson *et al.*, 2000; Grenfell and Ellery, 2009), by estimating the volume of sediment trapped in reservoirs (Msadala *et al.*, 2011; Foster *et al.*, in press), by estimating the mass of terrigenous mud stored on continental margins (Martin, 1987; Compton *et al.*, 2010).

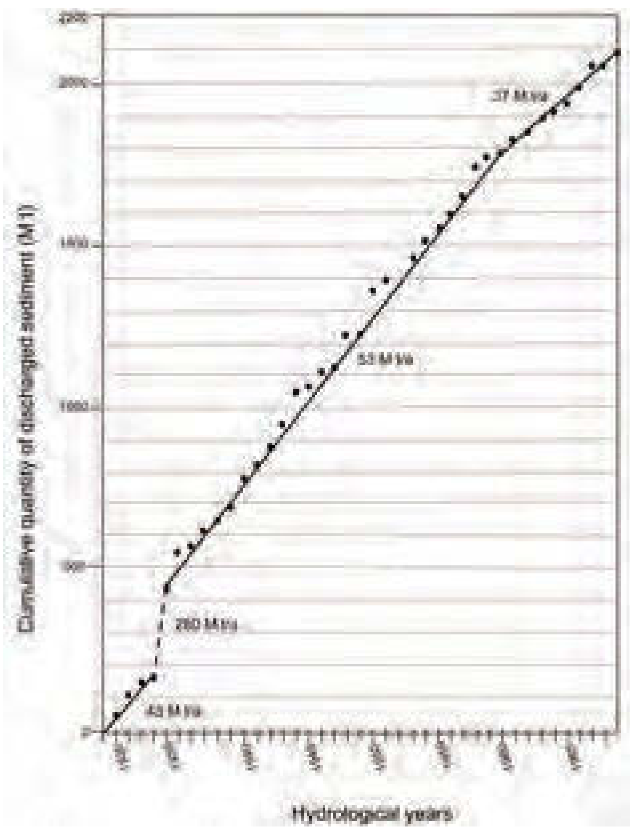

Figure 5.7. Cumulative load of suspended sediment at Prieska/Upington (modified from Rooseboom and Harmse, 1979).

One of the few rivers to be monitored for a significant period of time is the Orange River. A plot of cumulative sediment loads for the Orange River at the combined monitoring stations of Prieska and Upington constructed by Rooseboom and Harmse (1979) is presented in Figure 5.7. They claimed that these data show that, over the period 1930-1970, the average sediment yield of the Orange decreased by 50%, which they attributed to a decrease in sediment supply. A reappraisal of their data (here presented by Rowntree) indicates a more moderate 30% reduction in yield after 1959, from 53 million to 37 million t/a. Rooseboom and Harmse's (1979) graph indicates an average annual yield of 90 Mt between 1929 and 1934, the figure quoted in their paper. They provide evidence that the catchment area contributing most of this high load lay downstream of the present Gariep Dam. The bulk of this load, however, can be attributed to 1934, when 280 Mt of sediment were measured. This is clearly an anomaly. An Internet search revealed e-mail correspondence between Frykberg and Meyer (2003) that mentions that a severe drought was broken in November 1933 by widespread floods. Further correspondence between the author and Frykberg confirmed that the Orange at Upington was "a huge, wide, orange torrent" and that south flowing tributaries between Upington and Kimberly were also in flood (e-mail, 9 April 2011). Thus heavy rains over the normally dry northern Orange River catchment could have mobilised sediment that had accumulated in the landscape since the last major rainfall event.

The importance of the relationship between sediment availability in the catchment and sediment load in the river has been recognised since the pioneering work of Walling (1977) and others. Rooseboom and Harmse (1979) pointed to the poor relationship between sediment concentration and discharge for the Orange River and emphasised the importance of antecedent conditions and sediment supply. Likewise, Grenfell and Ellery (2009) stated that suspended sediment concentrations in the Mfolozi River were poorly correlated to discharge (Figure 5.8a). They identified an annual hysteresis loop in which the highest sediment concentrations occurred in early summer (Figure 5.8b), when availability of sediment is likely to be highest after the dry winter. Flow rates (specifically mean velocity) were found to be a better predictor of bedload in the sand bed channel of the lower Mfolozi River (Grenfell and Ellery, 2009). Bedload, however, made up a relatively small proportion of the total load over the flow range monitored, with suspended load varying from 2.7 times to 10 times the bedload.

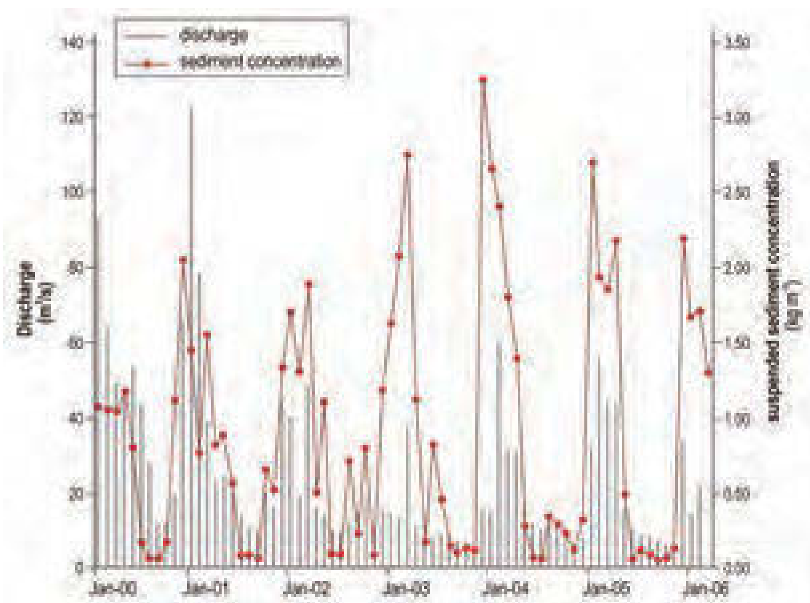

Figure 5.8(a). Relationship of suspended sediment load to discharge in the Mfolzi River: monthly values of discharge and suspended sediment (after Grenfell and Ellery, 2009).

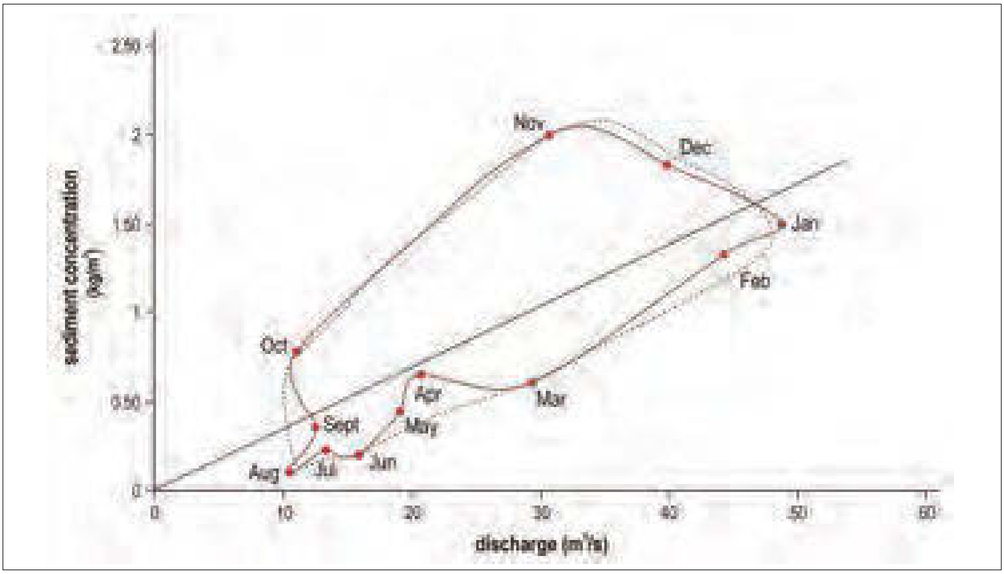

Figure 5.8(b). Relationship of suspended sediment load to discharge in the Mfolzi River: averages changes in monthly suspended sediment showing hysteresis effect (after Grenfell and Ellery, 2009).

McCarthy *et al.* (1991) present results of a study of sediment inputs to the endoreic Okavango Delta, where the Okavango River, which rises in the Angolan Highlands, enters the panhandle. Based on field measurements, they estimate a bedload of between 140 000 and 170 000 t/a and a suspended load of ~30 000 t/a. The bedload flux of moderately well sorted sand was well correlated to flow velocity. They suggest that 90% of the bedload is deposited in the lower panhandle, before reaching the main fan. As explained below (section 5) the main sediment contribution to the lower fan is in the form of soluble salts.

The sediment load of ephemeral rivers is subject to deposition as stream power decays with transmission losses in to the channel bed or banks. Jacobson *et al*. (2000) have carried out one of the few studies of downstream changes in sediment load in an ephemeral stream. Sampling during 12 flood events in the Kuiseb River gave a mean value for total suspended sediments of 35.3 g/l at the peak discharge; the highest value was for a bore sample at the start of a flood, measured at 139.7 g/l. A significant proportion (11.8%) of the suspended sediment was fine particulate organic matter. Deposition of both clastic and dissolved sediment resulted in the river acting as an important sink. The transport of organic matter provides an important carbon source for downstream communities in the river's lower reaches. The distance that sediment is carried downstream varies with the extent of the flood wave, but the authors report that the deposition of sediment in the zone equivalent to the average reaches of floods results in a convexity in the river's longitudinal profile.

Accumulation of sediment in dams has been used to construct sediment yield maps and algorithms for the whole of South Africa. A study by Rooseboom *et al.* (1992) has recently been revised by Msadala *et al.* (2011). This later work is based on sediment yields calculated for the catchments of over 150 dams, subdivided into nine regions. The highest average yields are for KwaZulu-Natal, at 756 t/km^2/a, followed by the western and northern Drakensberg, at 501 t/km^2/a. Both these regions have relatively high rainfall and areas with highly erodible soils associated with Karoo Supergroup mudstones and shales. The regions with the lowest average yields are the southern Cape (76 t/km^2/a) and the southern tributaries of the Limpopo River (54 t/km^2/a). Both these regions are underlain by geology that weathers to give a sandy soil so the suspended sediment load is likely to be low.

Figure 5.9. Island formation in the Okavango Delta has been ascribed either to accretion around termite mounds, forming small islands such as seen in the foreground, or to abandonment of raised channels due to aggradation of sandy sediment, such as seen in the linear island in the background. Water levels were still high when this photograph was taken, leading to flooding of the backswamps.

Msadala *et al.* (2011) combine the reservoir surveys with a national soil erosion prediction map created using the Revised Universal Soil Loss Equation (RUSLE) in a GIS environment (Le Roux *et al.*, 2008). Although this map may highlight areas of greatest erosion risk, it does not take account of sediment storage on valley floors so is not a direct measure of the channel sediment load. Moreover, many reservoirs are located in upper reaches of rivers; the yields should not necessarily be extrapolated to whole drainage basins due to downstream storage. For example, Rooseboom *et al.* (1992) estimated the sediment yield of the Mfolzi River catchment to be 2.36 x 10^6 t/km^2/a, based on sedimentation in a reservoir in the upper catchment, while Grenfell and Ellery (2009) estimated the sediment yield of the Mfolozi River to be only 61 t/km^2/a. They attributed this considerable difference to storage on the Mfolozi River floodplain. Watson *et al.* (1996) provide supporting evidence for floodplain storage during large floods. They state that floods produced by the tropical cyclone Demoina deposited 80 x 10^6 m^3 on the Mfolozi River floodplain.

Sediment yield from small catchments in the Sneeuberg area of the Karoo have been estimated from farm dams (Boardman *et al.*, 2010, Foster *et al.*, 2012). Measured yields for the period from 1935 to 2007 ranged from 115 to 654 t/km^2/a, but a detailed analysis of dam sediments showed that sediment delivery to the reservoirs has varied considerably over the period since dam construction. The highest rates were measured for the catchment at Ganora Farm that had a significant area of badland erosion in the form of a dense gully network. The rate peaked at 1 600 t/km^2/a in the late 1970s.

The Holocene sediment yield from the Orange River catchment was estimated by Compton *et al.* (2010) from the thickness and composition of Holocene sediment on the continental shelf and slope between the Orange River Delta and the Cape Canyon. They calculated a mean Holocene mud flux of 5.1 Mt/a, which they ascribed mostly to erosion of Elliot Formation mudstones at the base of the Drakensberg Escarpment. By comparing their figure to the 52.5 Mt/a suspended sediment flux estimated by Rooseboom and Harmse (1979) (Figure 5.7), they suggest a tenfold increase in sediment yield between 1930 and 1969 relative to the average Holocene rate.

2.3 Longitudinal profile

The river longitudinal profile is an expression of the distribution of potential energy along the course of the river; the steeper the slope, the greater the potential energy. When combined with the flow discharge we get a measure of stream power (see Box 1), which is a measure of the capacity of the river to do work in the form of erosion and transport of sediment.

Box 1

Total stream power can be calculated using the equation:

$\Omega = \rho gQS$ (W)

and

Specific stream power as:

$\Omega = \rho gQS/w$ (W/m^2)

where Ω is the stream power, Q is hydraulic discharge (m^3/s), S is channel slope, w is channel width, ρ is the density of water (1,000 kg/m^3), g is acceleration due to gravity (9.8 m/s^2), W energy in watts.

(See Petit *et al.*, 2005.)

The shape of the longitudinal profile is the product of the long-term development of the landscape and expresses the combination of tectonic uplift and down wearing by the river. A classic long profile is concave upwards, with steep headwater channels gradually decreasing in gradient towards the base level set by the sea. A selection of longitudinal profiles for South African rivers (Figure 5.10), demonstrates that South African rivers do not necessarily have such a simple long profile. According to Maud (2012), the long profiles of rivers draining the eastern escarpment have been strongly impacted by Miocene and Pliocene uplift. Variable lithology also affects the long profile, with more resistant strata forming steps in the profile. For example, the Thukela River has a concave upper profile, but downstream the gradient steepens as the river descends through gorges that extend close to the coast. The Berg River in the Western Cape has a more classic concave profile; this region has been little affected by recent uplift. The profile shape of the Mokolo River in the north, a tributary of the Limpopo, is closely related to the underlying bedrock. Rising in the sandstones of the Waterberg, the upper reaches of the Mokolo River are dominated by bedrock (Figures 5.10 and 5.11). True alluvial sections are found in the lower reaches where it crosses the low gradient Eastern Limpopo Flats (see Figure 5.2).

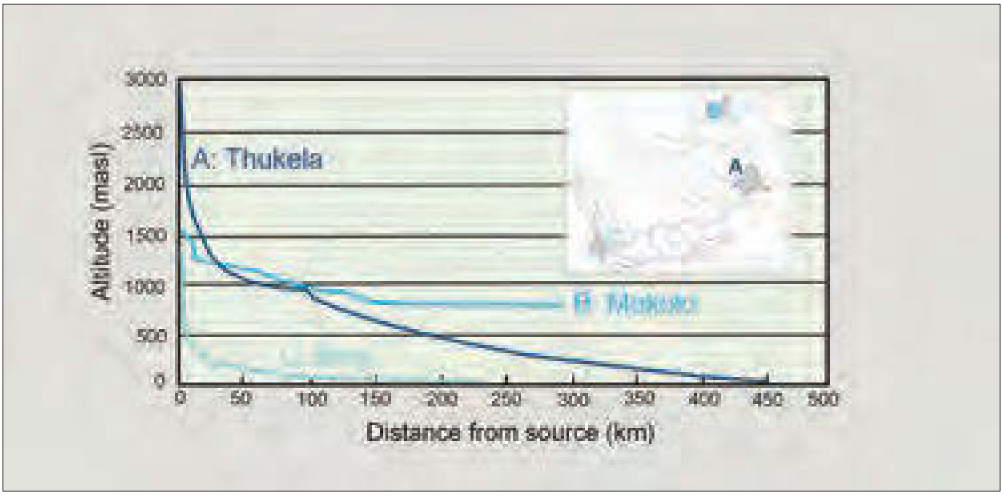

Figure 5.10. Longitudinal profiles of the Mokolo, Thukela and Berg Rivers. Map shows the locations of their catchments. Profiles were constructed from 1:50 000 topographic maps, with a 20 m contour interval.

Figure 5.11. Rapids over bedrock in the middle Mokolo River, Limpopo. The main channel is indicated by white water over the rapids, flood water inudates woody vegetation on the flood bench. The longitudinal profile of the Mokolo steepens in its middle reaches as the river incises through the Waterberg Formation sandstones.

Early ecological typologies of stream type developed by ecologists (see for example Noble and Hemens, 1978) recognised the importance of geomorphological change related to a river's longitudinal profile. Rowntree *et al.* (2000) developed the geomorphological aspects of this typology further as a geomorphological zonation system in collaboration with river ecologists. The river types in Rowntree *et al.*'s (2000) system recognise both position along the river course and local channel gradient. It also recognises changes in hillslope-channel coupling (see Harvey, 2002) and changes to valley floor width. The geomorphological zonation system is presented in Table 5.4. Rejuvenated reaches affected by uplift occur as discontinuities along the classic graded profile, so are classified separately.

Following work in a number of different rivers around the country, it has become evident that channel gradient is a good first indicator of channel morphology and bed material and that changes in channel type down the profile can be identified from an analysis of gradients taken off a 1: 50 000 or 1: 250 000 topographical map. Tooth and McCarthy (2004a) also found that channel gradient was a good discriminator of bedrock and alluvial reaches in the Orange River above Augrabies Falls. Likewise Rountree *et al*. (2001) relate channel type to gradient in the lower Sabie River.

Table 5.4. Geomorphological zonation of river channels (after Rowntree and Wadeson, 1999; Rowntree *et al*. 2000, with acknowledgement to Harrison, 1965; Olif, 1960; and Chutter, 1967).

GEOMOR-PHOLOGICAL ZONE	CHARAC-TERISTIC GRADIENT	DIAGNOSTIC CHANNEL CHARACTERISTICS	HILLSLOPE-CHANNEL COUPLING
Source zone	not specified	Low gradient, upland plateau or upland basin able to store water. Spongy or peaty hydromorphic soils.	Moderate
Mountain headwater stream	>0.1	A very steep gradient stream dominated by vertical flow over bedrock with waterfalls and plunge pools. Normally first or second order. Reach types include bedrock fall and cascades.	Strong
Mountain stream	0.02 – 0.099	Steep gradient stream dominated by bedrock and boulders, Locally cobble or coarse gravel in pools. Reach types include cascades, bedrock fall, step-pool, plane bed, pool-rapid or pool-riffle. Approximate equal distribution of "vertical" and "horizontal" flow components.	Strong
Foothills (cobble bed)	0.005 – 0.019	Moderately steep, cobble-bed or mixed bedrock-cobble bed channel, with plane bed, pool-riffle or pool-rapid reach types. Length of pools and riffles/rapids similar.	Moderate to strong – narrow valley floor, stream undercutting on one side
Foothills (gravel bed)	0.001 – 0.005	Lower gradient mixed bed alluvial channel with sand and gravel dominating the bed, locally may be bedrock controlled. Reach types typically include pool-riffle or pool-rapid, sand bars common in pools. Pools of significantly greater extent than rapids or riffles.	Moderate – valley floor widening, floodplain often present.
Upland floodplain	<0.001	An upland low gradient channel, often associated with uplifted plateau areas as occur for example beneath the eastern escarpment; meandering channels, often associated with floodplain wetlands. Downstream bedrock control is often present.	Weak – channel formed within significant floodplain or floodout.
Lowland sand bed	<0.001	Low gradient alluvial sand bed channel, typically regime reach type.	Variable – often confined, but fully developed meandering pattern within a distinct floodplain can develop in unconfined reaches where there is increased silt content in bed or banks.
Rejuvenated bedrock fall / cascades	>0.02	Moderate to steep gradient, often confined channel (gorge) resulting from uplift in the middle to lower reaches of the long profile, limited lateral development of alluvial features, reach types include bedrock fall, cascades and pool rapid. Bedrock channels widespread.	Strong – confined gorge
Rejuvenated foothills	0.001 – 0.02	Steepened section within middle reaches of the river caused by uplift, often within or downstream of gorge; characteristics similar to foothills (gravel/cobble bed rivers with pool-riffle/pool-rapid morphology in a mixed alluvial/bedrock channel) but of a higher order, with higher stream power. A compound channel is often present with an active channel contained within a macro channel, flooded only during infrequent flood events. A narrow floodplain may be present between the active and macro-channel.	Moderate to strong – confined gorge, but narrow floodplain may develop.

Figure 5.12. The Orange River gorge below the Augrabies Falls. The flat surface into which the gorge is incised is ascribed to the African erosion surface that formed during the Cretaceous period. Uplift of the sub-continent during the Miocene and Pliocene resulted in the incision of the lower courses of many rivers.

3. Space and timescales, and hierarchical classifications

Hierarchical classification frameworks are commonly used as the basis of river classification systems, especially by river scientists engaged in linking the geomorphological template to riverine ecology. Rowntree and Wadeson (1995; 1997; 1999) adapted the scheme presented by American ecologists (Frissell *et al*., 1986) for application to South African rivers. This classification was devised to meet the need

> *... to integrate ecological and geomorphological thinking ... through a classification system that would describe geomorphological processes across a wide range of scales ... that was both relevant and meaningful to ecologists (Rowntree and Wadeson, 1999:2).*

A similar scheme was developed contemporaneously for rivers in the Kruger Park (Van Niekerk *et al*., 1995; Moon *et al*., 1997). The idea of a hierarchical structure has since been taken up by a number of authors (Thoms and Sheldon, 2002; Dollar *et al*., 2007; Parsons and Thoms, 2007) as a way to integrate interdisciplinary approaches to river science and management.

The basic assumption underlying a hierarchical classification is the geomorphological premise that the structure and dynamics of a river system are determined by the surrounding catchment, and landform structure at one level is driven by processes at a higher level. At the same time, each higher level is comprised of an amalgamation of lower level units (Parsons and Thoms (2007) use the term 'holon'). Hierarchical systems are thus built from the bottom up, but driven from the top down. The flux of water and sediment through the channel network is determined by processes in the contributing catchment, the channel morphology for a given reach of river (see below for a definition of a reach) is determined by the sediment budget for that reach, which in turn depends on the flux of sediment coming from upriver.

Parsons and Thoms (2007) defined boundaries between holons in terms of scale and process rates. They stated that higher levels have slower rates of change and are subject to lower frequencies of events. These spatial scales of the hierarchical classification thus can be linked to timescales of change and also to the most relevant timescale for studying process-response systems. For example, a drainage basin may change its overall form over millions of years (cyclical time) whereas the form of a morphological unit may change during a single flood event (steady-state time) or more gradually through a series of flood events (graded-time). Rowntree and Wadeson (1999) point out that the relationship between time and space scales at the reach scale and lower may be different for alluvial and bedrock rivers because of the greater resistance to change displayed by bedrock reaches.

A hierarchical scheme for describing and classifying a river system is detailed in Table 5.5. This is based on the South African schema of Heritage et al. (1997), Rowntree and Wadeson (1999), and Dollar *et al.* (2007).

Table 5.5. A geomorphic hierarchy (after Heritage *et al.*, 1997; Rowntree and Wadeson, 1999; Dollar *et al.*, 2007).

HIERARCHICAL LEVEL	DEFINITION	DEFINING FEATURES
Province	Cross-catchment regions with similar geomorphological characteristics.	Boundaries based on discontinuities in the river long profile and valley form.
Catchment	Land surface that supplies water, sediment and other materials to a channel network.	Drainage divide; line of maximum relief containing the channel network.
Segment	A length of channel with no significant change in the imposed flow discharge or sediment load.	Significant tributary junctions.
Reach	A length of stream channel within which uniform boundary conditions result in a characteristic river morphology.	Uniform slope gradient and valley floor width, resulting in a characteristic channel pattern/ type and assemblage of morphological units.
Morphological unit	Basic erosional or sedimentary structure comprising the river channel.	Discrete area subject to erosion or deposition processes.
Habitat patch	Area within a channel that provides a homogenous habitat for a given organism or community of organisms.	Depends on organism or community of concern.

At the highest level is the geomorphological province (Partridge *et al.*, 2010; Figure 5.2). The second level, the catchment, is the landscape unit that provides the source of water and sediment to the channel network; it is also the main surface for human activity and biological productivity by way of terrestrial ecosystems. Catchments can be described or classified in terms of their component provinces.

The hierarchical organisation of the river channel has four levels. The segment, the reach and the morphological unit are physical channel features. Nested within the morphological unit is the habitat patch – the morphological component of the river ecosystem. Key features of these four levels will be described in turn. Further details can be found in Rowntree and Wadeson (1999).

3.1.1 Segment

The segment is a portion of the channel network that provides a subdivision that is linked to concepts of stream order. Along the length of a segment there should be little change (or no abrupt change) in external drivers (flow discharge, sediment load and sediment calibre), but there should be significant changes in one or other of these variables between segments. Although external drivers do not change, significant change of channel morphology within a segment can be expected due to changes in the local channel gradient and boundary constraints, resulting in the subdivision of a segment into reaches. Segments can be classified according to the geomorphological zonation in Table 5.4.

3.1.2 Reach

The reach is probably the most common unit of geomorphological study. A reach has a characteristic channel morphology that is the result of a spatially uniform channel slope, valley floor width, underlying bedrock resistance and vegetation on the channel banks. Reaches have been variously described in terms of their reach type (Rowntree and Wadeson, 1999), channel type (Heritage *et al.*, 1997) or river style (Brierley *et al.*, 2002; Brierely and Fryirs, 2005).

A simple classification of river type is into bedrock and alluvial channels, formed respectively within the underlying bedrock or within the sediment that is carried and deposited by the river. Bedrock channels are said to be supply-limited because the transport capacity of the flow is greater than the supply of sediment. Alluvial channels form where there is a plentiful supply of sediment relative to the capacity of the stream to transport it and are said to be transport-limited. The channels of many South African rivers locally alternate between a bedrock and alluvial character (i.e. mixed channels) (Van Niekerk *et al.*, 1995; Rowntree and Wadeson, 1999; Tooth *et al.*, 2004; Tooth and McCarthy, 2004a).

Reach types for alluvial and bedrock systems have been described by Rowntree and Wadeson (1999). Typical alluvial reach types found in South African rivers include step-pool (in headwater streams with large clasts), plane bed (in cobble bed mountain streams), pool-riffle (in gravel bed foothill streams) and regime (in sand bed lowland streams). Tooth and Nanson (2011) use the term floodout to describe a reach in a 'loosing' stream where a marked reduction in stream power due to loss of flow, lack of confinement or a reduction in gradient results in loss of channel definition and deposition of sediment over the valley floor. Cascades, bedrock fall, planar bedrock and pool-rapid are common reach types of bedrock systems.

Reach types can be further separated in terms of channel pattern: single thread (straight or braided) or multi-thread (braided, anabranching). Heritage *et al.* (2000) identified five main channel types in the lowveld rivers of the Kruger Park (e.g. the Sabie River): alluvial braided, alluvial single thread, pool-rapid, mixed anastomosing, bedrock anastomosing. 'Bedrock anastomosing' is the term they used for multi-thread channels developed in bedrock. Tooth and McCarthy (2004a) recognised similar features in the Orange River, but termed them 'anabranching channels', reserving the term 'anastomosing' for multi-channel meandering systems formed in fine-grained sediments. Tooth and McCarthy (2004a) linked anabranching channel formation to outcrops of well-jointed rocks such as granite and gneiss.

Conditions leading to straight, meandering or braided channels have been distinguished on the basis of stream power expressed as a graph of channel slope versus some measure of dominant discharge (Leopold and Wolman, 1957). Church (2006) distinguishes a number of different channel patterns in alluvial rivers that can be related to channel stability, sediment load and sediment calibre.

3.1.3 Morphological unit

Morphological units are the smallest scale geomorphological channel form that result from local scale fluvial processes. Thoms and Sheldon (2002) termed these functional units. Rowntree and Wadeson (1999) describe morphological units observed in bedrock and alluvial channels in South Africa. Those in bedrock channels are strongly controlled by bedrock structure and resistance. Bedrock pavement is a common morphological unit developed in horizontal rocks such as the sandstones and mudstones of the Karoo Supergroup. Dolerite results in more blocky structures. Bedrock pools, waterfalls, cascades and rapids are other common bedrock morphological units. Van Niekerk *et al.* (1995) identified bedrock core bars as being important features of the lowveld rivers such as the Sabie and Letaba Rivers. These features form islands where fine sediment accumulates over a bedrock core during periods without extreme floods, but can be stripped off during extreme events.

Sedimentary morphology in alluvial reaches is controlled by the calibre (size) of the sediment and the flow regime. Common in-channel features include various types of bars; lateral features include narrow-flood benches, channel banks and the more extensive floodplain. A set of alluvial morphological units commonly described in the international literature is the pool-riffle sequence, best developed in gravel bed reaches with a meandering plan form. Such a reach type is relatively uncommon in South Africa, as are pool-riffle sequences.

3.1.4 Linking geomorphology and ecology – the habitat patch

The habitat patch is the lowest level of the hierarchy and provides the link between the geomorphological template and biotic communities. The scale of the habitat patch depends on the particular community, life cycle stage and even diurnal pattern of activity (Dollar *et al.*, 2007). Wadeson and Rowntree were among the first South African geomorphologists to develop a formalised typology of instream habitat that was relevant to fish and invertebrate ecology (Wadeson, 1994; Rowntree and Wadeson, 1996; 1999; Wadeson and Rowntree, 1998; 2001). Working in collaboration with English geomorphologists they developed the concept of the hydraulic biotope (see Newson and Newson, 2000 for a general discussion). The hydraulic biotope is an integration of variables significant to instream organisms – bed material size and structure, flow depth and flow velocity. The relationship between flow depth and velocity can be expressed as the Froude number, and particular combinations of flow depth and velocity give rise to the surface flow type. This facilitates mapping of habitat patches over a range of discharges (Rowntree and Wadeson, 1996). Rowntree and Wadeson (1999) describe eight flow types: no flow and barely perceptible flow, smooth turbulent flow, rippled flow, standing waves, free falling flow, chaotic flow and upward vertical flow. These can be related directly to terms commonly used by ecologists to describe flow conditions (rather than channel morphology) – pool, glide, run, riffle, cascade and boil respectively. Flow type mapping has been taken up as a means of rapid assessment of habitat conditions in Australia (Howes and Stewardson, 2005; Dyer and Thoms, 2006; Reid and Thoms, 2009) and the United Kingdom (Heritage, Milan *et al.*, 2009).

The hydraulic biotope recognises that habitat is a dynamic, flow-dependent entity, modified by the underlying channel morphology. It is limited to instream habitat. Other river habitats are also dependent on both morphology and flow. Boucher (2002), a botanist working in the Western Cape, developed a typology of vegetation zones linked to the frequency and duration of inundation. He categorised vegetation into three principle zones: aquatic, wet bank and dry bank. These were subdivided into a further six zones linked to channel morphology and inundation frequency as shown in Figure 5.13. Whereas the aquatic zone is wetted by baseflow for at least 50% of the year, the wet bank and dry bank zones are flood dependent. The wet bank is permanently moist and is inundated by floods with a recurrence interval of less than one year whereas the dry bank is inundated by floods with a recurrence interval greater than one year. The dry bank thus lies outside the limits of the bankfull discharge and is characterised by a greater component of terrestrial vegetation. Boucher's (2002) typology, or derivations of it, is widely applied to the assessment of environmental flows in South Africa (King *et al.*, 2003).

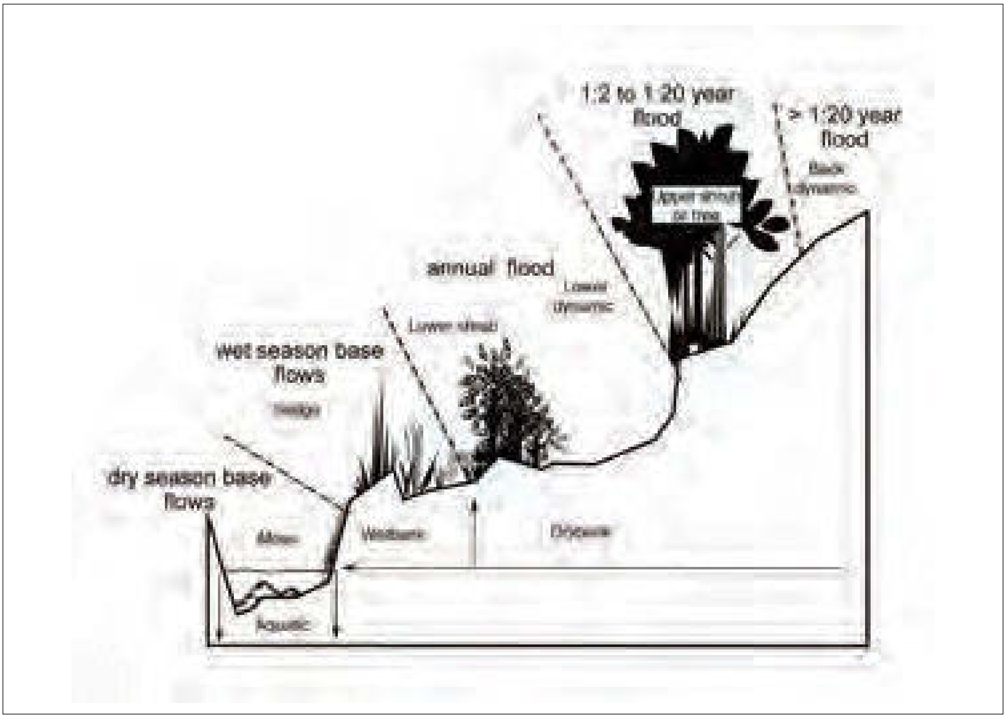

Figure 5.13. Relationship between channel morphology, inundation frequencies and vegetation zones (modified from King *et al*., 2003).

4. Abiotic-biotic links: riparian vegetation

Hydrogeomorphic factors are important in providing suitable habitat for vegetation communities; vegetation in turn is a key geomorphological control. Vegetation plays an important role in stabilising channel banks, bars and, in some cases, channel beds (Graf, 1988; Clarke, 2002; Anderson *et al*., 2004; Trimble, 2004, Nanson *et al*., 2002), but a possible consequence of increased local stability is induced instability in the bigger system (Rowntree and Dollar, 1999). In-stream plants slow the water down during high flows, increasing the height of the water column and forcing water onto the floodplains (Graf, 1988; Clarke, 2002), possibly resulting in the formation of new channels. Vegetation plays a distinctive role in dryland rivers, such as are common in southern Africa, because of the increased water availability along a river relative to the rest of the landscape. In many dryland rivers, deep-rooted woody vegetation that relies on groundwater (phreatophytes) lines the banks of rivers. Moreover, because of the long time periods between significant flood events, vegetation can take advantage of the greater water availability below the bed and encroach onto channel bars and the bed itself (Graf, 1988).

Tooth and McCarthy (2004b) and Rowntree and Dollar (1996) have shown how dense bank vegetation can effect the transition between different channel patterns as first put forward by Leopold and Wolman (1957). Tooth and McCarthy (2004b) investigated the transition from meandering to straight channels in the Okavango Delta, where losses by leakage through permeable vegetation-lined banks caused a downstream reduction in discharge, but minor faulting has caused an increase in gradient. Leopold and Wolman's (1957) model, or derivations of it, did not explain the transition from meandering to straight channels in the Okavango Delta. Rather, they found that Parker's (1976) model where the slope/Froude number ratio is plotted against the depth/width ratio provides a better explanation. They attributed this in part to high bank stability afforded by the vegetation.

Figure 5.14. The Bell River in the Eastern Cape provides an example of how vegetation can cause channel instability. The willows in the center of the picture mark the line of a former channel meander that became constricted due to increased bank stability. A meander cutoff occurred between 1969 and 1975 and has since lead to channel widening to form a braided channel. Increased bedload due to gully erosion in the catchment may have contributed to channel instability.

Rowntree and Dollar (1996) investigated channel instability in the Bell River, a perennial river draining the Drakensberg Range in the Eastern Cape. Both stable and unstable reaches plotted above the threshold line for braiding on Leopold and Wolman's (1957) diagram. However, if resistance due to bank vegetation was included (after Ferguson, 1984), most sites plotted close to or below the threshold line (Rowntree and Dollar, 1996). The authors invoked the combined effect of an increased sediment load due to gully erosion in the catchment (Dollar and Rowntree, 1995) and bank stabilisation due to introduced willows to explain a series of meander cut-offs that resulted in channel straightening and a more braided channel pattern (Figure 5.14). Crack willow (Salix caprea) planted along the banks increased stability, but also caused a reduction in channel capacity and increased overbank flooding across the neck of meanders, leading to channel avulsion (Rowntree and Dollar, 1999).

The invasion of riparian zones by alien vegetation such as Acacia mearnsii impacts on channel processes and form. Rowntree (1991) reviewed the literature on the impact of vegetation on channel form to assess the potential impacts of invasive woody vegetation on geomorphological processes.

Field investigations have since been carried out by Esau (2005) and Pietersen (2009). Pietersen (2009) found that, in small upland rivers in the Cape Fold Mountains, channels under a dense canopy of Acacia mearnsii were on average 40% wider than those under an indigenous fynbos cover and the cross-section area increased by an average of 71%. Observed effects of Acacia mearnsii included a lack of protective ground cover and a shallow rooting system that encouraged undercutting. If trees subsequently fall into the channel, debris dams encourage further widening.

In the Sabie River, Van Coller et al. (1997) related vegetation communities to channel types. Trees such as Breonadia salvinica were able to colonise bedrock areas, including bedrock core bars, whereas the reed Phragmites mauritanus colonised and stabilised sand bars. Beilfuss *et al.* (2001) analysed vegetation change on the lower Zambezi River floodplains and delta following the construction of upstream dams. They found a marked increase in woody savannah species, caused by the reduced frequency of flooding.

Broadhurst *et al.* (1997) carried out an in-depth study of factors affecting channel resistance in the Sabie River. Vegetation was shown to be an important component of roughness, especially at high flows. Heritage, Birkhead *et al.* (2009) found that stripping of cohesive sediments in the Sabie River by the 2000 flood with an estimated return period of between 60 to 200 years was greatly reduced by the presence of vegetation. The recognition that reeds in the lowveld rivers are important in inducing sedimentation and increasing resistance to erosion led to a series of laboratory experiments by James and co-workers (e.g. James *et al.*, 2001; Jordanova and James, 2003; Sharpe and James, 2006. James *et al.* (2002) developed a rule-based model that simulated an event driven response of reed beds. The role of small to moderate flood events was to provide sufficient water to support further reed growth; the ability of large floods to strip out vegetation depended on the stage of vegetation growth.

Extensive research on abiotic-biotic relationships in the Okavango Delta (or, more correctly, the Okavango Alluvial Fan (Stanistreet et al., 1993)) has been carried out by McCarthy, Ellery and co-workers. The contribution of vegetation to determining channel pattern has been mentioned above (Tooth and McCarthy, 2004b). A detailed description of the morphology, sedimentology and channel systems of the Okavango system is given by McCarthy et al. (1997). Two main lines of research are presented here: channel dynamics and island development. Both are strongly affected by the interaction of hydrology, sediment dynamics and biota.

The hydrology of the fan is dominated by a seasonal flood wave that peaks at Mohembo at the top of the pan handle in April, three months after the rainfall peak (McCarthy and Bloem, 1998). It takes a further four months for the peak to reach the distal end of the fan. Water volumes decline down the length of the channels due to transmission losses through the permeable peat banks. A high evaporation rate means that outflow from the fan is minimal, 1.5% of the total annual input of 15 339 Mm^3. As a consequence the load carried by the floodwaters must be deposited within the fan, which is therefore an area of continuous accumulation. Using a mass balance approach, McCarthy and Metcalfe (1990) calculated that 250,000 t/a of chemical sediment accumulates per year, compared to 40 000 t/a of clastic load. They also postulated that precipitation of $CaCO_3$ and SiO_2 in soils causes a significant volume increase. McCarthy and Metcalfe (1990) point to this being the dominant aggradational process at present. The salts are concentrated in islands due to evapotranspiration by deep-rooted trees. Ellery, Ellery *et al.* (1993) showed that vegetation gradients on islands were related to gradients of surface salt concentrations and groundwater salinity. These chemical gradients are in turn a response to vegetation processes (transpiration) (see also McCarthy *et al.* 1993). Island development requires an initial nucleus on which vegetation can grow to start the process of evaporation and salt accumulation. McCarthy *et al.* (1998) suggest that termite mounds commonly form such a nucleus, which expands by lateral accretion to form a more or less circular island. Another morphological feature that can become an island is an abandoned channel (Ellery *et al.*, 1993). Abandoned channels form linear or sinuous islands.

As is typical of fan systems, the channels of the Okavango Fan are characterised by distributaries subject to channel switching over the long term. Channel dynamics are the result of a complex set of processes involving transport and deposition of the sandy bed material, development of peat associated with papyrus swamp, peat fires (Ellery *et al.*, 1989) and creation of pathways by hippopotamus (McCarthy *et al.*, 1998). Flow is lost through the erosion resistant, but highly porous peat that forms the channel banks, causing a loss of stream power and deposition of the sandy bedload. The channels aggrade and become elevated with respect to the surrounding backswamps. McCarthy *et al.* (1998) explain how hippopotamus may prolong the life of these channels by incising pathways along them. They also create paths from the main channels to the backswamps, and cause incision of sinuous paths through the backswamps themselves. These paths can cause flow and sediment to be diverted from the main channel to the backswamps, where aggradation can cause raised channel beds to develop (McCarthy *et al.*, 1998). Once channels decline through aggradation of sediment, they become prone to encroachment by papyrus, leading to a more permanent blockage (Ellery *et al.*, 1995). Following avulsion, the downstream channel and adjacent peat dries out, making the peat vulnerable to peat fires (Ellery, Ellery, Rogers *et al.,* 1989). Terrestrialisation follows, with possible reflooding at a later stage due to a lowering of the ground surface.

5. Connectivity

Connectivity is being embraced increasingly by hydrologists, geomorphologists and ecologists as a concept that allows integration of landscape structure and function at a number of time and space scales (Ward, 1989; Harvey, 2002; Hooke, 2003; Kondolf *et al.*, 2006; Bracken and Croke, 2007; Fryirs *et al.*, 2007; Turnbull *et al.*, 2008; Lexartza-Artza and Wainwright, 2009). Connectivity is a measure of the potential for materials (water, sediment, nutrients) and energy to move over the landscape and may be either structural or functional (Turnbull *et al.*, 2008). Structural connectivity describes the physical links between landscape components; in a drainage basin the channel network provides structural connectivity. Functional connectivity is a measure of the movement of materials (water, sediment, dissolved salts, organic materials); it is process related. Together structural and functional connectivity are well suited to describe process response systems in fluvial geomorphology. Moreover the interaction of structural and functional connectivity allows for non-linearity in landscape response (Fryirs *et al.*, 2007; Turnbull *et al.*, 2008) and thus addresses the concerns of Church (2010) presented above. Connectivity will be used here as a framework to explore further the fluvial geomorphology of South Africa.

The ecologist Ward (1989) introduced the idea of four-dimensional connectivity that acts in the longitudinal, lateral and vertical directions through time (Figure 5.15a). In a river system longitudinal connectivity refers to the down system movement of materials along the channel network. It can include both the main channel and tributaries. Lateral connectivity refers to the movement of materials from the channel outwards onto the banks and floodplain. Vertical connectivity refers to interactions between the channel bed surface and underlying material or hyporheic zone. This may refer to surface water – groundwater interactions or turnover of bed sediment. Kondolf *et al.* (2006) used three-dimensional connectivity as a framework for ecological river restoration. They argued that hydrological connectivity is the defining feature of all riverine ecosystems and link river degradation to disconnection. While stressing the negative aspects of disconnection, these authors also recognise the negative impacts of artificially increasing connectivity and stress the need to restore the natural connectivity regime.

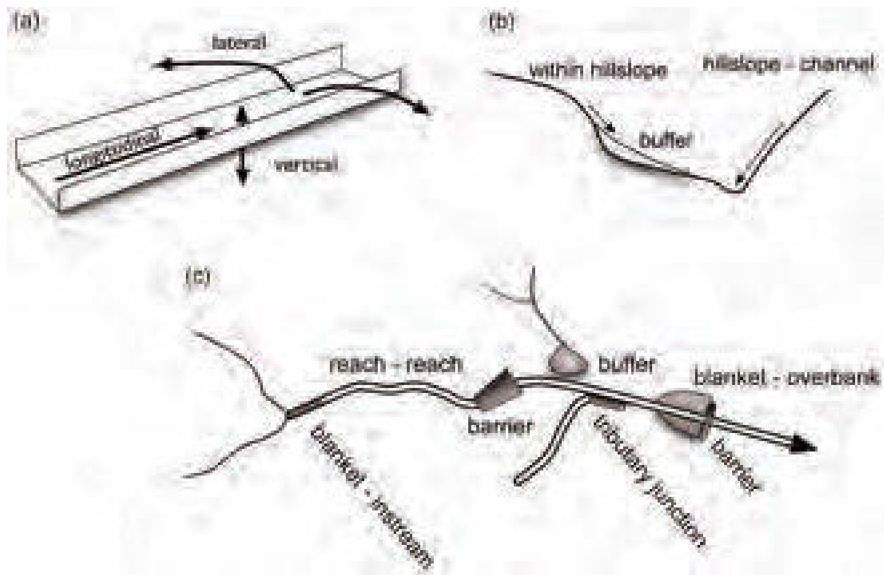

Figure 5.15. Connectivity concepts after Ward (1989), Harvey (2002) and Fryirs *et al*. (2007).

Ward's (1989) concept of connectivity is strongly functional, being concerned with the movement of materials that drives ecosystems. This functional connectivity is dependent on the structural connectivity in the landscape, as expressed by the configuration of hillslopes, valley floor and channels, channel long profile and the channel morphology. Geomorphologists such as Harvey (2002), Hooke (2003) and Fryirs *et al*. (2007) focused on the movement of sediment as a key geomorphic process. They all argued that sediment connectivity controls the propagation of change through the landscape, regulates the sensitivity of response to perturbations and determines the longer-term evolution of the fluvial system. All these authors stress the importance of sediment storage, where storage zones represented disconnectivity in the system (Fryirs *et al*., 2007)

Discussing coupling within landscapes, where coupling refers to the movement of sediment between landscape compartments, Harvey (2002) identified two scales of coupling – local and zonal. Harvey (2002) describes local scale coupling as including within-hillslope coupling, hillslope-channel coupling, tributary junction coupling and reach to reach coupling (Figure 5.15b and c). At the larger scale he differentiated zonal coupling from regional coupling. Zonal coupling occurs between two discrete major zones of a river system (*sensu* Schumm, 1977), whereas regional coupling potentially affects the whole of the drainage basin and is linked to tectonically induced base-level changes. Harvey (2002) linked temporal scales of change to the different spatial scales described above.

Fryirs *et al.* (2007) focussed their attention on sediment stores and sinks. They adopted the term (dis)connectivity to stress the idea that it is as important to consider sediment storage as it is to consider its movement. They distinguished four groups of phenomena that affect (dis)connectivity: buffers, barriers, blankets (Figure 5.15c) and boosters. Buffers are features such as floodplains or fans that prevent sediment from entering a channel; they impact on the hillslope-channel coupling and tributary-channel coupling of Harvey (2002). Barriers disrupt longitudinal connectivity. Natural barriers include bedrock barriers, valley constrictions, zones of sediment storage such as wide, shallow channels with reduced stream power and floodouts. Artificial barriers include dams. Blankets are described as areas of sediment deposition such as floodplain sand sheets, fine grained material in gravel bars or, conversely, bed armour. In all cases of blankets, the surface material is changed relative to that underlying it and is deemed to protect the underlying sediment, thus reducing connectivity. Boosters are configurations that increase connectivity, such as a gorge, where coupling between hillslopes and the adjacent river channel is increased. Boosters (gorges) are a common feature of South African river systems.

The degree of connectivity in any landscape system is linked to effective timescales or to the magnitude-frequency of effective geomorphic events that exceed thresholds for reworking the sediment. Fryirs *et al.* (2007) envisage a series of switches distributed through the landscape that are linked to connectivity. If an event is of sufficient magnitude to rework sediment in storage the switch is turned on and connectivity is established. Clearly, the bigger the event the greater the degree of connectivity, and the greater the catchment area that will contribute sediment down-system.

In the following section, the connectivity framework will be applied to a number of southern African examples: wetland development, geomorphological processes in the Baviaanskloof in the Eastern Cape, sediment dynamics in the Karoo, and the impact of dams and interbasin transfer schemes.

5.1 Wetland development

The upland floodplain category in Table 5.4 is often associated with a natural barrier that disconnects upstream and downstream reaches in rivers subject to rejuvenation. Tooth *et al.* (2004) and Tooth and McCarthy (2007) demonstrated how dolerite intrusions are associated with the development of floodplain wetlands in the Highveld region of South Africa. Tooth *et al.* (2004) present a model of wetland development for South Africa (Figure 5.16) that can be equated to the idea of (dis)connectivity of Fryirs *et al.* (2007). In their model, Tooth *et al.* (2004) link wetland formation to the development of a floodplain with a meandering stream upstream of a barrier – for example a dolerite dyke. These floodplain wetlands form over many thousands of years, but can be reconnected or switched on, to use Fryir *et al.*'s (2007) terminology, if long-term channel erosion breaches the bedrock barrier. Once the wetland becomes connected in the down-system direction, lateral connectivity is compromised by channel incision that reduces the frequency of overbank flooding on to the floodplain. Tooth *et al.* (2004) argue that some instances of incision in wetlands that have been attributed to human induced degradation may be due to natural geomorphic breaching.

A similar example of barriers causing down-system (dis)connectivity is given by Rountree *et al.* (2001). They describe braided reaches upstream of bedrock barriers in the Sabie River. In the Sabie River there is evidence that the main 'switch' operating on a decadal timescale is the rare flood event that causes stripping of sediment deposits (Rountree *et al.* 2001).

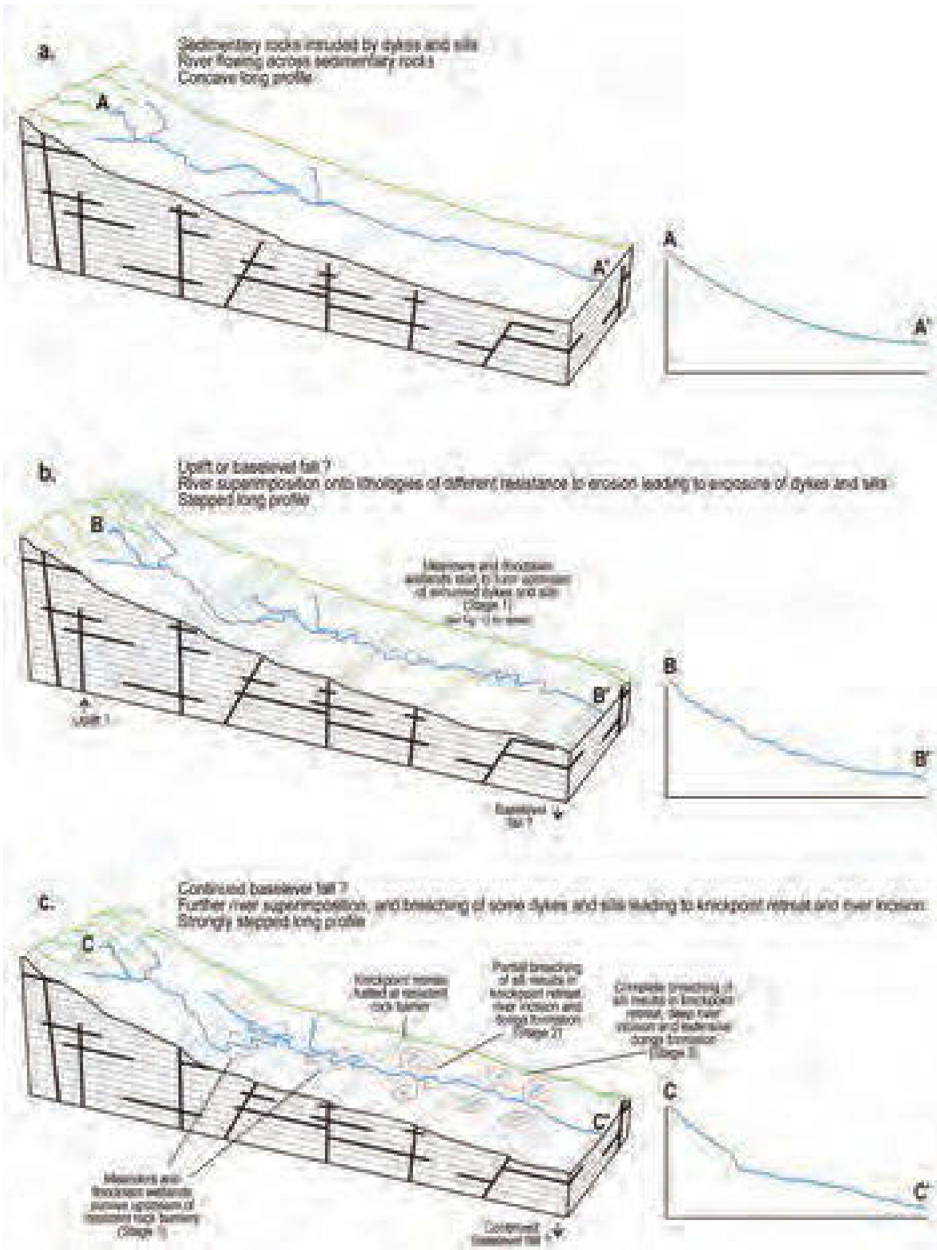

Figure 5.16. Model of wetland development (from Tooth *et al.*, 2004).

5.2 Connectivity in the Baviaanskloof

The Baviaanskloof in the Eastern Cape is a good example of the role of valley constrictions as barriers, alluvial fans as buffers and the impact of human activity on modifying connectivity. Proposed restoration plans are largely about restoring the natural connectivity of this fluvial system. Rowntree *et al*. (2011) present on-going research on fluvial systems of the Baviaanskloof within a connectivity framework.

The kloof is a fault-bounded basin in the Cape Fold Mountains, a region characterised geologically by highly folded and fractured Table Mountain Group sandstones. The valley floor consists of alternating unconfined and confined reaches between 7 and 16 km in length. The unconfined reaches are normally approximately 900 m wide, seldom exceeding 1 000 m, whereas the confined reaches are between 30 and 100 m wide. The channel morphology in the two reach types is distinctive. In the unconfined section, the ephemeral channel takes on a braided form with a higher floodplain and terraces on either side. In locally confined sections, a perennial or intermittent channel is more often single thread with inset benches.

Changes in lateral and vertical connectivity are translated into different hydrological responses in the two reach types. The coarse gravel and cobble of the unconfined sections support a groundwater aquifer that under normal conditions does not intersect the channel surface. Flow is only observed in the channel during or immediately after floods, except in exceptionally wet periods. Floods propagated down the channel are subject to transmission losses in the storage zones so that floods generated in the headwater areas may not reach the end of the valley.

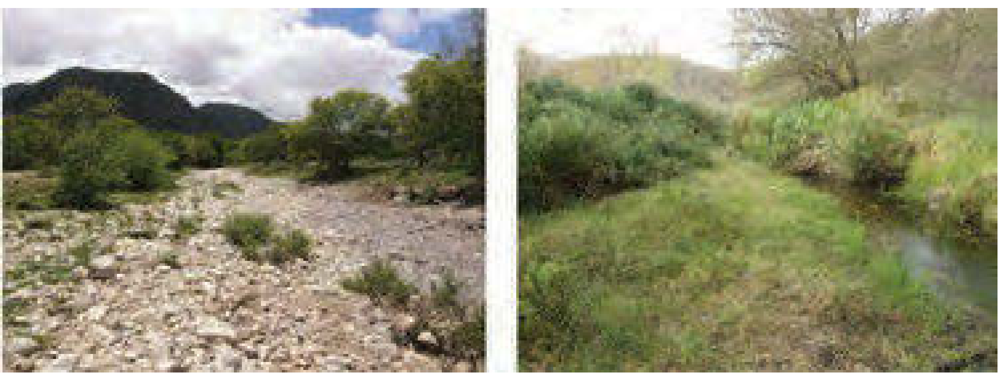

Figure 5.17. Baviaans River: Contrasting channel form in the Baviaanskloof (left) ephemeral channel in an unconfined section of the valley (right) perennial channel in a confined section of the valley.

The hydrology of the confined sections is very different. Surface flow continues for much longer periods of time and some sections are perennial. As groundwater moving slowly down the valley comes through a confined section it is forced to the surface, resulting in baseflow within a narrow inset channel. The presence of permanent surface water promotes vigorous riparian vegetation that contributes to bank strength and inhibits widening of the low flow channel during floods. The wider, higher, less stable channel takes the floods.

Farming activities since the 1980s have negatively impacted connectivity. Lateral connectivity between the channel and floodplains in the unconfined reaches has been reduced through the construction of berms or artificial levees along the channel banks, to prevent floods from inundating cultivated fields on the floodplain. Confining floods to the channels has the effect of increasing the transport capacity of the flow and possibly leading to channel incision, further reducing lateral connectivity. Restoration plans aim to remove the berms and increase connectivity between the river and floodplain, with the aim of increasing groundwater recharge.

Another distinctive feature of the Baviaanskloof is a series of alluvial fans that are located wherever significant tributaries join the main valley in the unconfined reaches. Opposing fans can meet on the valley floor, forming a barrier across the river. Evidence of organic horizons in the river banks indicates that valley floor wetlands may have been widespread in the past. Incision of the main channel through the fans, possibly due to artificial river confinement as described above, has removed these barriers, resulting in a loss of wetlands. The fans also act as a buffer between tributaries and the main channel, storing water and sediment. If channels crossing the fans become incised, their buffering function is lost. Fan incision may be a natural process due to toe trimming by the main channel, or the consequence of farmers channelising flow to prevent damage to infrastructure on the fans. Restoration plans are being implemented to restore the buffering function of incised fans by blocking incised channels and forcing water back onto the fan surface.

5.3 Connectivity and sediment dynamics in the Karoo

Rowntree and Foster (2012) and Foster and Rowntree (in press) invoked changes in connectivity and storage to explain changes in sediment yield from two catchments in the Sneeuberg of the eastern Karoo. Both catchments drained into farm dams, which provide sinks for sediment eroded from the catchment. Sediment cores from the two dams were dated using the radionuclides ^{137}Cs and ^{210}Pb as described by Rowntree and Foster (2012) and Foster and Rowntree (in press). The source of sediment was determined by matching the magnetic signature of the sediment in the core with that of potential source areas in the catchment.

The results for the small 2.78 km^2 catchment on Ganora Farm (Rowntree and Foster, 2012) showed that the rate of delivery to the dam remained low until the 1960s, despite reports from the early twentieth century of widespread overgrazing in the Karoo and the presence of badland erosion in the catchment evident on aerial photographs from 1945. At some point in the 1960s, sediment delivery increased dramatically, peaking in the 1970s. Heavy rains and flooding in part explain the high rates in the 1970s, but not the rapid increase in the 1960s.

The change in sediment delivery rates in the 1960s can be explained by looking at the magnetic signatures of the sediment. The sediment in the lower section of the core, when accumulation rates were low, had a signature that matched soils derived from dolerite, present in the eastern side of the catchment. The increased accumulation after the 1960s is coincident with a shift of magnetic signature towards soils associated with badland erosion in the western half of the catchment. A close examination of aerial photographs revealed that this was coincident with the badland area becoming connected to the main channel draining that side of the catchment. Prior to that connection, eroded soil was stored on the hillslope below the badland area. Thus it was a change in connectivity rather than a change in erosion rate that determined the rate of sediment delivery to the reservoir.

5.4 The impact of water developments on connectivity

Artificial barriers are common on many southern Africa rivers; few major rivers do not have at least one large reservoir that disrupts longitudinal connectivity. Table 5.6 shows the distribution of South African reservoirs by capacity. Seven dams account for 66% of South Africa's storage capacity; only one of these (Pongolapoort Dam) is not in the Orange System. Farm dams smaller than 1 000 m^3 are numerous and widespread, but account for less than 2% of the total storage capacity. Petts and Gurnell (2007) present a global review of the geomorphological effects of dams. The main effects of the dam as a barrier are twofold. First, dams change the magnitude-frequency distribution of floods and, second, they reduce the transmission of fine sediment and prevent coarse sediment from moving downstream.

Rowntree and Dollar (2008) illustrate how the Isandile Dam on the Keiskamma River in the Eastern Cape has reduced the annual flood by 30%. The capacity of the Isandile Dam is 27 500 m^3, with an upstream catchment area of 360 km^2. Beck and Basson (2003) describe how the Pongolapoort Dam reduced the 1:10 year flood from 1 877 m^3/s to 759 m^3/s. They reported that the river below the dam had not experienced any large floods since the dam was built in 1973. The Cyclone Demoina flood of 1984 with a peak inflow of 13 000 m^3/s was almost completely absorbed by the dam, which had very low water levels prior to the flood.

Large amounts of sediment are stored in South Africa's dams. According to Msadala *et al.* (2011), 16% of South African dams have lost over half their storage capacity; by 2010, five billion tonnes of sediment were stored in dams, predicted to double by 2050. Grassridge Dam on a tributary of the Fish River in the Eastern Cape accumulated 38 Mm^3 (51.3 Mt) of sediment over a 50-year period, an average volume of 0.76 Mm^3 (1 Mt) a year that was prevented from moving down the river system (Msadala *et al.*, 2011).

Table 5.6. Dam capacity in South Africa, compiled from the DWA's dam database.

DAM CAPACITY ($M^{3)}$	NUMBER OF DAMS	TOTAL CAPACITY (MILLION M^3)	PERCENTAGE OF TOTAL FOR SOUTH AFRICA
1,000,000-10,000,000	7	32.03	66.24
100,000-999,999	45	10.38	21.46
10,000-99,999	128	3.96	8.20
1,000-9,999	405	1.19	2.45
100-999	2,429	0.72	1.49
1-99	1,193	0.07	0.15

Through their impact on the magnitude and frequency of flood events, dams also affect lateral connectivity in downstream reaches. If the flood magnitude-frequency regime shifts towards fewer, smaller floods, the connectivity between the active channel and benches or floodplain will be reduced (a loss of functional connectivity). Lateral connectivity with the floodplain will also decrease below the dam if a reduction in available sediment results in channel incision (a loss of structural connectivity). Beilfuss *et al.* (2001) describe how the Zambezi Delta exhibits both these impacts. A severe loss of lateral connectivity has resulted from the building of the Kariba Dam in 1958, dams on the Kafui River in the 1960s and 1970s, and the Cahora Bassa Dam in 1974. Only the most extreme floods, such as that of 2000, inundate the floodplain. The main channel, once a wide, unstable braided system, is now confined to one permanent channel that meanders between consolidated, vegetated islands.

The flow of water and sediment is also interrupted by many smaller obstructions in the form of weirs and causeways. To the author's knowledge, the number in South Africa has never been documented, but, as an example, the Kat River in the Eastern Cape has, on average, three weirs per kilometre of river, some reaching a height of six meters. These weirs store water, increase evaporation losses and trap sediment. Weirs are semi-permanent features, subject to breakage. Boardman and Foster (2011) point to the considerable volumes of sediment that could be released downstream if a weir or dam wall is breached. They estimate that in an area of 95 km^2 in the semi-arid Sneeuberg in the Eastern Cape, up to 72 million m^3 of sediment could be released following breaching of farm dams.

Interbasin transfer schemes (IBTs) increase longitudinal connectivity, in this case between basins. Increased baseflow is a common result, often up to the level of the annual flood. Rowntree and Du Plessis (2003) provide one of the few studies of the geomorphological impacts of an IBT, focussing on the Skoenmakers River in the Eastern Cape. This river is part of the Orange-Fish-Sundays IBT. The flow regime of the regulated Skoenmakers River changed after 1978 from a first order ephemeral stream to a baseflow dominated, perennial river with a maximum average daily flow of 4 m^3/s. After 1985, the maximum daily flow increased further to 22 m^3/s. This resulted in dramatic changes to the river morphology as shown by a comparison with the adjacent unregulated Volkers River (Figure 5.18). In the upper reaches there was massive incision and widening, from a shallow channel 0.5 m deep and 5 m wide to an incised channel 6 m deep and 25 m wide. In the middle reaches, the macro-channel was both deeper (6 m) and wider (55 m) than for the Volkers River (1 m and 35 m respectively), but the active channel was confined between benches that had formed on both banks, giving an active channel width of 27 m and depth of 2 m. In the lower reaches, slight widening of the channel and significant deposition of sediment in the form of braid bars had occurred. The width of the macro-channel was similar (60 m in the Skoenmakers River and 50 m in the Volkers River), while the depth of the Volkers River actually exceeded that of the Skoenmakers River (3 m against 2 m).

Thus there was a loss of lateral connectivity in the upper reaches between the floodplain and active channel due to incision, but an increased lateral connectivity in the lower reaches, between the active channel and benches, due to the increased level of baseflow and, in the most downstream reach, deposition in the channel.

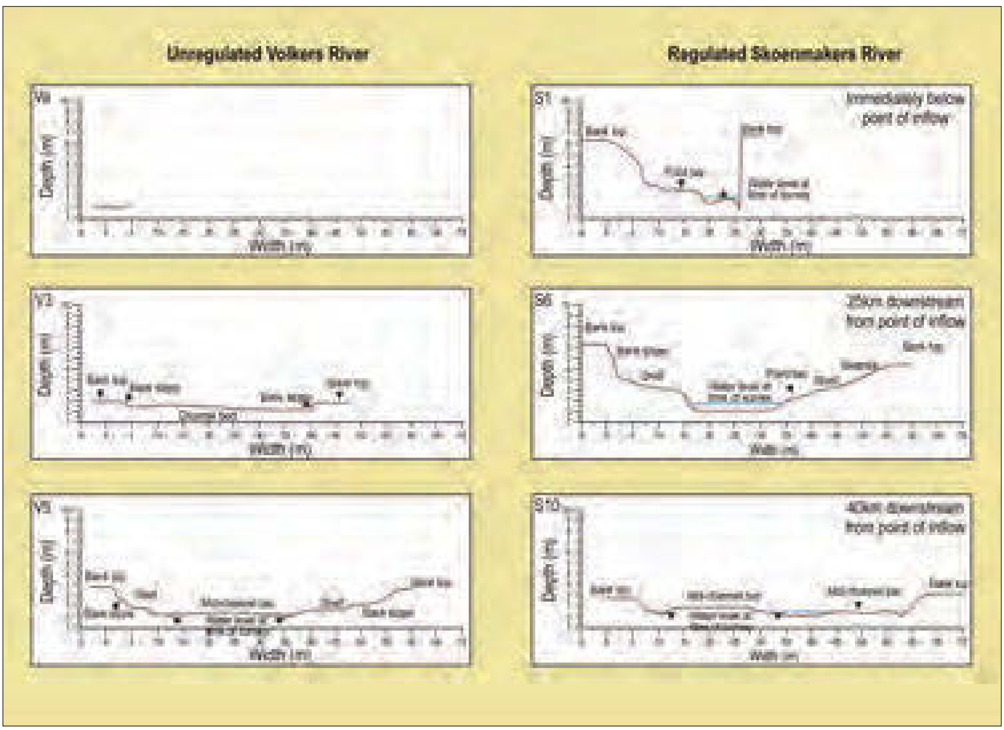

Figure 5.18. Channel change resulting from the interbasin transfer into the Skoenmakers River. The adjacent Volkers River was used as a reference (redrawn from Rowntree and Du Plessis 2003).

6. Geomorphology and river management

Since the mid-1990s South African geomorphologists have made a significant contribution to developing tools for application to river management, largely in response to environmental protection clauses included in the National Water Act of 1998 (Republic of South Africa, 1998). Chapter 3 of this act is concerned, among other things, with the protection of aquatic and associated ecosystems through the ecological Reserve. The ecological Reserve is defined in the Act as "the water required to protect the aquatic ecosystems of the water resource" (National Water Act Chapter 3, Part 3). A complex set of tools has been developed under the umbrella of Resource Directive Measures for assessing and complying with the ecological Reserve. Embedded within these measures is the recognition that the morphology of the river channel provides the abiotic component of the river ecosystems, and that geomorphology is an important driver of aquatic and riparian ecosystems. An overview of river management for ecosystem protection in South Africa and the role of geomorphologists is given by Dollar (2000) and Rowntree and Du Preez (2008). King, Brown *et al*. (2003) provide an overview of environmental flow assessment methods employed in South Africa.

The ecological Reserve is primarily concerned with flow; recommendations must be made as to what flows are necessary to protect the river ecosystem so that "excess" water can be allocated to water users. Floods, a key driver of river morphology, comprise the main flow component of concern to geomorphologists. Rowntree and Wadeson (1998) present guidelines for determining geomorphologically effective flows. They identify three groups of flow: low flows that provide hydraulic habitat (hydraulic biotopes) within a stable channel morphology; within-bank flows that cause seasonal flushing of fine sediment from the coarser matrix of bed material so as to maintain an open bed structure; bankfull flows and greater that are important for creating and maintaining channel morphology. It is common practice to recommend a bankfull (or bench full) flow with a frequency of recurrence of one to two years, in line with the concept of dominant discharge, but this will vary with the natural flow regime (i.e. less frequent in more arid areas).

An important consideration when setting the ecological Reserve is the desired Management Class. There are four classes: natural, moderately used/impacted, heavily used/impacted, and unacceptably degraded. The amount of water use that can be allocated to various users increases from the natural to heavily used class; to manage for the unacceptably degraded class is not permitted by law and restoration measures should be put in place for such degraded river reaches. The Management Class is linked to the Ecostatus, measured in terms of five Ecological Categories, A to F, which range from unmodified or natural to extremely modified, where the degree of modification is linked to the Reference Condition for that river reach. The Reference Condition is the presumed condition in the absence of human impact. An assessment index is used to determine the Ecostatus for each ecosystem component – hydrology, water quality, geomorphology, vegetation, fish and macro-invertebrates. Further details are given by Rowntree and Du Preez (2008).

The main purpose of the Geomorphological Assessment Index (GAI) is to give an indication of the extent to which the physical habitat template has changed from natural (the Reference Condition). It would therefore be logical to base GAI on observed changes to channel morphology. Du Preez and Rowntree (2006) point out that this is problematic due to the possible disequilibrium in many South African river systems, the timescales over which natural change takes place, and the frequent irreversibility of that change. A good example is the channel stripping by floods and the consequent build-up of sediment that has been observed in lowveld rivers, such as the Sabie River (Rountree *et al*. 2001). Which morphological condition represents the Reference? This problem is aggravated by the lack of good historical data on channel morphology for the majority of South African rivers. To get around this problem, Du Preez and Rowntree (2006) defined the reference condition as "the geomorphological system that supports the natural ecosystem, where a system is a set of components connected through flows of energy and matter to accomplish a set function." (Du Preez and Rowntree, 2006, 43) and proposed that GAI be based on the extent that system drivers have been modified by anthropogenic activities. GAI is therefore based on changes to the magnitude-frequency signature of floods, the sediment load and to change in perimeter conditions, especially those related to the stabilising effect of vegetation. All the above changes are placed within a framework of longitudinal, lateral and vertical connectivity. Observed or assumed changes to channel morphology are used as confirmation of the impact of driver changes, rather than as the basis of the assessment per se.

A significant limitation of current environmental flow methodology is linked to the unpredictable nature of rivers and the inherent difficulty in determining how channel morphology will respond to a new flow regime. Current methods tend to assume that the present channel morphology will persist despite changes to the flow regime, which is clearly untenable. This has been addressed by a number of river scientists working in South Africa who have developed models that could be used to predict future channel change under different management scenarios. For example, qualitative rule-based models have been developed as part of the Kruger National Park River Research Programme to predict the response of systems to natural and anthropogenic factors affecting water supply (Jewitt *et al.*, 2001). Heritage *et al.* (1997) developed a sediment transport and storage model for predicting annual change at the channel type scale, while Birkhead *et al.* (2000) presented various geomorphological change models as part of a project aimed at developing a decision support system for water resource management. The challenge remains to integrate these models into the Resource Directed Measures tool kit.

Recent attention has been paid to the possibility of restoring the ecological function of degraded rivers (King, Scheepers *et al.*, 2003b; Uys, 2004). Successful river restoration must begin by addressing hydro-geomorphic functioning of the system (Kondolf *et al.*, 2007). Geomorphologists can play an important role in providing evidence for the historical reference condition, mapping the trajectory of change over time and identifying causal factors and making recommendations for future possible states. It is often not feasible to attempt to return to the reference condition when geomorphologic thresholds have been crossed; rather restoration must aim to establish a modified but sustainable ecosystem that takes account of recent ecological adaptations to changed conditions. McCarthy et al. (2010) exemplify this for the Seekoeivlei floodplain wetland on the Klip River, a tributary of the Vaal River.

7. Conclusion: the future of fluvial geomorphology in southern Africa

Applied fluvial geomorphology in southern Africa is closely linked to initiatives by ecologists to protect water related ecosystems. This has led to an integrated approach to river management and river science in which ecologists and fluvial geomorphologists work closely together (Dollar *et al.*, 2007; Rowntree and Du Preez, 2008). The acknowledgement of the interrelationships between ecology and geomorphology is a distinctive feature of much of the research carried out in southern African, be it in the lowveld rivers of the Kruger Park, the Kuiseb River of Namibia, the Okavango Delta or the lower Zambezi River. Pressure from managers to answers questions such as "what is the geomorphological reference condition of this river channel?" and "what magnitude and frequency of floods are required to maintain the ecological integrity of this river?" has forced geomorphologists to question whether basic concepts such as dominant discharge apply to the southern African fluvial environment where a semi-arid climate and bedrock dominated channels are common rather than the exception. We still have incomplete knowledge on which to base to predictions of how channel morphology may change in the future in response to continued development of water resources, land use change in catchments, climate change and natural long-term geomorphological evolution.

Fluvial geomorphological research in southern Africa has been carried out by a small number of researchers from both within and outside the region. There are clearly many questions and considerations that provide opportunities for further research. These can be grouped as follows.

- We need to have a better understanding of the channel process-response relationships at the three timescales proposed by Schumm and Lichty (1965). How have southern African fluvial landscapes evolved over the long term in response to tectonic uplift or warping and climate change? How does channel morphology respond to catchment drivers in graded time? Should we consider landforms to be in equilibrium or disequilibrium? How do channels respond during a single event, in steady state time? Can we develop more realistic sediment transport models that inform us about the morphological response? Can we develop an integrative framework that encompasses these different time and space scales such as that provided by connectivity?
- How do these relationships differ between the different geomorphological provinces and climatic regions? There are strong environmental contrasts between humid catchments, semi-arid and arid catchments; to what extent do these result in different process form relationships? The downstream connectivity between geomorphological provinces is also an important consideration.
- How can we improve our understanding of geoecological relationships through collaboration with river ecologists? Geomorphologists need to be aware of the habitat requirements of different groups of organisms in order to develop improved models that translate an understanding of fluvial processes into habitat change. Ecologists in turn must be made aware of the dynamic nature of river systems and the possible prevalence of disequilibrium.
- How can geomorphologists best contribute to river management, for example through developing improved methods for ecological flow assessment or through river restoration projects? Only by developing a sound conceptual understanding of our fluvial systems will geomorphologists be able to develop better tools that can guide decision-making.

This chapter has aimed to provide an overview of recent developments in the science of southern African fluvial geomorphology within a conceptual framework borrowed from the broader literature. In a single chapter it has been possible only to give a sketch of the significant developments in fluvial geomorphology in the region since the mid-1990s. It is hoped that it has provided an introduction that will inspire the reader to follow up ideas presented in order to be able to address some of the research areas outlined above.

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6

Aeolian Systems

Aeolian Systems

David SG Thomas and Giles FS Wiggs

1. Introduction

Wind-blown deposits and landforms are a significant geomorphological feature of the southern African landscape. After Hodson's (1912:21) early non-scientific account of the "dreary and depressing" sand hills of southern Bechuanaland, Lewis (1936) provided the first scientific assessment of these dunes, from the neighbouring Northern Cape Province of South Africa. This is the driest area of the Kalahari, where the predominantly linear dune forms, though largely vegetated, display a propensity to wind activation during drought periods. In the same decade, far to the north, the ecological traverses being conducted in Barotseland (western Zambia) by Colin Trapnell in 1933 made reference to east-west belts of loamy, bush covered ridges "... as though they were the relics of regular dune belts" (Smith, 2001:138).

Since these early accounts of the complex aeolian geomorphology of the Kalahari, there have of course been major scientific advances in their interpretation. These have been based on detailed field investigations of forms and sediments, on the monitoring and measurement of aeolian processes and the controls thereon, on vital aerial survey, and through the application of techniques that allow the accumulation histories of dunes to be established.

Although by far the most extensive landscape in southern Africa with an aeolian heritage, the Kalahari represents only part of the sub-continent's response to the geomorphological capacity of the wind. The southern half of the hyper-arid Namib Desert comprises one of the most impressive dune landscapes on earth, and is one where the role of the wind today is significant in shaping landforms. Beyond the Kalahari and Namib, marked, but more localised, evidence of the geomorphological efficacy of the wind also exists, including in the Karoo, the Free State Province of South Africa, and in coastal regions. Coastal dunes are dealt with in Chapter 10: this chapter addresses the aeolian features and processes of the interior, and their associations with aridity present and past. The aim here is not to give an account and description of all aspects of regional aeolian features: this to some degree is provided elsewhere in, for example, Lancaster (1989) for the Namib and Thomas and Shaw (1991a) for the Kalahari. Instead, the focus is on key themes and issues relating to the controls on aeolian processes past and present including dust as well as dune systems, distributional differences in aeolian forms, and the potential responses of aeolian systems to future changes in environmental conditions. Figure 6.1 shows the general distribution of aeolian features in southern Africa, and areas discussed in the analysis in this chapter. Pan areas, where lunette dunes are often present are also shown, but are discussed elsewhere in this volume (Thomas and Shaw, Chapter 7).

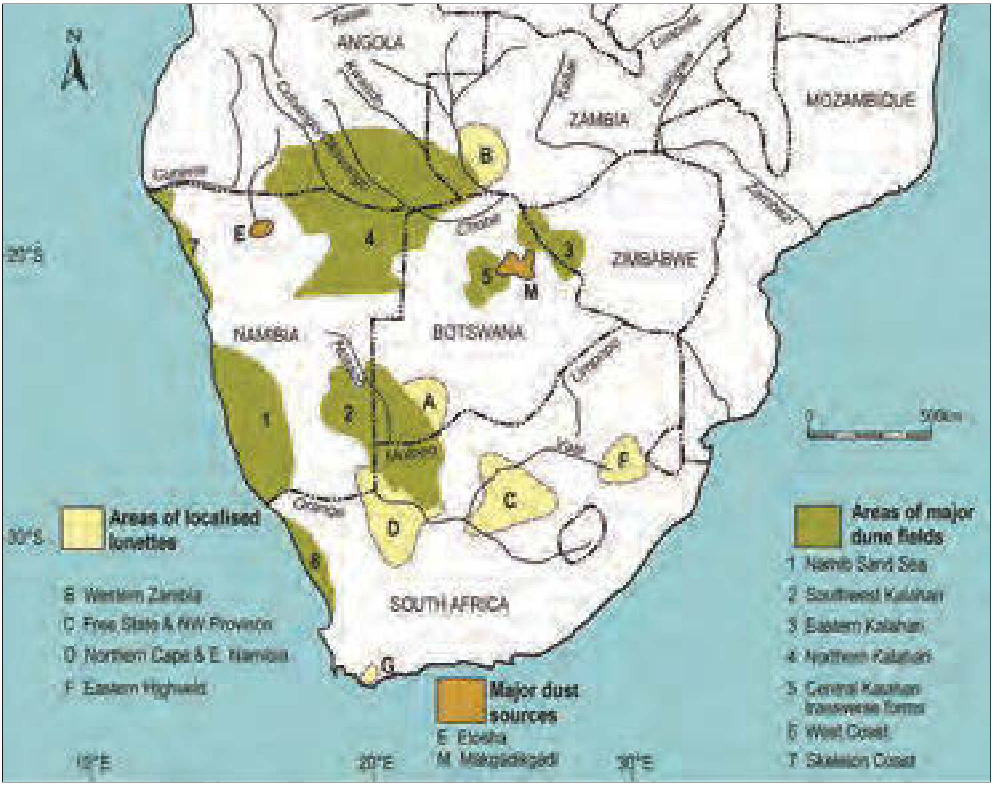

Figure 6.1. The distribution of continental aeolian features and principle dust sources in the southern African landscape. 1-6 refers to major dunefields, including several (no. 2-5) in the Kalahari, where Kalahari sediments over a wide area are regarded to have aeolian affinities even where dunes are absent. Kalahari Dunefields are dominated by linear dune forms though area 5, which has received virtually no detailed research, is comprised of transverse dune forms. The West Coast Dunefield, although having an association with coastal sediment supply (as does area 1, the Namib Sand Sea), is included as a continental dunefield because dune forms extend inland. Pan depressions in southern Africa often, but not exclusively, having fringing lunette sand dunes. These are discussed in Chapter 7 with main areas shown as A-D and F. Aeolian dust sources include Etosha and Makgadikgadi, as well as individual pans and agricultural land throughout southern Africa.

1.1 Interpreting the formation and environmental significance of southern African aeolian systems

In aeolian geomorphology, four recent scientific developments have provided the opportunity for improvements in understanding the nature of and behaviour of aeolian systems (Thomas and Wiggs, 2008). These have been advances in the capacity to make effective empirical measurements of the relationships between wind energy and sediment flux (e.g. Arens, 1996; Van Boxel *et al.,* 2004); mathematical modelling of both dune and dust dynamics and their relationships to key environmental controls (e.g. Baas and Nield, 2007; Zender *et al.,* 2003); developments in remote sensing systems, allowing sediment movement to be monitored from space and the internal structures of sand dunes to be viewed and interpreted (e.g. Washington *et al.,* 2003; Bristow *et al.,* 2000); and advances in dating techniques that allow the timing of dune formation to be established in detail, particularly through the development of Optically Stimulated Luminescence dating (e.g. Aitken, 1994).

Research into southern African aeolian systems has both benefitted greatly from some of these advances and been the source of some of these developments. For example, sand flux studies carried out in the Skeleton Coast Dunefield by Weaver and Wiggs (2011) have benefited considerably from new technological advances and are contributing to advances in recognition of the significance of atmospheric turbulence to sand flux. In the Namib Sand Sea, Bristow *et al*. (2000) used ground penetrating radar (GPR) to great effect to investigate the internal structures of linear dunes, which then led to targeted sampling for OSL dating that improved understanding of dune migration, while advanced applications in remote sensing have been used in studies of dust flux from the Makgadikgadi Basin in the Kalahari by Bryant *et al*. (2007). In terms of understanding the complexity of dune system development, the Kalahari is now perhaps the most intensively OSL-dated aeolian system on earth (see Thomas and Burrough, 2012), yielding complex records of dune accumulation as well as contributing to developments in how OSL age sequences should be interpreted (Stone and Thomas, 2008). These, and other developments, contribute to the analyses and interpretations that follow.

2. Dunes of southern Africa

The continental dune features shown in Figure 6.1 extend well beyond the present-day occurrence of hyper-aridity and aridity, into areas where conditions might be deemed too wet today for aeolian activity to be effective. Dryness, leading to limited vegetation cover, is important for the wind to be an effective geomorphic agent, through its impact on the erodibility of surfaces that might yield sediment to the wind. This simple assertion is complicated by several factors. First is that however dry, without significant wind energy (erosivity), aeolian activity will not occur. Second is that however dry and windy conditions are, aeolian activity will not take place without a supply of suitably available sediment. The interaction between erodibility, erosivity and sediment availability is the key to understanding the occurrence of aeolian processes both in space and time (refer to Wiggs, 2010 and other sources for reviews of the controls and operation of aeolian processes, which are not considered from first principles in this chapter). Temporal changes in aeolian activity in the southern African landscape have occurred, significantly in response to the impacts of decadal to millennial scale climate variability and change, with the principle evidence for this coming from marine core records (see chapter by Meadows) and from recent attempts to unravel the records of past depositional periods present in the sediments within dunes themselves.

There are two main areas in southern Africa where continental dunefields occur. First is the extensive Kalahari region, extending from the Northern Cape northwards through Botswana and eastern Namibia, to Angola, western Zambia, and western Zimbabwe (Thomas and Shaw, 1991a). Second is the extreme west of the subcontinent, including the Skeleton Coast, the Namib Sand Sea (Lancaster, 1989) and the western margin of South Africa north from Cape Town to the Orange River (Chase and Thomas, 2007). The Kalahari Dunefields occur in the elevated interior continental basin with a mean altitude of approximately 900-1 100 m a.s.l. which has accumulated continental sediments since the basin was created following the division of Gondwanaland approximately 200 Ma. Mean annual rainfall today ranges from about 100-150 mm in the southwestern Kalahari of the Northern Cape to over 1 000 mm in northwestern Zambia. The Kalahari Dunefields, therefore, embrace arid to tropical humid conditions today.

The Namib and western margin dunefields lie between the western arm of the Great Escarpment and the Atlantic in a sandveld and desert zone that varies in width from about 20 km near the Orange River, to 60 km in the south at Cape Columbine and over 100 km in the central and northern zones of the Namib, near the Kuiseb River. North of Cape Town, dunes occur in part of the winter rainfall zone that receives approximately 400 mm pa precipitation, rapidly decreasing in a northern direction, while the Namib is of course hyper-arid and one of the driest places on earth, with parts receiving no perceptible annual rainfall and precipitation often dominated by moisture gained from fog (Olivier, 1995).

Taken as a whole, the continental dunefields of southern Africa include all the major dune types that geomorphological classifications (e.g. Lancaster, 2011) encompass (Figure 6.2), though overall linear dunes dominate the southern African aeolian landscape. Dune form varies on account of a range of factors (e.g. Wasson and Hyde, 1983) of which annual wind directional regime variability is perhaps most dominant. Thus linear dunes tend to form through extension, in bi-directional wind regimes, transverse forms (including barchan dunes) in unidirectional regimes that encourage dune migration, and star dunes in multidirectional regimes that encourage sand accumulation. That linear dunes are southern Africa's, and the world's, most common dune type (Lancaster, 1982), is probably due to a combination of the fact that unidirectional wind regimes are relatively unusual, and that the extending and accumulating nature of linear dune development favours their preservation relative to more transient and mobile transverse forms (Livingstone and Thomas, 1993).

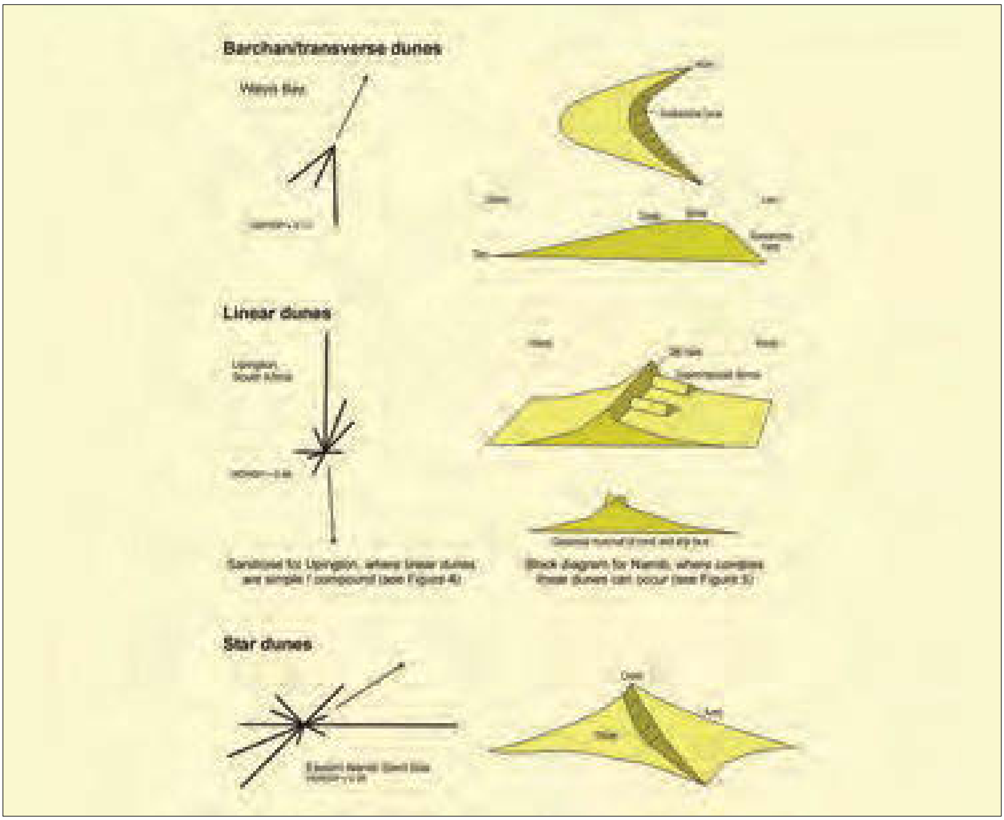

Figure 6.2. Classic dune types found in southern Africa and their associated formative wind regimes. Note that dune forms vary considerably, and "text book" examples are relatively rare compared to the patterns of compound, complex and multigenerational dunes that often occur. Illustrated here are sand transport regimes, in the form of Fryberger (1979) sand roses, for example locations in southern Africa where barchans, linear and star dunes are found.

A simple transverse-linear-star dune classification (Figure 6.2) disguises the actual complexity and diversity of forms that can occur in response to the complexity of wind regimes, both directionally and seasonally, in different places. Beyond the major dunefields of southern Africa, dunes also occur, particularly as a consequence of local sediment supply factors and topographic impacts, for example dunes banked against hills in the Karoo where sediment is sourced from dry channel systems.

2.1 The Kalahari

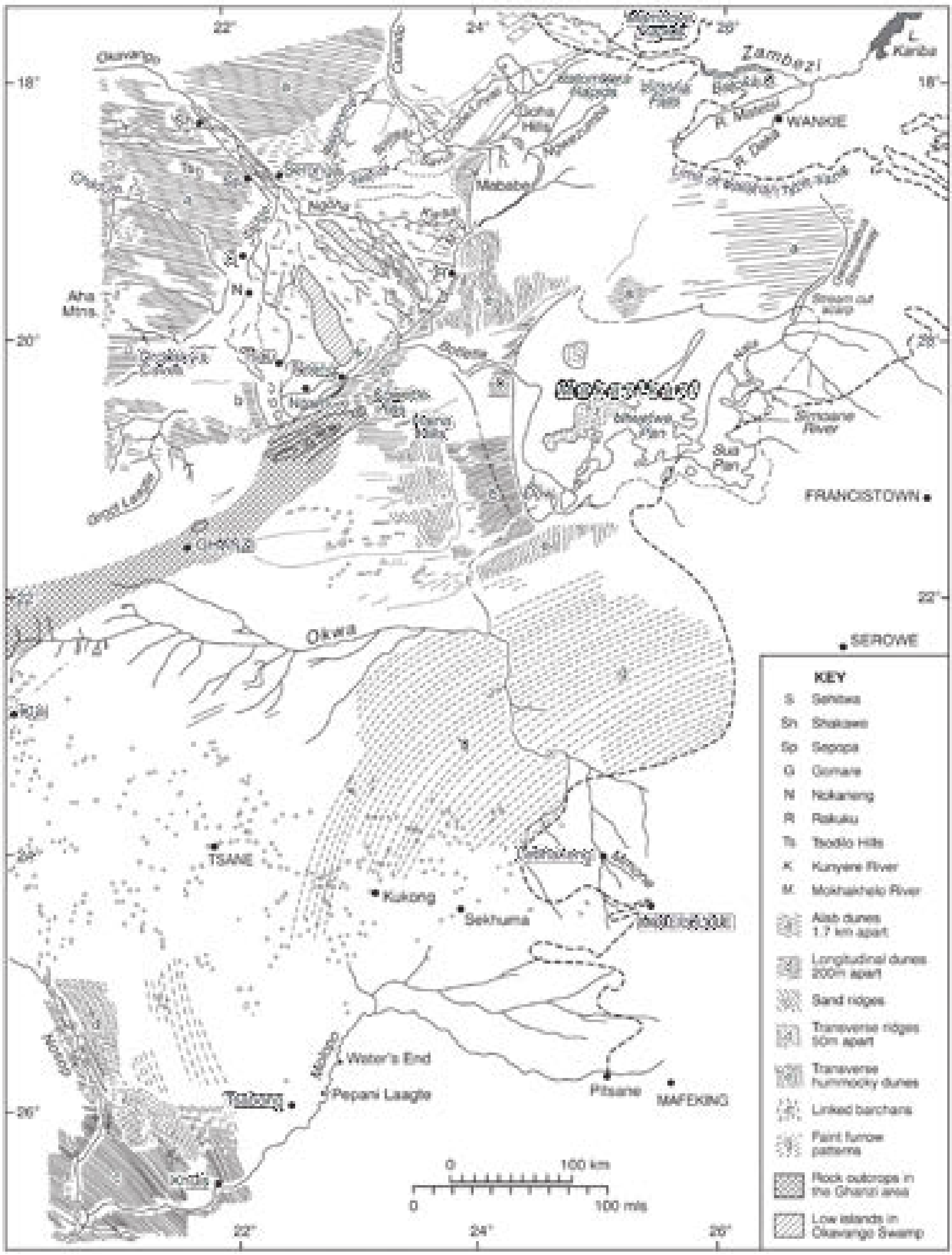

Figure 6.3. Modified version of Grove's (1969) map of Kalahari landforms. The map demonstrates the significance of both aeolian and lacustrine components of this generally very flat landscape.

Mapping from aerial photographs and field reconnaissance led in 1969 to the publication of a paper that provided critical baseline data on the landforms of the then rather remote and difficult to access Kalahari (Grove, 1969). The key elements of this work resulted in a map (Figure 6.3) than demonstrated the complexity of the landscape, mainly comprising depositional features formed in the sandy units of the Kalahari beds. Within this context, several areas of sand dunes were recognised, many comprising fields dominated by linear dunes (called both *alab* and *seif* dunes in the work of Grove 1969), but also including a zone of transverse forms both within and adjacent to the western side of Ntwetwe Pan, within the Makgadikgadi depression. Also mapped were many pan depressions, especially in the southern Kalahari, with lunette dunes on their downwind margins.

This critical paper led to several subsequent studies that focussed on pans, the Makgadikgadi Lake Basin (both discussed in the next chapter) and the linear dunefields. The area of transverse forms to the west of Ntwetwe (in the Central Kalahari Game Reserve) has not been the subject of research effort, while the barchans features on the floor of the pan itself are only now beginning to be analysed (Burrough *et al.*, 2012).

Figure 6.4. Linear dunes in the southwest Kalahari, northern Cape. All photographs were taken within a 30 km radius of Andresvale, Northern Cape, but in different years, illustrating the varieties of linear dune form in the region and the effect of climate variability on dune vegetation and aeolian activity. Clockwise: a) 1984, during a drought period. Low vegetation cover on dunes due to combined effect of drought and ensuing pressurised grazing, resulting in dune crests and flanks being devoid of plant cover and subject to sand transport; b) 1991 during a wet period, with significant vegetation cover on all dune surfaces and no evidence of sand transport; c) 1994, bare rolling linear dune crest; d) 1996, steeper, active upper dune flanks, but vegetated lower slopes.

The Kalahari linear dunes have often been treated as forming three distinct dunefields (Lancaster, 1981; Thomas, 1984): a southern dunefield (Figure 6.4) in the Northern Cape, southwestern Botswana and southeastern Namibia, where dunes are largely grassed but can have bare crests on a seasonal or inter-annual basis, an eastern system centred on Hwange National Park in Zimbabwe where dune ridges are broader and flatter and covered in woodland, and a northern system, found in western Zambia, the Caprivi Strip of Namibia and southeastern Angola, and like the eastern dunes displaying a vegetation cover comprising various woodland and shrub communities. Thomas (1984) should be consulted for discussion of details of dune forms and vegetation communities, and Thomas and Shaw (1991a) for detailed accounts of the forms in each of these systems.

Closer analysis suggests that this tripartite division is a simplification of the situation on the ground. There is much morphological variability within, especially, the southern and northern groups of dunes, and in the case of the northern dunes, there are in fact several distinct zones of ridges each emanating westward (downwind) from major river valleys (the upper Zambezi, Chobe and Okavango Rivers), which suggest that the fluvial systems may have played some role in sediment supply to the aeolian features. In the field, the northern dunes, especially in more easterly areas such as western Zambia, are most readily recognised by vegetation patterns rather than morphological expression. In contrast, dunes northwest of the Okavango Delta are up to 25 m high (Lancaster, 1981). Ridges vary in width in northern and eastern areas from 500-2 000 m, and have wavelengths of up to 2 500 m (Thomas, 1984; Lancaster, 1981). That these are severely degraded features (by processes of runoff and sheet wash) was proposed by Flint and Bond (1968) and more recently reiterated by MacFarlane *et al.* (2005). This implies that the features present today are both lower, and wider, than when these relic dunes were constructed, and that interdunes have received sediments from the dunes themselves, contributing to a progressive levelling out of the landscape since major phases of dune formation.

The features of the southern Kalahari are more readily identifiable as dunes in morphological terms. Ridges are up to 25 m high, 250 m wide and, at the northern end of the dunefield in eastern Namibia, over 2 km apart (Thomas, 1988). Within the region, morphological variability is marked (Bullard, 1994), embracing near-straight, widely spaced dunes in upwind locations through to sinuous, merging and closely spaced features (such that they might almost be termed network dunes) in the centre of the dunefield. Between the Aoub and Nossob Valleys in the Transfrontier Kgalagadi Park, several small patches of distinct parabolic dunes have been shaped out of the linear ridges (Thomas and Shaw, 1991a).

Despite their relative *freshness* compared to Kalahari linear dunes further north, linear dune formation is not occurring today in the southern Kalahari. Though droughts and fire regularly reduce vegetation cover on the dunes of the southern Kalahari (Bullard *et al.*, 1996; Wiggs *et al.*, 1995; Thomas and Leason, 2005), the sand transport that occurs on the dune surfaces is relatively limited and is normally restricted to crestal locations (Livingstone and Thomas, 1993; Wiggs *et al.*, 1995). The dunes of the Kalahari are therefore commonly regarded as features inherited from times past that were more favourable to widespread aeolian action. This assertion creates two issues, considered later in the chapter, that are of central importance in the geomorphology of this part of southern Africa: the role of vegetation in dune activity, and trying to establish when the dunes were in fact constructed

2.2 Namib and western margin dunes

The Namib Sand Sea comprises dunes that demonstrate in their broad distribution and morphology a spatial progression that reflects wind regimes and sources of sediment to the system. Though mineralogical studies show that sand has reached the Namib from both the west and from channels emanating from the eastward Great Escarpment (Besler, 1980, Waldren and White, 1997), and sand transport is affected today by SW and E-NE winds, much of the sand that has built this spectacular sand sea has arrived at its destination through the combined effects of river, sea and wind transport (Lancaster and Ollier, 1983). This is because sediment transported to the Atlantic via the Orange River system is transported northwards by coastal currents that deliver sand to the beaches of the Namib. From here the dominant westerly-southwesterly winds have moved sand landward into the dunefield.

Dunes vary in size in the system, with the largest in more central locations where complex linear/star ridges may be up to 150-300 m high, and spaced at up to 2 km apart (Goudie, 2002). Mean annual sand flow direction shows spatial variation within the dune system of the Namib Sand Sea. On the western littoral 80-90% comes from the SSW-SW sector, decreasing to 55-65% in the interior and 35-40% at the easterly margin, while E-NE *berg* winds, which generally occur intensively for short periods in July-August, account for less than 10% of sand flow at the coast, rising to 30-55% in the interior and 60-65% in the east of the system (Lancaster, 1985; 1989). Overall, therefore, there has been a predominantly eastward movement of sediment, while the interplay and spatial variation in dominance of winds from the western and eastern sectors has left its mark in the broad spatial distribution of dune forms: transverse dunes (including barchans) move sand eastwards from the coast, linear dunes trending northwards dominate where SW and NE winds most effectively interact, and star dunes occur in eastern-central areas where net sand accumulation is highest and wind regimes most complex. This simple spatial arrangement of dunes, though holding true at the large scale, does in fact disguise local complexity to the patterning of dune forms that detailed analysis of satellite imagery reveals (Breed and Grow, 1979; Livingstone *et al.,* 2010).

The interplay between westerly and easterly sector winds can lead to seasonal migration of the crests on linear dunes (Livingstone, 1989). The application of OSL dating to samples taken from cores across a linear dune profile has demonstrated that long term (millennial timescale) there is an imbalance within these seasonal movements (Bristow *et al.,* 2007). Ages from an intensively studied dune in the northern part of the sand sea showed a subtle easterly migration of the linear form over time.

Figure 6.5. Long, high (>80 m) complex linear dunes in the central Namib Sand Sea with small transverse dunes migrating up towards the crest .

The large volume of sediment in this sand sea contributes to some very large, and high, dunes in the Namib, which, at up to 300 m from interdune to crest, makes them amongst the highest on earth (Goudie, 2002). All sand dunes modify airflow over their surface by intruding into the atmospheric boundary layer, but the size of these dunes, and the volume of sediment availability, impacts notably on dune form, giving rise to some of the dune pattern complexity referred to above. In some locations, smaller dunes develop on the flanks of large features, so that, for example, barchan dunes that migrate up the windward flanks of the large linear ridges (Figure 6.5).

The hyper-aridity of the Namib Sand Sea and development of dunes is not a recent phenomenon. The Namib is widely regarded as having been extremely dry for a very long time, perhaps since the Tertiary (Tankard and Rogers, 1978) or earlier (Van Zinderen Bakker, 1978), which may account for the size of dunes, reflecting long periods of sediment supply and accumulation (Vermeesch *et al.*, 2010). Wetter episodes during the Late Quaternary have impacted the dunefield though, supplying fluvial sediments to interdunes that are sourced in the escarpment and which have either terminated in the dunefield or at times extended to the coast (see Lancaster and Teller 1988; and Stone *et al.,* 2010 for discussion). Marine core evidence also suggests several periods of stronger westerly circulation that may have brought moister interludes to the desert (Shi *et al.,* 2001).

The western coast sandveld (area 6, Figure 6.1) including the sometimes-called Namaqualand Sandy Namib (Lancaster, 1989) is less spectacular than the Namib Sand Sea, but is none the less an extensive dune region. This region is a combination of dune systems comprising reticulate/parabolic features up to 4 m high, and sand sheet areas (up to 100 km^2) where thin (subc1m thick) aeolian deposits are of insufficient volume for dunes to develop (Chase and Thomas, 2007). Overall this is regarded as a sediment supply-limited, high energy, aeolian environment where sand transport has been dominated by westerly/southwesterly winds. As humidity increases in a southerly direction, so the aeolian deposits become better vegetated and less active today.

The Skeleton Coast Dunefield (area 7, Figure 6.1) comprises a narrow belt, ranging from 6-22 km in width, of transverse and barchanoid dunes in northwestern Namibia. This coastal erg stretches approximately 150 km between Torra Bay in the south to the Hoarusib River in the north (Krapf *et al.*, 2003) covering an area of approximately 2 000 km^2. The strong, unimodal, onshore SSW winds in this region result in a high energy sand flow from southern and western coastal sources into the lower energy interior (Lancaster, 1985) resulting in the accumulation of dunes reaching up to 50 m in height (Baddock *et al.*, 2007), although average heights of 5-30 m are more common (Lancaster, 1982). In the southern area the dunefield is characterised by relatively low, sinuous barchanoid forms that coalesce on the western edge to produce 30-50 m high sand *wall* running sub-parallel to the South Atlantic coast. Krapf *et al.* (2003) describe how the dunefield effectively dams several E-W trending ephemeral river valleys, particular in the northern (downwind) section of the dunefield where an increased dune-belt width is associated with larger transverse features.

2.3 Other dune forms

Southern Africa has many dune types including dunes that are controlled by topography. Climbing and falling dunes, found locally in the Karoo, adjacent to parts of the western escarpment and to the south of the Kalahari (notably south of the Orange River) occur where windblown sand encounters topographic obstacles, and deposition is facilitated In some instances, these topographically-controlled forms take on the characteristics of sand ramps: features comprised not only of wind-blown sand, but of material derived from the weathering and erosion of the hill against which the accumulation occurs, and sometimes fluvial sediments too. Sand ramps may also be gullied when they intercept runoff from the topographic obstacle. The most developed studies of sand ramps have taken place beyond southern Africa, in the Mojave Desert, California (Tchakerian, 1991; Lancaster and Tchakerian, 1996), and in Iran (Thomas *et al.,* 1997). In southern Africa, Bertram (2003) described and investigated sand ramps in southwest Namibia, between the Tilas Mountains and Aus. In this area two generations of ramps were identified, with windward (climbing, west facing) and leeward (falling, east facing) forms identified both comprising aeolian and fluvial wash units. Provisional chronologies placed the main phase of accumulation between 40-25 ka.

3. Dune activity

The discussion above makes reference to both dunes that are active to the wind today and those that are heavily vegetated and are not. This division is sometimes crudely translated into *active* and *fixed* (or *relic*) dunes. This simple classification attempts to capture the two extremes of dune environmental contexts (Thomas and Shaw, 1991b): locations where surface conditions today are dry, where aeolian transport can operate unhindered by significant vegetation, and those where the vegetation cover prevents all aeolian sediment transport, regardless of the level of wind energy. This *on* or *off* approach to dune activity does not fully capture the situation in many dunefield contexts, no better exemplified than in the southern Kalahari.

Livingstone and Thomas (1993) discussed, in a southern African context, how sand transport potential occurs along an environmental continuum, so that in many places the potential for activity requires more complex consideration, especially in locations where wind energy and precipitation displays marked seasonal and inter-annual variability. These are locations where episodic sand transport is likely to take place, and it is no surprise, given the temporal variability in environmental conditions, that the southern Kalahari should be a region where variable aeolian activity has been heavily investigated (e.g. Bullard *et al.,* 1995, Thomas and Leason, 2005).

3.1 Interactions between controlling variables

Regardless of the level of surface vegetation, Livingstone and Thomas (1993) noted that

> *... often the argument has been, first, that vegetated dunes are fixed relicts, and second, that fixed dunes indicate past aridity [...] neither of these maxims is necessarily true. (Livingstone and Thomas 1993:91).*

We have previously noted that sand transport and dune activity result from the interaction between surface erosivity and erodibility. One of the first studies that aimed to model the interaction between the two and the consequent impacts on dune dynamics was based in the southern Kalahari (Lancaster, 1988). This was an attempt to establish how modern average environmental conditions might differ from those at the time of dune construction.

Lancaster (1988) developed a *Mobility Index* (M), where $M = W/(P/PE)$. W = the amount of time wind blows above a threshold for sand entrainment, and P/PE captures rainfall effectiveness through the relationship of precipitation to potential evapotranspiration. Values of M were calibrated, by examining dunes in different states of activity today, as >200 = fully active dunes, 100-200 = plinths and interdunes vegetated, 50-100 = only activity on dune crests, and <50 = dunes fully inactive. More sophisticated variants of this index have since been produced, for example incorporating rainfall lag effects on soil moisture and rainfall seasonality (Thomas *et al.,* 2005), but importantly, Lancaster's version was used to assess the degree of integrated (erodibility and erosivity) environmental change that had occurred since dunes in the southern Kalahari were formed (Figure 6.6).

The operation of aeolian processes on dunes also varies over much shorter timescales than those considered by Lancaster (1988), and dune activity can be related both to dune construction (which Lancaster, 1988, was concerned with) and the simple movement of sand on dune surfaces, which may not lead to any aggregate changes in dune form. This has been assessed in the southern Kalahari through dune process monitoring studies and through analysis of meteorological and remote sensing data.

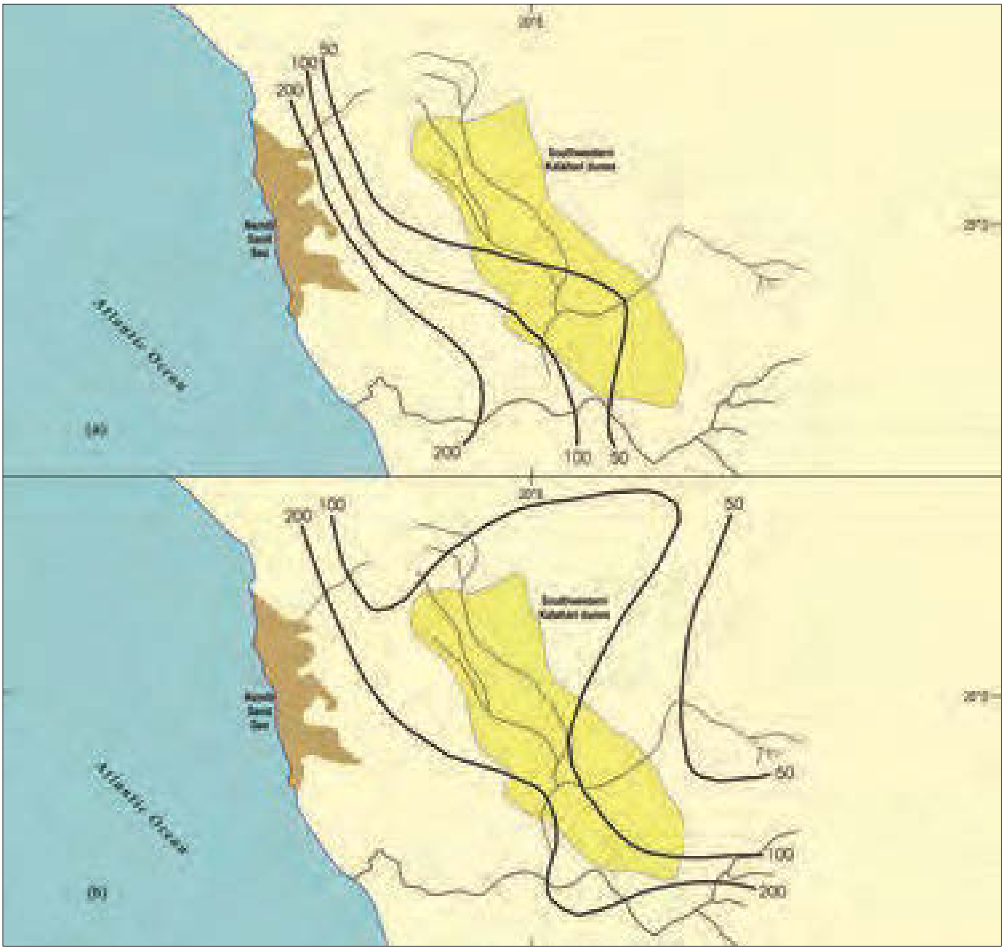

Figure 6.6. Mobility Index values for the southern Kalahari Dunefield today and at a time of assumed dunefield construction. See text for details (after Lancaster, 1988).

A 30-year analysis of the temporal variability of climatic parameters that impact on dune activity in the southern Kalahari was conducted by Bullard *et al.* (1996; 1997). This showed that both erodibility and erosivity elements of Lancaster's (1988) M vary at sub-decadal time scales. Interestingly, drier periods (leading to greater erodibility) tend to correspond with times of enhanced wind energy (erosivity) making the potential for sand transport during drought years significant. Wiggs *et al*. (1994a; 1994b; 1995) conducted surface monitoring studies and how this translates into sand transport via linear dune surface changes. These studies showed that while a single threshold vegetation cover level cannot be identified for all positions across a dune profile, a cover level of about 14% characterises the limit to marked sand transport on dune crests. This cover value was subsequently utilised to classify surface cover variability of an area of linear dunes within satellite imagery by Thomas and Leason (2005) (Figure 6.7), illustrating both the dynamic nature of dunefield vegetation in response to rainfall variability, and the spatial dynamics of potential dune surface activity.

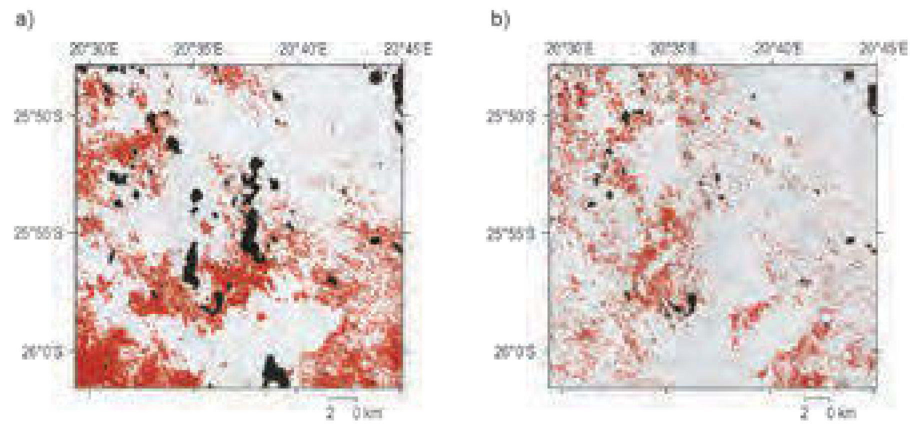

Figure 6.7. Temporal variability in dunefield vegetation cover in response to a) wet and b) drought years. Image classification shows how areas potentially susceptible to aeolian sand transport vary in response to rainfall-driven vegetation cover change. The susceptible areas remaining in the "wet" year analysis are a consequence of livestock grazing impacts on the dune ecosystem (after Thomas and Leason, 2004).

3.2 Dunes in the past

The process studies conducted on southern Kalahari Dunes in the 1990s have pointed towards there being insufficient dune activity and energy today, at least in the monitoring period, to build sand dunes. This is in contrast, for example, to conditions in at least the coastal areas of the Namib where transverse dunes build and transport new sediment in to the heart of the dunefield. In a Kalahari context, therefore, dune construction represents a legacy from past periods when environmental conditions facilitated significant sand transport.

Establishing when the linear dunefields of the Kalahari were constructed has long been of interest. Prior to the advances in OSL dating that now allow the timing of dune sediment accumulation to be determined, Grove (1969) noted that:

> *The morphological evidence [...] is difficult to interpret. No dates can be attached so far to the various features that have been described (Grove 1969:210).*

What was evident however was that many of the aeolian landforms, found both in the driest southern regions and in much wetter locations to the north, were not forming under modern conditions.

Prior to the capacity to directly date dunes using luminescence dating, a number of inventive approaches were taken to estimate the timing and relative ages of development of the three linear dune systems in southern Africa's interior. These focussed upon the relative degrees of morphological degradation to suggest differing periods of time over which dunes had been exposed to post-depositional modification in the three dune systems (Thomas, 1984), and differences in the orientations of the systems used to suggest changes in atmospheric circulation between phases of dune construction (Lancaster, 1981). On these grounds, the southern Kalahari Dunefield was considered to be the youngest, and the northern, most degraded system, the oldest, possibly predating 30 000 ka (Lancaster, 1981).

In 1997 the first OSL ages were published for Kalahari linear dunes, following field sampling in all three-dune systems (Stokes *et al.,* 1997). This study demonstrated for the first time that none of the dunefields were formed in single episodes, since multiple depositional events were preserved in dune bodies in all three systems (Figures 6.8 and 6.9). Further detailed studies in individual components of the southern and northern Cape and Namibia (Blumel *et al.,* 1998; Thomas *et al.,* 1998); eastern Zimbabwe (Stokes *et al.,* 1998); and northern Zambia and Namibia (O'Connor and Thomas, 1999; Thomas *et al.*, 2000) reinforced the evidence for multiple accumulation phases within each of the dunefields, as have analyses of OSL records from the aeolian deposits of the West Coast Dunefield (Chase and Thomas, 2006; 2007). These are discussed in the context of Quaternary climatic changes in southern Africa (see Chapter 12).

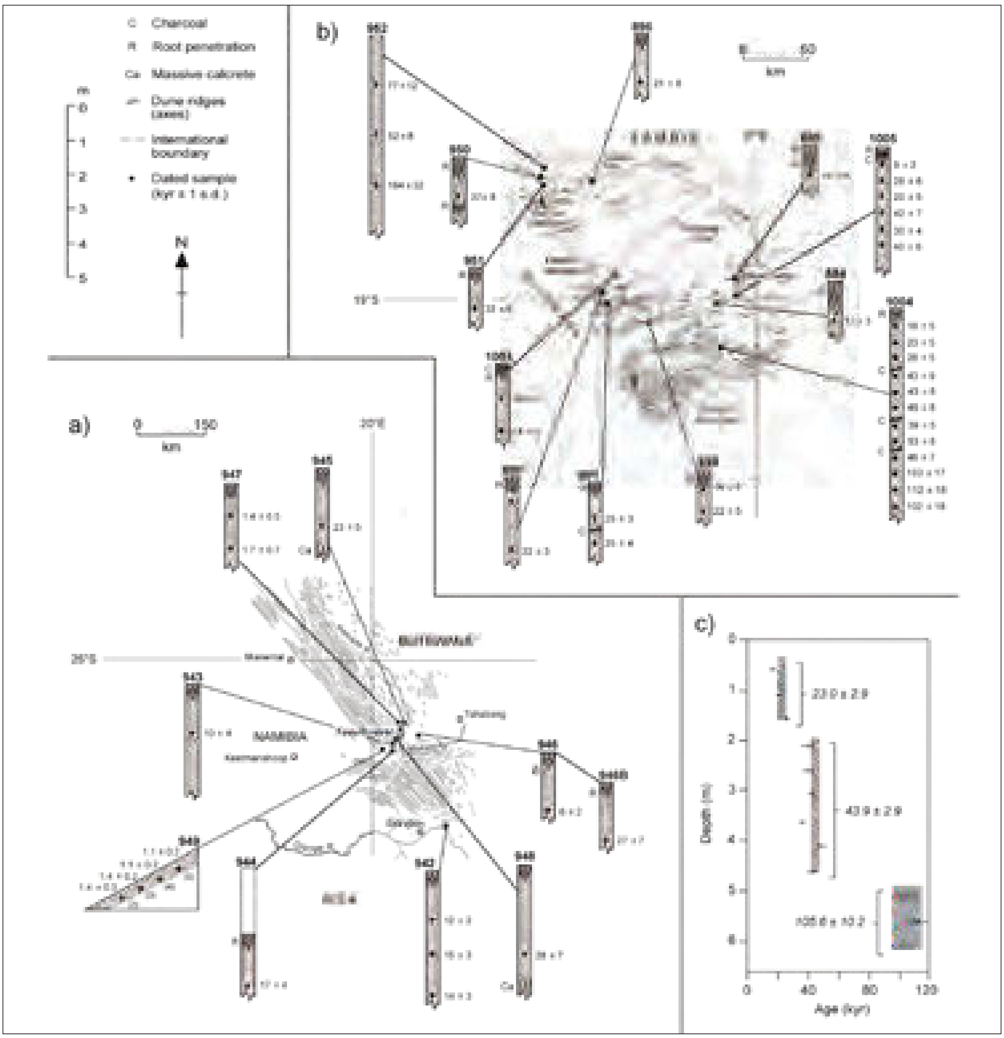

Figure 6.8. Chronology of linear dune deposition in the Kalahari Desert, based on OSL dating of sediment samples, largely collected from hand-dug pits in dune crests. Ages were interpreted as demonstrating discrete periods of dune deposition during the late Quaternary (after Stokes *et al.*, 1997).

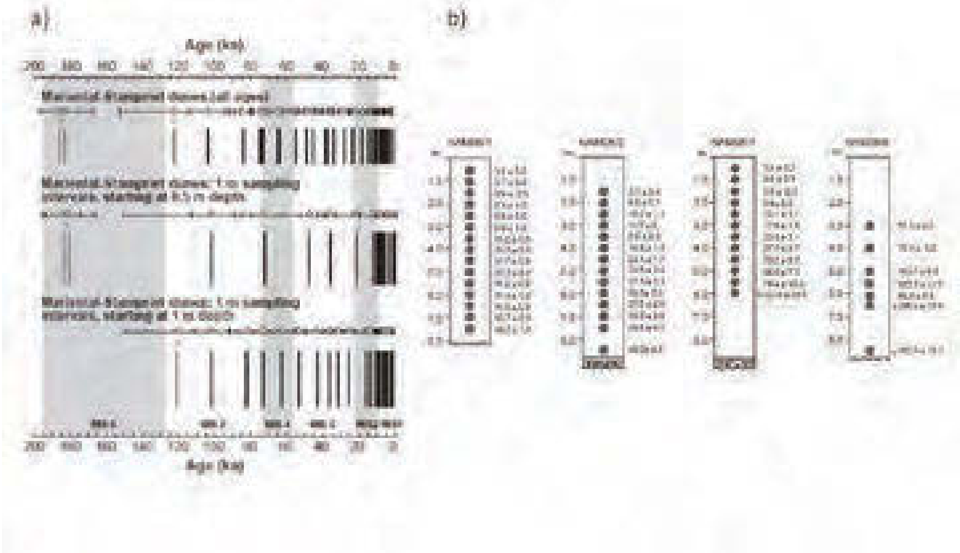

Figure 6.9. a) Chronology of linear dune accumulation in the Stampriet-Mariental area, Namibia, derived from the age data from samples collected and dated using OSL from four sediment cores (shown in b). The top row in a) shows the chronology based on all ages which, when age one sigma errors are taken account of, suggests almost continuous accumulation over c 150 ka. However, if fewer ages are included, as might occur if a single core is dated at e.g. 1 m sample intervals, then alternative dune accumulation records may be deduced. This is shown in the middle and lower rows, which also illustrate how the depth at which samples are first taken impacts on resultant ages, even if the same sampling interval is retained (after Stone and Thomas, 2008).

Dune systems are not however as simple to interpret in palaeoclimatic terms as was once thought (Thomas and Shaw, 1991a), where fixed dunes were simply regarded as a proxy for rainfall changes, with a number of issues arising in recent work (Stone and Thomas, 2008; Chase, 2009). Linear dunes, that dominate the Kalahari record are, with their *extending* mode of development, perhaps the most stable of all major desert dune types even when undergoing sediment transport. This may be one reason why they persist in the geomorphological record, but it also complicates their clear palaeoclimatic interpretation. This is particularly the case in contexts such as the southern Kalahari where dunes are subject to episodic surface activity under the impact of climatic variability, as previously noted.

The palaeoclimatic signal that can be gained from stable dunes, therefore, to some degree depends on their location relative to contemporary environmental conditions (Thomas and Burrough, 2012) with, say, the degraded dunes in humid western Zambia providing clearer evidence of climatic shifts than those in the arid northern Cape. The relative role of surface cover (erodibility) change and wind energy (erosivity) change is currently a hotly debated topic (Thomas *et al.*, 2005; Chase and Brewer 2009) and in such instances other independent proxies of past environmental conditions, such as records of increased windiness from dust accumulations in off-shore marine cores, may be important (e.g. Stuurt *et al.*, 2002). A further issue is that where dunes are heavily degraded, the ages obtained from sampling the forms present today may not provide a simple or complete record of all phases of dune formation.

As more OSL ages have been produced from samples from southern African dunefields, the emerging picture of dune accumulation has become more complex, and is well illustrated by the example of the southern Kalahari. In 1997 seven OSL ages were published for linear dunes from this region (Stokes *et al.,* 1997). These suggested that the southern linear dunefield had accumulated in two periods. By 2003, studies from Blumel *et al.* (1998) and Bateman *et al.* (2003) had added 11 new ages, extending the period over which accumulation was recorded to include the Holocene. By 2007, a total of 88 ages had been published (Telfer and Thomas, 2007). Many fell within the period covered by the earlier studies but, because deep drilling of dune profiles was now contributed to sampling, these also extended the period during which dune accumulation was recorded to over 110 ka.

A further 48 OSL ages, this time determined from samples collected from four full-dune profile dune cores in the eastern Namibian Kalahari, were produced by Stone and Thomas (2008). Many of these ages fell within the previously recorded 30 ka accumulation period, as well as back to approximately 110 ka, but two further ages also representing deposition in the 170-200 ka period. With more published ages, the discrete phases of accumulation that were present in initial studies have disappeared from the record, with apparent continuous deposition represented by the full set of age determinations (Stone and Thomas, 2008; Figure 6.9). Several factors could contribute to this situation including the error range on ages, possible post-depositional sediment mixing, and the complex suite of factors that affect dune activity and sediment preservation, but until these issues are resolved it means that unravelling dunefield accumulation histories, and the environmental changes responsible for dune activity, remain difficult to achieve.

4. Dust in southern Africa

Sources of aeolian dust are often found in topographic depressions where silt-sized material derived from weathering and abrasion of sediment is concentrated by the action of water and, once dried, acted on by erosive winds (Washington and Wiggs, 2010). Such sources may include ephemeral lakes and pans, alluvial surfaces, floodplains and dry river valleys (Bryant *et al.*, 2007; Prospero *et al.*, 2002). Whilst such potential source areas are well represented in southern Africa, the amount of dust produced is relatively small when compared to other source areas, particularly in the northern hemisphere. Long-term satellite data of the Aerosol Index (AI) from the Total Ozone Mapping Spectrometer (TOMS) identifies two significant sources in southern Africa; the closed basins of the Makgadikgadi Pans in Botswana and the Etosha Pan in northern Namibia (Washington *et al.*, 2003; Table 6.1; Figure 6.1).

Table 6.1. Ranked data for maximum mean Aerosol Index (AI) Values for major global dust sources determined from TOMS (after Washington ***et al.,*** 2003).

LOCATION	MEAN AI VALUE	
Bodélé Depression of south central Sahara	> 3.0	
West Sahara in Mali and Mauritania	> 2.4	
Arabia (southern Oman/Saudi border)		> 2.1
Eastern Sahara (Libya)	> 1.5	
Southwest Asia (Makran coast)		> 1.2
Taklamakan/Tarim Basin	> 1.1	
Etosha Pan (Namibia)	**> 1.1**	

LOCATION	MEAN AI VALUE	
Lake Eyre Basin (Australia)	>1.1	
Makgadikgadi Basin (Botswana)	**>0.8**	
Salar de Uyuni (Bolivia)		>0.7
Great Basin of the United States		>0.5

Etosha pan is a 5 000-6 000 km^2 salt lake that occupies the sump of a larger basin draining the interior of northern Namibia and southern Angola. The pan is fed by a large and complex network of ephemeral channels (oshanas) entering from the north (see Thomas and Shaw, Chapter 7) that can inundate the pan surface in the wet season (Bryant *et al*., 2007). The Makgadikgadi Pans also cover an area of approximately 6 000 km^2 within a much larger basin system with river systems draining into it from southern Angola, the Okavango Delta and Zimbabwe. The pan system is divided into two with the westerly Ntwetwe Pan being slightly larger than Sua Pan to the east (Thomas and Shaw, 1991a). Satellite analysis by Vickery (2010) reveals a very strong seasonality in the emission of dust from these two major sources with the majority of deflation evident between May and October. Further, Vickery (2010) discovered that specific areas within each of the pans appeared far more susceptible to erosion. Such a spatial inhomogeneity was particular evident for Sua and Ntwetwe Pans where four dominant emissive regions on the northern and southern margins were evident.

The complex dynamics of emission of dust from such source *hot spots* is poorly understood, but the production of suitable material is commonly related to local groundwater depth and salinity as well as rainfall amounts and rates of evaporation. All of these influence the accumulation of erodible surface evaporite minerals (Reynolds *et al*., 2007) and the formation and strength of surface crusts. Mahowald *et al*. (2003) and Bryant *et al*. (2007) have also noted that dust emission from Etosha Pan and the Makgadikgadi Pans is significantly controlled by the supply of erodible material provided by intermittent summer flood inundation from their contributing ephemeral river flows. In this way, winter dust emission becomes more intense in years following large flood inundations. Bryant *et al*. (2007) hypothesise that the amount of dust emanating from these sources may therefore be linked to changing rainfall patterns driven by the El Niño-Southern Oscillation (ENSO).

A further significant regional dust source in southern Africa has been found to be the ephemeral river valleys of the west coast of Namibia (Eckardt and Kuring, 2005). Vickery (2010) describes more than 12 ephemeral river catchments in western Namibia ranging in size from 2 000-30 000 km^2. Sediment supplied from the wetter headwaters of these channels provides material prone to deflation in the drier western parts of the catchments as easterly winds are channelled into the river valleys (Figure 6.10). Analysis of satellite data by Vickery (2010) revealed that the 200 km of coastline between the Hoanib and Ugab rivers produces the highest density of dust plumes whilst the Kuiseb River Valley is the most significant single contributor.

Additional aeolian dust is also derived from the dry pans of the Northern Cape (area D in Figure 6.1). The >500 small (10-15 km^2) pans prevalent in this semi-arid region and interspersed between the dune ridges show little evidence of surface drainage and so the dynamics and susceptibility of their surfaces to wind erosion are assumed to be controlled by changes in groundwater levels (Lancaster, 1986). Hakskeenpan, covering an area of approximately 140 km^2 in the northwest of this region, has been identified by Vickery (2010) as the most consistent and significant dust source in the Northern Cape.

Figure 6.10. Dust plumes sourced from ephemeral river channels off the west coast of Namibia (NASA, 2010).

Erosion of dust from these sediment sources results from the migration of anticyclones that enhance the prevailing easterly trade winds across the region. These enable the transport of aeolian dust in an anti-clockwise circulation across southern Africa and out into the Indian Ocean and also provide a westerly transport pathway into the Atlantic Ocean off the west coast of Namibia (Figure 6.11). Up to 54% of aerosols are re-circulated at least one complete cycle before exiting the region (Tyson *et al.*, 1996).

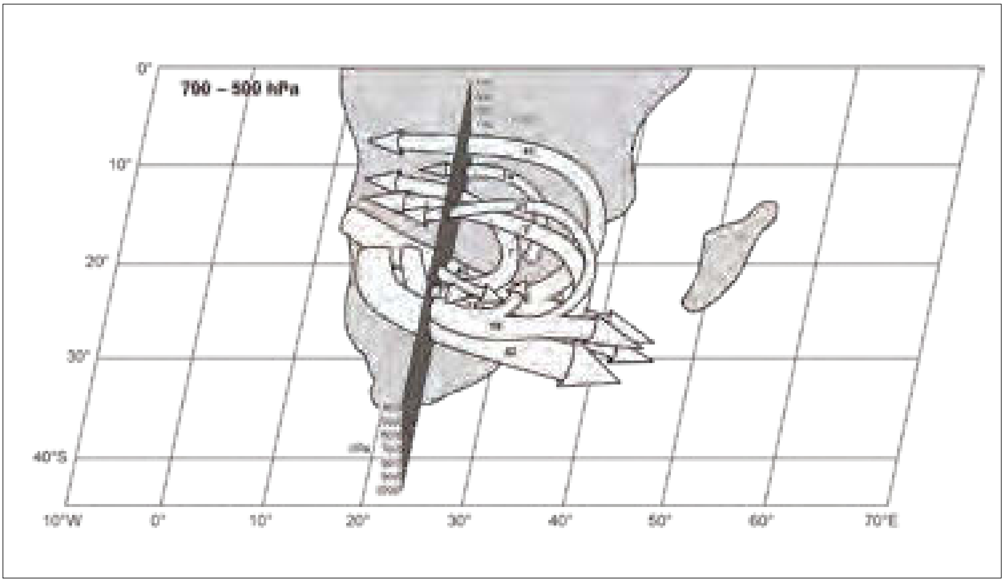

Figure 6.11. Mean atmospheric transport pathways for aerosols over southern Africa (after Tyson *et al*., 1996).

A further influence on aeolian dust activity in southern Africa is human activity, which can have a profound influence on generating new dust sources (Wiggs, 2011). Most significant in the southern African context is the disruption of stable surfaces by agricultural activity on susceptible soils. In particular, the west-central Free State (with typically <500 mm of rainfall) supports widespread commercial agriculture, but also suffers from significant periods of dust emission from bare fields in the dry winter months (Wiggs and Holmes, 2010). Whilst conservation strategies can be employed in agricultural production to limit the susceptibility of soil to wind erosion to some degree, there is concern that climatically driven environmental change may increase the wind erosion hazard from these human impacted sources in the future.

5. Aeolian futures

Notwithstanding some of the current issues in interpreting chronologies of late Quaternary dune development in southern Africa, the records that exist show how dynamics, and potentially sensitive, dune systems can be to climate change. Dust generation may be equally susceptible, but without dated chronologies of long-term dust flux this is presently hard to confirm. However, the fact that the main basins that emit dust today (Etosha and Makgadikgadi) have held large lakes on multiple occasions in the Late Quaternary (see Thomas and Shaw, Chapter 7) shows how these source areas are susceptible to climate and environmental change impacts.

By combining our knowledge of how aeolian sand transport relates to wind and precipitation regimes, mediated through the impact of variable vegetation covers, the potential behaviour of the extensive dune systems of central southern Africa has been modelled by Thomas *et al*. (2005). Extending from northern southern Africa to Angola and Zambia (Figure 6.2), these dunes are, with the exception of those in the southwestern Kalahari (Figure 6.3) currently stable due to low erosivity and well-developed vegetation cover. A modified version of the mobility index of Lancaster (1988; Figure 6.6) was developed to take account of climate seasonality such that:

$$A_{p,\ GCM} = \bar{U}_3/(P_{lag}/E_{p,\ lag} + P_{rainy}/E_{p,\ rainy})$$

where:

$\bar{U}_3$ = the cube of the mean wind speed

$P_{lag}/E_{p,\ lag}$ is the residual effect of recent rainfall and potential evaporation, such that $P_{lag} = (P_{-1} + P_0)/2$, where P_{-1} is precipitation in the previous month and P_0 is rainfall in the current month,

$E_{p,\ lag} = (E_{p,\ -1} + E_{p,\ 0})/2$, where $E_{p,\ -1}$ is potential evapotranspiration in the previous month and $E_{p,0}$ is potential evapotranspiration in the current month.

$P_{rainy}/E_{p,\ rainy}$ is the effect of rainy season precipitation and potential evaporation on soil moisture, such that $P_{rainy} = (P_N + P_D + P_{J...})/m$ and $E_{p,\ rainy} = (E_{p,\ N} + E_{p,\ D} + E_{p,\ J...})/m$, where m = N (November), D (December), J (January), and so on is the month under consideration within the rainy season.

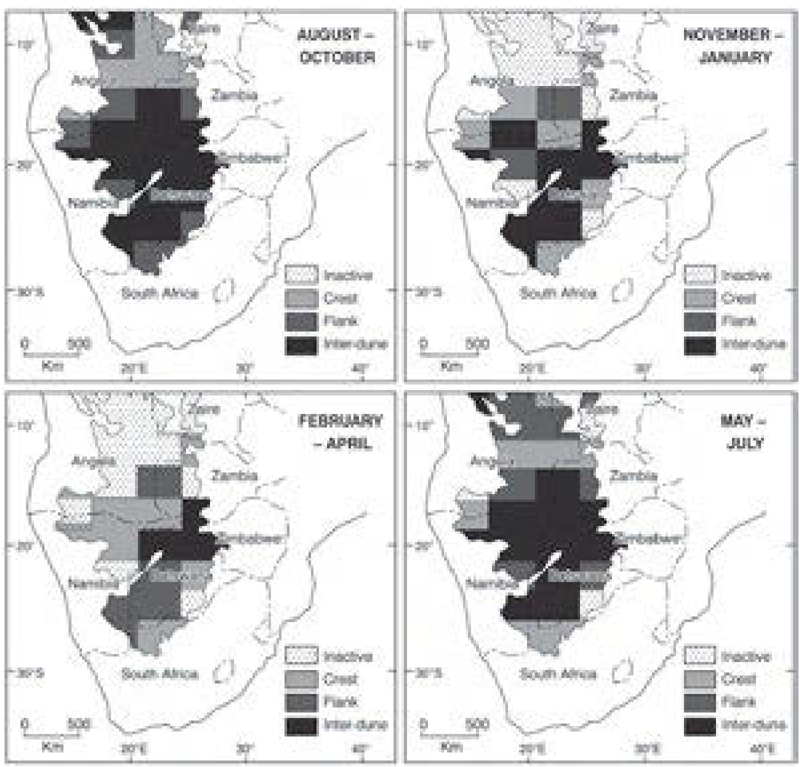

Figure 6.12. Modelled Kalahari dune system activity, expressed in three-monthly time slices, averaged for the period 2070-99. Shading differences suggest the degree of dune landscape activity through the seasons. "Interdune" implies aeolian activity occurs through the whole dune landscape: i.e. that it is not restricted to dune crests and flanks. "Flank" refers to activity on dune slopes and crests, and "crests" refers to activity on dune crests only (after Thomas *et al*., 2005).

The model was then driven with monthly climate data outputs from three GCMs for the twenty-first century, and calibrated Ap, GCM values used to establish how much of dune landscapes (crest, flanks, interdunes) would be susceptible to aeolian transport (Figure 6.12). The results suggest that regardless of the degree of climate change provided by the three GCMs, significant dune reactivation may occur by the end of the twenty-first century throughout the southern African interior. Whether these changes are borne out will depend on the actual changes in climate that occur in the future and the reliability of the dune mobility model, but this study does suggest that the aeolian landscapes of southern Africa have the potential to be dynamic as the next decades progress.

6. Conclusion

The contribution of aeolian processes to the geomorphology of southern Africa has been long known. Today, we are gaining a nuanced, but still far from complete, understanding of how aeolian processes have operated in the past, why they occur today, and how their occurrence and rates of operation may change in the future. Quantified studies, of the timing of dune development, rates of sand movement on different surfaces, and of dust emission, provide the data that informs the capacity to look back in time and aeolian dynamics and forward to the possible changes driven by global warming impacts. The systematic knowledge that has developed in the past twenty years, through the application of luminescence dating and conduct of reductionist process studies, has taken this understanding to levels not envisaged forty or fifty years ago. The Kalahari and Namib have been key areas for processes studies that have impacted on aeolian geomorphology globally, and the same can be said for the luminescence dating applications conducted in the subcontinent. The Kalahari perhaps possesses the most completely dated dune systems anywhere on earth and from these studies has come greater knowledge – but also many unanswered questions. There remains much still to understand about these fascinating landscapes.

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7

Terminal Basins: Lacustrine and Pan Systems

Terminal Basins: Lacustrine and Pan Systems

David SG Thomas and Paul A Shaw

1. Introduction

In the previous chapter, the contribution of aeolian processes to the development of southern African landscapes was discussed, in the context of the aeolian systems that are active today or have been in the past. Issues of interpreting some of these systems in terms of past aridity manifest themselves in the debates surrounding dune systems that are inactive, or only displaying limited activity, today. At an apparent opposite end of humid-arid transitions in the Quaternary period are a suite of landforms that suggest much wetter past climates within the subcontinent's dry or highly seasonal interior. At one extreme are the large palaeolake basins of the Makgadikgadi and Etosha systems, complexes of ephemeral lakes and saline lake flats, which, although termed *pans* should be described as playas or salt lakes. At the other are the so-called pan depressions, found in a wide range of current climatic regimes in southern Africa. In some areas, such as the southern Kalahari and parts of the Free State Province of South Africa, they are one of the most dominant and distinctive landscape units present (Figure 7.1). Though there has been significant research effort invested in southern Africa's palaeolakes and pans since their geographic distribution was so effectively mapped by Grove (1969), there remain many uncertainties regarding the modes and timings of their active development. This chapter discusses palaeolakes and pans in terms of their distributions, modes of development, controlling variables, and the most recent research that has attempted to establish the age and frequency of the development of these hydrologically-important components of the southern African landscape. Collectively, these lake basins and pans might all be termed *terminal basins* since a common characteristic of most is the absence of surface outflows.

As with dune systems, it was the application of optically stimulated luminescence (OSL) dating during the 1990s that gave new impetus towards resolving many of the uncertainties surrounding lake and pan development, but as is often the case, new research tools not only offer the potential to answer old questions, but have the potential to create new questions that generate further challenges to unravelling the geomorphological development of arid regions.

1.1 A short research history

Grove (1969) provided insights to the complex geomorphology of the Kalahari that in many regards captured two simple questions: what are the environmental controls of landform development, and when did that development occur. The author of that paper would be the first to acknowledge that the research tools available to him could not provide the required answers, but what he did achieve was a systematic analysis that provided suggestions towards the relationships between features regarded as indicating the importance of both wetter and drier conditions. The enigmas presented by dunes, lake and pan features had been recognised earlier by others, for example Allison (1899), Passarge (1911), Wellington (1943; 1955), and Flint and Bond (1968), but it was not until the late 1970s that a new era of systematic investigations, combining detailed fieldwork, mapping and analysis began, leading ultimately to a situation where it is recognised today that description alone will not generate successful conclusions to complex arid environmental research questions, with successful investigation requiring account of chronology and process, and due consideration given to other factors of importance including geochemistry and tectonics.

Figure 7.1. Pans in the southern Kalahari Dunefield, Northern Cape Province of South Africa stand out due to their dry, salt crusted, surfaces. November 2006.

For the lake basins of southern Africa, critical contributions of this period began with the work of Cooke (1976; 1979; 1980), Cooke and Verstappen (1984), Heine (1978) and Shaw (1985a; 1985b), focussing on the Makgadikgadi system. Precise mapping of shorelines in a pre-digital terrain model era, attempts at early chronologies of hydrological change (based problematically on radiocarbon) and recognition of the potential for subtle tectonic movements to alter hydrological configurations in the Makgadikgadi resulted. For pans, the key work was that of Lancaster in the Kalahari (Lancaster 1978a; 1978b; 1979). This work viewed pans with accompanying fringing lunette dunes as bearing evidence of conditions both wetter and drier than today. There followed a phase of pan-mapping sub-continent wide (Goudie and Thomas, 1985, subsequently reiterated in global surveys by Goudie, 1991; and Goudie and Wells, 1995) in a quest to establish broad factors controlling the distribution of these depressions. Such studies all laid foundations for placing interpretation of the southern African palaeolakes and pans in a context that could draw on relevant work from other continents, showing that uniqueness was not really an attribute of these features, and that models of development had to embrace the role of climatic changes in the Quaternary, and required a strong chronometric foothold to be able to do so.

2. Basins along a hydrological gradient

In many regards, however, it is research from Australia in the 1980s that best provides a framework for interpreting the geomorphic identities of lake basins and pans in southern Africa. Lake Eyre, in the Australian arid zone, and the many pans (more usually termed *playas* in Australia) found across that island from drier to wetter locations, were the focus of intensive investigation in the 1980s as part of the Salt Lakes, Evaporites and Aeolian Deposits (SLEADS) research programme. This contributed to development of a hydrological model to explain the variations in sediments and forms found in terminal basins (Bowler 1986), which also provide a useful framework for the analysis of southern African basins that follows (Figure 7.2). This presents basins along a continuum, from permanently full of water to almost permanently dry, showing likely surface sediment geochemistry as well as illustrative morphological expressions of basin forms. While southern African basins lack the level of geochemical analyses of their Australian counterparts, the adaptation of the diagram to a southern African context gives examples of basins from this region that fall along the wet-to-dry continuum (Figure 7.2). One geomorphological element that this diagram critically displays is the potential relationship between lacustrine and aeolian processes in shaping basins.

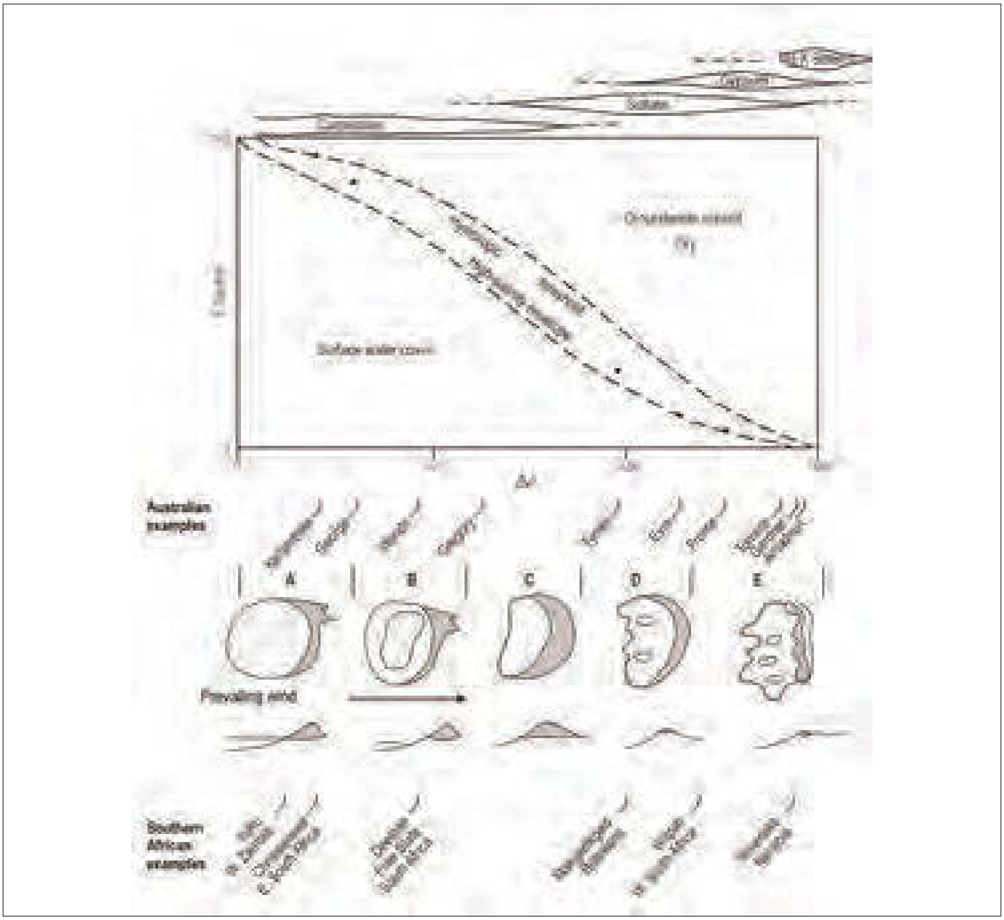

Figure 7.2. Terminal basins of southern Africa. This diagram is an adaptation of a figure produced by Bowler (1985) to explain the hydrological and geomorphological contexts of terminal basins in Australia. The classification of basins is based upon the interplay today of surface water presence and dry conditions (groundwater control) as expressed through a Disequilibrium Index (how far a basin deviates from conditions needed to give rise to permanent water in the basin). The original figure includes representation of basin geochemistry (this is done hesitantly in the context of southern African basins) as well as form. This version retains the names of Australian example basins and adds basins for southern Africa. Some of these examples are estimates, but Makgadikgadi is included on the basis of the Disequilibrium Index value of −298 calculated by Burrough and Thomas (2009a), and Silverlake Playas (Namib) on basis of geochemical information in Eckardt *et al*. (2001).

3. Shorelines and lunettes

The largely dry (*senso* recent decadal-century scale) lake basins of southern Africa have received much of their analysis in terms of the fossil shoreline systems present on downwind (ca. westerly) margins (e.g. Cooke, 1979; Shaw, 1988; Burrough *et al.,* 2009a). A focus in the analysis of pans has been the lunette dunes (Hills, 1940) also found on downwind margins of many basins (e.g. Lancaster, 1978a; Goudie and Thomas, 1986). Their development in the context of basin hydrology has been somewhat difficult to unravel, particularly as to whether they are a function of deflation from dry pan floors or deflation from beaches formed by wave action in wet pans (Bowler, 1973; Lancaster, 1978a). The concept of a continuum of forms and process-interactions across a hydraulic gradient, embodied in Figure 7.2, does away with the need to compartmentalise basins – or the processes operating in them – into wet or dry. Although the continuum of basin forms in Figure 7.2 was developed by analysis of Australian playas across the hydraulic gradient of today, the diagram can equally effectively represent the ranges of process-form relationships that can shape a basin as environments experience climatic changes that impact on basin hydrological status.

For southern Africa, there are two issues concerning basin development that require particular consideration. First is that lunette dunes may not, contrary to widely held views, be best developed in conditions of substantially negative hydrological balance. This is because sediment deflated from a basin may not remain anchored at the basin margin under very dry conditions, due to a lack of sediment-trapping vegetation. Thus, lunettes in the Australian model are portrayed as best developed in wetter conditions, especially when their sediments include a clay-pellet fraction. Second is that a clear distinction may not always exist between shorelines and lunettes, i.e. there may be aeolian contributions to lake shorelines that are primarily a function of wave action. This issue is explored in detail in the context of the Makgadikgadi system, but with wider applicability by Burrough and Thomas (2009). Aeolian action may contribute sediment to a lacustrine beach ridge during both ridge constructional phases (when a lake is present in a basin and wave action is forming a shoreline) and during post ridge formation and lake regression phases. In some contexts sedimentary variations may allow wave and wind elements of a ridge system to be distinguished, but Otvos (1999; 2000) has shown that this may be difficult to achieve. In the Kalahari, where many depositional landforms have developed from sandy sediments with little systematic sedimentary variation, this is also the case (Burrough and Thomas, 2009).

As with the linear dune systems discussed in the previous chapter, prior to the direct age determination of lunette and shoreline sediments using OSL dating, it was commonly regarded that individual lunette dunes and beach ridges were formed in single episodes. For example Lancaster (1978a) interpreted the two lunette dunes found on the downwind margins of several southern Botswana pans as each representing a distinct period of pan excavation by the wind (Figure 7.3). Similarly, the multiple beach ridges found on the western margins of the individual basins within the Makgadikgadi system were ascribed formative events, correlated between basins based on altitude (e.g. Shaw, 1988), and used to determine system-wide lake stages (Figure 7.3).

OSL dating of southern African lunette dunes has shown that often they have developed during multiple accumulation events (Thomas *et al.,* 2002, Holmes *et al.,* 2008) that may include complex lateral variability along dunes (Telfer and Thomas, 2007). Shorelines also display evidence of multiple accumulation events, with up to seven recorded in the Makgadikgadi system in the last approximately 100 ka (Burrough *et al.,* 2009b). However, interpretation is complicated by the role of ridge migration during shoreline development (Burrough and Thomas, 2009; Burrough *et al.,* 2007). This can contribute to erosion as well as accumulation of shorelines and may impact on the potential of shorelines to fully preserve all the lake high stages that a basin has experienced in the late Quaternary. Not surprisingly, therefore, as more data become available, so the records of terminal basin histories are becoming more complex than previously assumed. The sections that follow now examine the nature of lake systems and pans in southern Africa, their geomorphology and developmental histories.

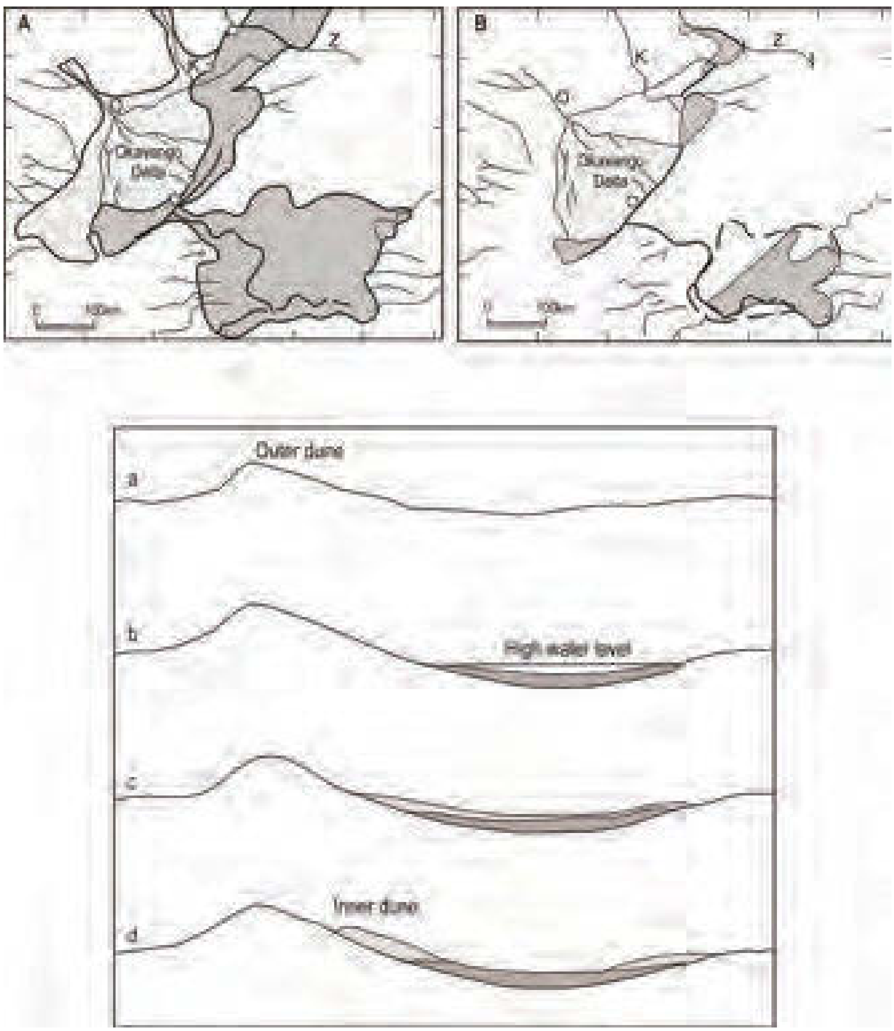

Figure 7.3. Early models of southern African shoreline and lunette dune development. i) Shaw's (1988) model of Makgadikgadi Lake stage development based on the altitudinal correlation of lake shoreline ridges across sub-basins, giving rise to recognition of A) *Lake Palaeo-Makgadikgadi Stage* and B) *Lake Thamalakane Stage*. ii) Lancaster's (1978a) model of multiple lunette development on Kalahari Pan margins. a) Outer lunette dune deflated. b) Deposition of clay sediments in wet pan and c) sandy sediments in drying pan. d) Deflation of sandy-clay mix to form inner lunette dune.

4. Lake basins

The principal lake basins of the southern African interior are the Makgadikgadi complex of Botswana and Etosha in Namibia. Etosha has received relatively little lacustrine system research relative to Makgadikgadi, with the recent papers of Brook *et al*. (2007; 2010) and Hipondoka *et al*. (2005) following work by Smith and Mason (1991), Buch *et al*. (1992) and Rust (1984). Makgadikgadi was more systematically investigated in the 1970s and 1980s as noted above, with further detailed geochemical (Ringrose *et al*., 1999; 2005; 2009) and palaeoenvironmental studies in recent years (Burrough *et al*., 2007; 2009a; 2009b; Burrough and Thomas, 2008).

Both these basins have some element of structural control impacting on their existence. Tectonism has affected the development of the Kalahari from the outset at the division of Gondwanaland, given it is located in a downwarped interior basin. Tectonic activity in the western extension of the East Africa rift, which shapes the distal end of the Okavango Delta, has undoubtedly also affected the further development and subsidence of systems and drainage patterns in the region (Thomas and Shaw, 1991; Ringrose *et al*., 2008). Nonetheless, as Burrough *et al*. (2009) note, while tectonic movements cannot be ruled out as impacting on the geologically-recent history of the lake basin, there is currently no evidence to suggest that it has been the major control on lake basin dynamics and fluctuations in the Quaternary period.

4.1 Makgadikgadi

The Makgadikgadi system today comprises three main sub-basins: Ngami, Mababe and Makgadikgadi, the latter consisting of two major pans, Sua and Ntwetwe (Figure 7.4). Dated sedimentary evidence indicates that during wetter periods (i.e. greater positive hydrological budgets than today: Burrough *et al*., 2009b), these basins formed the lowest parts of a complex lake system that at its maximum extent would have formed a single water body embracing all three palaeolake basins. Collectively this is known by various names in the literature, but Megalake Makgadikgadi (Burrough *et al*., 2009b) would have covered an area of over 60 000 km^2; approximately the size of Lake Victoria today. Recent remote sensing investigations of the area west of the Gidikwe Ridge (e.g. McFarlane and Eckardt, 2008) suggest that the system may, at some unknown time, have been even larger, though this hypothesis requires further detailed investigations before it can be substantiated.

The basin fringing shorelines of Makgadikgadi, mapped through the analysis of imagery and careful field surveying, occur at a range of altitudes from approximately 912-945 m a.s.l. OSL dating, including sampling via deep drilling of shoreline ridges within the system has provided a record of high water levels within the system as a whole and within the sub-basin components (Figure 7.5). Other studies have utilised diatom analysis to investigate past hydrological conditions, particularly within the extensive Ngami diatom beds (Shaw *et al*., 2003) and exposures along the Boteti River component of Makgadikgadi (Shaw *et al*., 1997). Burrough *et al*. (2009b) have considered the dynamics of basin filling.

Makgadikgadi is the end point of an allogenic and endoreic hydrological system ranging over 12° of latitude, extending from the Angolan highlands at approximately 12°S, where tributaries of the Okavango, and other major south-flowing rivers are sourced, to the southern margins of the Makgadikgadi Depression at 24°S. This catchment also ranges over approximately 10° of longitude (Figure 7.4). Close to the source region annual rainfall varies between 800 and 1 400 mm/yr^{-1}, while in the sump of Makgadikgadi today it is approximately 400 mm/yr^{-1}. Annual evaporation rates are approximately 1 900 mm in headwaters to over seven times precipitation amounts in the basin itself. These rivers reach the Okavango-Zambezi graben and in wet years (e.g. 2009 and 2010) flow from the distal end of the Okavango Delta, via the Boteti River, can reach the margins of the depression sump.

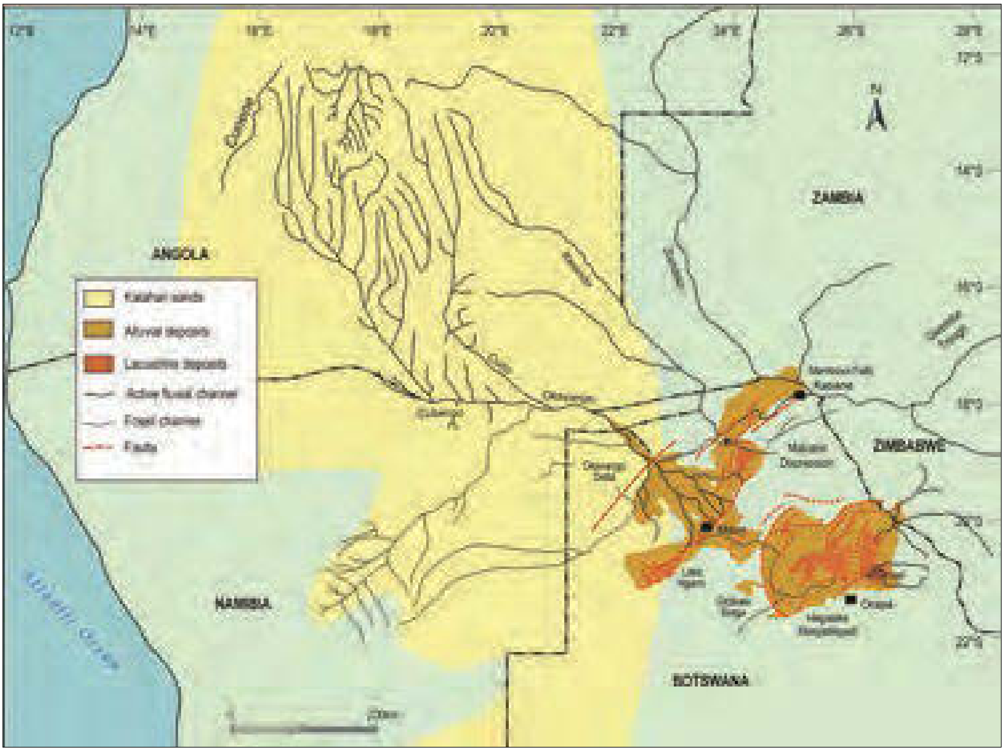

Figure 7.4. The geomorphology of the Makgadikgadi system, including sub-basins (Ngami, Mababe, Makgadikgadi) of Megalake Makgadikgadi. The lake system relationship to its major catchment system to the north is shown. Data from various sources including Burrough *et al.* (2007).

It has been hypothesised that an increase in outflow to the basin from the Okavango system alone would be insufficient to fill the megalake basin. Grove (1969) and Shaw (1985b) proposed that inflow from the Zambezi River would also be required to fill the basin. This is geomorphologically feasible due to basalt-bar constrictions on the Zambezi (and Chobe) diverting flow westwards towards the basin during modern wet years. Connections occur via linkages between the Zambezi, Mababe Depression and thence to the Okavango and Makgadikgadi (Shaw and Thomas, 1988; Burrough and Thomas, 2008). Subtle tectonic movements in the graben system may have contributed to flow redirection in the past, but given there is evidence for multiple phases of basin occupancy by a substantial water body, tectonics without climatic change are an improbable cause of Megalake Makgadikgadi.

There is a further consideration that may contribute to the development and then maintenance of a mega lake of sufficient size and duration to build substantial marginal shoreline features. Climate modelling (Burrough *et al.,* 2009b) indicates that a full lake basin would alter the local and regional climatic-hydrological system sufficiently to increase rainfall, allowing the system to be self-maintaining beyond the initial forcing mechanism of increased rainfall in its distant catchment. Clearly this is, in hydrological, tectonic and climatic terms, a complex system to interpret, but recent advances are starting to provide data that can contribute to a greater understanding of how this system has evolved.

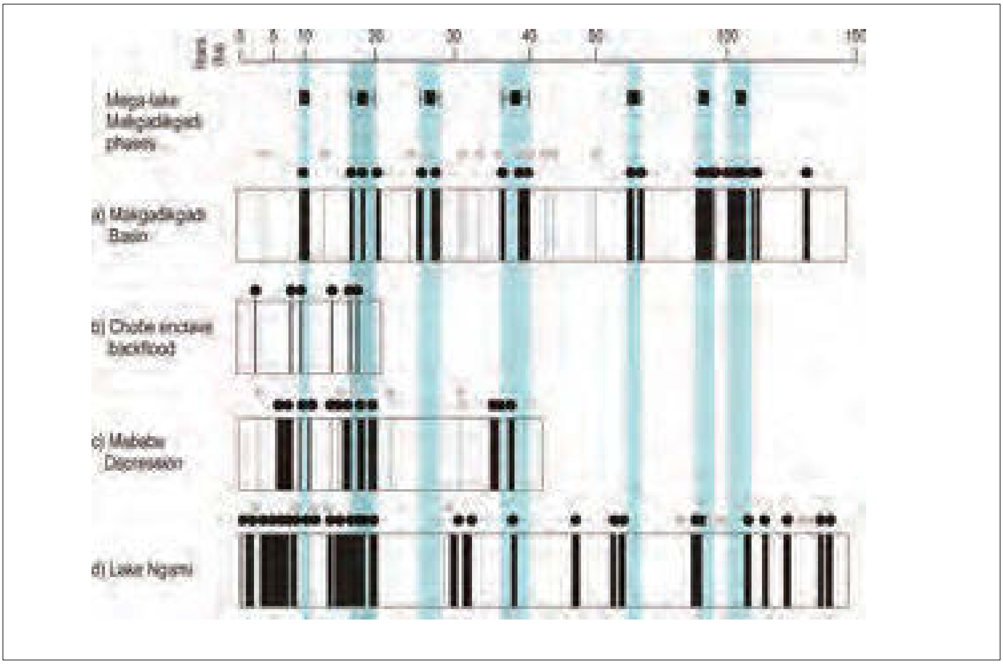

Figure 7.5. The current record of lake stages with the Megalake Makgadikgadi system. After Burrough *et al.* (2009b), using data from Burrough *et al*. (2007), Burrough and Thomas (2008) and Burrough *et al* (2009a). For each sub-basin, including the Chobe enclave at the modern confluence zone of the Zambezi and Chobe Rivers, an individual OSL-dated record of shoreline construction exists. These are shown as a), b), c) and d), with individual ages and their one sigma errors shown as well as inferred phases of high stand. Hatched bars are periods where directly-dated evidence of lake absence is recorded from other records. When all four components are integrated and probable overall lake phases are statistically determined from the record, the upper overall record is produced, and is shown by the pale bars that run the whole vertical length of the diagram.

4.2 Etosha

Etosha, in northern Namibia is another large basin containing a series of pans, the largest of which, Etosha, only experiences surface inundation on a seasonal basis. Smaller than the two main pans of the Makgadikgadi system, Etosha Pan nonetheless covers approximately 6 000 km^2 (Thomas and Shaw, 1991) with a maximum edge-to-edge dimension (east to west) of approximately 120 km (Brook *et al.,* 2010). Being to the west of Makgadikgadi and thus further from the centre of the Kalahari Basin as a whole, the floor of Etosha has a higher overall elevation than Makgadikgadi, averaging 1 080 m a.s.l. (Lindeque and Archibald, 1991). Ephemeral streams (*oshanas* or *omurambas*) enter the pan from the north, northeast and northwest today (Figure 7.6) and, while there are no inflows from the south under current conditions, the presence of spring mounds on the southern margins of the basin suggest that this may not always have been the case (Brook *et al.,* 2010).

Structurally, Etosha occupies part of the Owambo Sub-basin of the Kalahari, which at over 400 m has one of the greatest thicknesses of mid-Neogene to Quaternary age (Kempf, 2000) Kalahari Group sediments in the whole interior of southern Africa (Thomas, 1988; Haddon, 2000). A structural element has been proposed for the initial development of the Etosha Basin as it sits within the Owambo Sub-basin of the Kalahari, but it is in terms of climatic change that the geomorphic development of Etosha has more commonly been interpreted. In this regard, the investigations of Hipondoka (2005) and Brook *et al.* (2007; 2010) have made significant contributions to interpretations that had otherwise changed little since the brief overview provided by Thomas and Shaw (1991).

There have been two broad and contrasting theories put forward to explain the existence of the large pan in the Etosha Basin. First is that the pan is a relic of a former palaeolake that occurred in the Miocene/Pliocene. Secondly, the pan owes its existence to aeolian deflation of soft Kalahari sediments during the late Quaternary. Although these views are sometimes presented as opposing theories, as is often the case, the existing evidence, when scrutinised with care and with the benefit of modern geochronometric applications, suggest a role of both water and wind agencies in the development of Etosha.

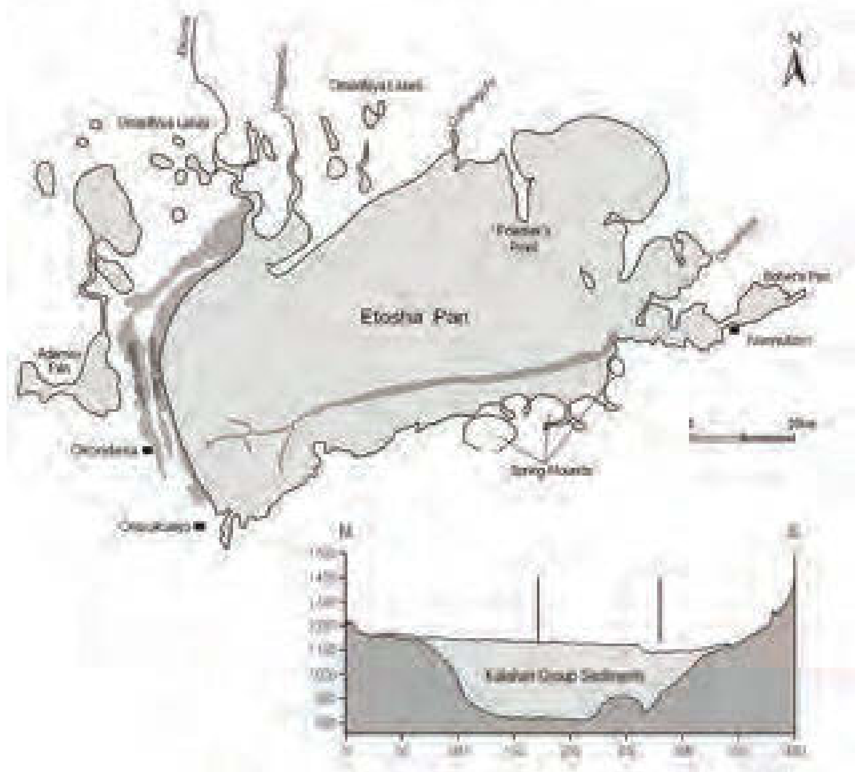

Figure 7.6. The Etosha Basin, showing modern inflows and other geomorphic features referred to in the text. Inset: cross section of the Owambo Sedimentary Basin form north to south, showing relative location of Etosha in a structurally-controlled sedimentary basin. Data from various sources including Hipondonka (2005) and Brook *et al.* (2007).

Wellington (1938; 1939) proposed, on the basis of elevation differences in fluvial systems in the region, that the present-day dry Etosha was the result of a lake in the basin ceasing to receive major inflows. This was ascribed to the seaward capture of a formerly southerly flowing Cunene River, which would have drained into the basin. Examination of basin sediments, and the proposal of a Miocene age for lacutrine stromatolites in upper basin clays, led Miller (2008) to lend recent support for the relic-palaeolake model. However, Hipondoka (2005) notes that despite the Neogene age of Cunene capture, which is relatively undisputed, some proponents of this model suggested that a lake persisted in the basin until about 35 ka.

The alternative model, which gained favour in the 1980s, suggested that aeolian deflation during the Quaternary, under predominantly seasonal semi-arid climatic conditions, was the major driver of basin development (Rust, 1984; Buch and Zoller, 1992). Evidence for this comes from lunette dunes on the westerly, downwind, margin of the basin (Figure 7.6), as well as Etosha being a major source today of atmospheric dust, demonstrating its propensity for deflation (see previous chapter).

Resolving the relative contributions of lacustrine and aeolian processes to basin formation requires careful mapping of landforms in the basin and strong chronometric control for the timing of key landform development and sediment deposition. This has come, as in the case of Makgadikgadi, through imagery and field survey (Hipondoka, 2005); the application of TL dating to lunette sediments (Buch and Zoller, 1992; Buch *et al.,* 1992), OSL dating of shoreline features associated with the lunettes on the western pan margin (Hipondoka *et al.,* 2006; Brook *et al.,* 2007); and careful AMS radiocarbon dating of both stromatolite carbonates and organic components (Brook *et al.,* 2010) to resolve previously ambiguous ages (e.g. Rust, 1984; Miller, 2008).

In summary, these recent studies indicate, not unsurprisingly given other regional records of environmental change in interior southern Africa, that the basin has experienced both dry deflational phases and the presence of significant water bodies in the late Quaternary, including the Holocene. Lunette sediments have accumulated over at least 140 ka (Buch and Zoller, 1992) attesting to deflation from the basin. Importantly, Brook *et al*. (2007) point out that not all Etosha lunette sediments are pure sand, with the presence of clay pellets suggesting that at least some deflational phases have occurred from sediments previously laid down under wet (lake) inundation. The shoreline components of the lunette complex, as dated by Brook *et al*. (2007), attest to several periods of basin inundation in the Holocene (ca. 7-5, 4.5-3.5, 2.5-1.7 and 1.0 ka, excluding age errors). The recent detailed radiocarbon AMS dating of the organic components of stromatolites in the east of the basin shows these to be much younger than surmised by Miller (2008). Falling within 3-19 ka, these ages provide evidence of hypersaline, but persistent lake full conditions with a higher rainfall than the approximate 500 mm p.a. of the present day required to sustain a water body in the basin, on several occasions in the Holocene. As is the case with Makgadikgadi, the detailed chronologies of environmental and climatic change are far from fully resolved, but there is clear evidence that these basins, as well as having prolonged dry phases, also contained persistent lake bodies on more than one occasion during the late Quaternary, including during relatively recent times in the Holocene.

4.3 The large basins today: sediments and landforms of the sub-basins

Contemporary basins reflect a balance of water, salt and sediment inputs and outputs between the surface, groundwater and intermediate (sediment, porewater) systems (see Torgersen *et al*., 1986 for discussion). In the case of Ngami and Mababe, the basins are floored with lacustrine sediments fed by ephemeral/seasonal inflows. The water table forms an inverted dome beneath the basins, the age of the water increasing rapidly with depth and distance from the basin floor itself. This type of lake can be classified as a *recharge playa* (Rosen, 1994), in which the accumulation of salts is not possible, beyond minor efflorescence when surface water evaporates. The most notable feature of the Ngami sedimentary sequence are the extensive diatom beds and burnt *Phragmites* deposits representing the last high stands of the palaeolake (Shaw *et al.,* 2003; Huntsman-Mapila *et al.,* 2006), overlying tens of metres of sand. At the periphery of both basins are extensive calcrete formations, the radiocarbon dating of which has suggested precipitation either from contemporary groundwater, or from high stands in the past (Shaw, 1985b).

By contrast the contemporary Makgadikgadi Pans form the lowest point of the largest groundwater catchment in the sub-continent, and can be defined as a *discharge playa* (Rosen, 1994) or *wet playa* (Reynolds *et al*., 2007), in which groundwater lies close to, but does not intersect, the pan surface. The pan bed is therefore composed of tens of metres of saline clays and sands, with a thin surface efflorescence derived from near surface capilliary action. Hydrochemical studies (Eckardt *et al.,* 2008) suggest that in its present form the basin is balanced between the continual input of hypersaline $Na-CO_3-SO_4-Cl$ type brines from the catchment and the episodic seasonal input of surface floodwaters from the Nata and other eastern rivers, with major flooding events (covering the whole of Sua Pan and much of Ntwetwe Pan) occurring on a *circa* decadal scale. Surface water evaporates to precipitate calcite and halite, but does not contribute to subsurface brines.

Surrounding the basin is an extensive area of calcrete (White and Eckardt, 2006) which has been interpreted as representing calcite precipitation from evaporation following high lake stands (Ringrose *et al*., 1999; 2005). The presence of silcretes and sil-calcretes within the peripheral calcrete matrix would suggest Si precipitation at high pH evaporative sites as the lake desiccated (Ringrose *et al.,* 2005), a process which occurs very rapidly on the pan surface at the present (Shaw *et al.,* 1990). The wetting and drying of the modern surfaces of Makgadikgadi and Etosha, and the ephemeral input of sediments from surface flows, mean that the Makgadikgadi Pans are also notable for being a major atmospheric dust source, as discussed in the previous chapter.

5. Pans

Pans are essentially enclosed basins ranging in size from a few square metres to larger features in the 1-16 km^2 range (Lancaster, 1978) and upwards to the component units of major basins such as Etosha and the Makgadikgadi. They are widespread throughout southern Africa, ranging from the Agulhas Plain (Figure 7.7) (Carr *et al.,* 2006) in the south to north of Mongu in western Zambia (Williams 1986), where they are known as plains. Latitudinally they extend from the Namibian coast (Eckhardt *et al*., 2001) to central Mozambique (Tinley, 1977).

Figure 7.7. Voelvlei, a pan on the Agulhas Peninsula, in November 2002. Looking north towards the Bredasdorp Mountains. On the right is the inner of two marginal lunette dunes, with wave action in this wet pan truncating the lunette sediments giving rise to a small cliff (ca. 2 m high). The pan and its evolution are discussed in Carr et al. (2006).

Attempts to explain the distribution and density of pans have relied primarily on climatic and lithological factors (Goudie and Wells, 1995). Thus most occur in the west and centre of the subcontinent, with less than 500 mm mean annual precipitation, or more than 1 000 mm p.a. free surface evaporation (Goudie and Thomas, 1985). Pan densities are highest in the western Free State; reaching one per km^2 around Bultfontein, while in the Kalahari the densities of the somewhat larger pans are greatest on the *Bakalahari Schwelle* (Passarge, 1904), the imperceptible watershed between the Molopo River and the Central Kalahari drainage of the Okwa and tributaries. Other major concentrations occur in the Mier Country between Upington and Keetmanshoop, the Agulhas Coast and, outside of the climatic limits, in the Mpumalanga Lake District (>750 mm p.a. rainfall), in western Zambia (>1 000 mm p.a. rainfall) and in Hwange National Park, where Goudie and Thomas report 2 449 pans at a density of 0.15 pans per km^2.

With a few exceptions, the pans are developed on two geological formations, the Karoo Supergroup or the Kalahari Group. With the former, tillites, shales, sandstones and mudstones of the Ecca Group and Dwyka Formation are particularly susceptible, along with the Ventersdorp lavas. All of these strata have low resistance to weathering and a tendency to yield salts; where salt content is low, as in the Beaufort Group sandstones (De Bruiyn, 1971), pan development is impeded. The extensive Kalahari Group sediments of sands and duricrusts underlie the pans of the Northern Cape Province of South Africa, Botswana, Zambia and Zimbabwe. Although the sands themselves are of low weatherability, the availability of solutes has led to extensive duricrust development – primarily calcretes and silcretes in the Kalahari, and ferricretes further north in Botswana and Zambia. Three dimensional studies of pans are uncommon; geophysical and geochemical investigations of two pans in the southern Kalahari (Butterworth, 1982; Farr *et al.,* 1982) revealed chemical alteration and calcrete formation to a depth of 30 m, with weathering of the underlying Karoo bedrock and dolerite intrusions. In other studies (e.g. Bruno, 1985) sub-pan and peri-pan calcrete layers, being relatively impervious aquicludes, have been identified as supporting perched water tables in the pan itself.

5.1 Pan morphology and origin

In plan, most pans are subcircular, or of kidney or clam shape. Pans may be elongated where topographic controls, such as linear dunes, are present (Thomas, 1984), and may form depressions as much as 20 or 30 metres below the surrounding terrain. Observation of the preferred orientation (e.g. Le Roux 1978 in the Free State) suggests that the long axes of pans develop transverse to the dominant wind direction, a factor previously noted in western Australia (Killigrew and Wilkes, 1974). Lancaster (1978a) also identified a secondary trend parallel to the dominant wind direction, although this may be a function of adjacent dune orientation. These observations suggest that dominant winds have an important role in shaping pans.

Pans have an approximately flat basin floor within the depression. Boocock and Van Straten (1962) differentiated between grassed, ungrassed (clay) and saline pans, though these terms must be taken as indicative as over time surface conditions in most pans vary according to antecedent hydrological conditions (Figure 7.8). In the Mpumalanga Lake District differences between open, grassed and reed covered pans can be explained by minor variations in near surface chemistry, particularly variance in sulphates (Russell, 2008). High salt accumulation would imply near-surface groundwater activity – however, only a few pans in southern Africa are of this type (see below), most are essentially clay floored, with water tables at depth.

Figure 7.8. Koopan Sud, Northern Cape Province of South Africa, looking towards a major lunette dune complex (over 20 m high) on the eastern pan margin. In April 2004 after a wet summer that lead to significant surface flooding. In September 2010 in the late dry season.

The origin of pans has raised vigorous debate over the last century. Early studies (e.g. Allison, 1899; Passarge, 1904) suggested their evolution could be attributed to the erosive activity of game animals congregating for water and sodium salts. Although this process undoubtedly occurs, it will be spatially limited and does not account for the size or distribution of the features, though it will account for the presence of waterholes in the pan bed. In the 1970s aeolian deflation was proposed as a major process (Lancaster, 1978a; 1978b), accounting for both pan orientation and the presence of lunette dunes. As noted above, wind activity clearly shapes pans, but does not account for their features at depth.

Goudie and Thomas (1985) provided a diagram that illustrates the complex array of factors that can influence the geomorphic development of pans and which predispose the landscape to pan development, namely a lack of integrated surface drainage, the presence of susceptible weatherable lithologies and a semi-arid climate (Figure 7.9). Marshall and Harmse (1992) concur with these pre-disposing factors, adding the presence of geological structures, such as faults, sills and dykes as conduits for groundwater flow. This implies that weathering a depth is a major contributory factor in pan development, with many pan systems forming along buried drainage lines, as proposed by Wellington (1943) for the Mpumalanga Lake District. Subsequent studies using DTM and geochemical data suggest that this hypothesis is well founded (Russell, 2008).

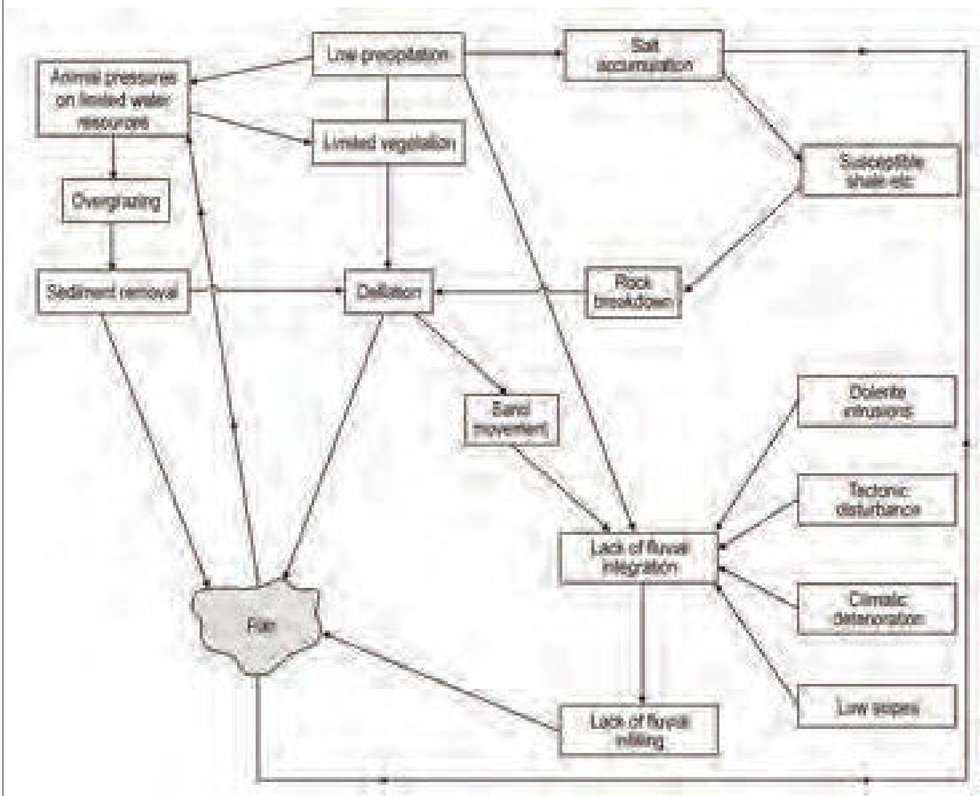

Figure 7.9. A model of pan development (after Goudie and Thomas, 1985).

A model of clay floor pan development was been proposed by Wood and Osterkamp (1987) for the Texas High Plains (Figure 7.10). Although developed in a context where pans occur over limestone and sandy-limestone lithologies, this is not inappropriate to southern African contexts where sandy lithologies are found in association with calcretes. In Texas, initial depressions have been created by drainage ponding, structural barriers and deflation, leading to the concentration of seasonal runoff and groundwater recharge. In turn recharge leads to sub-surface weathering, in particular oxidation and carbonate dissolution, resulting in piping, bulk loss and basin enlargement. As the pan develops insoluble clays accumulate in the lowest, usually central, areas, increasing groundwater seepage and leading to peripheral weathering. Where percolation is limited (i.e. the water table is at depth) weathering can be reversed to produce the accumulation of duricrusts. In turn clay accumulation occurs through inflow and loss by deflation. Such a model explains many of the features encountered in three dimensions in southern African pans, and provides a genetic link to associated landforms, such as the dry valleys (*mekgacha)* of the Kalahari (Shaw and De Vries, 1988, Nash *et al.,* 1994).

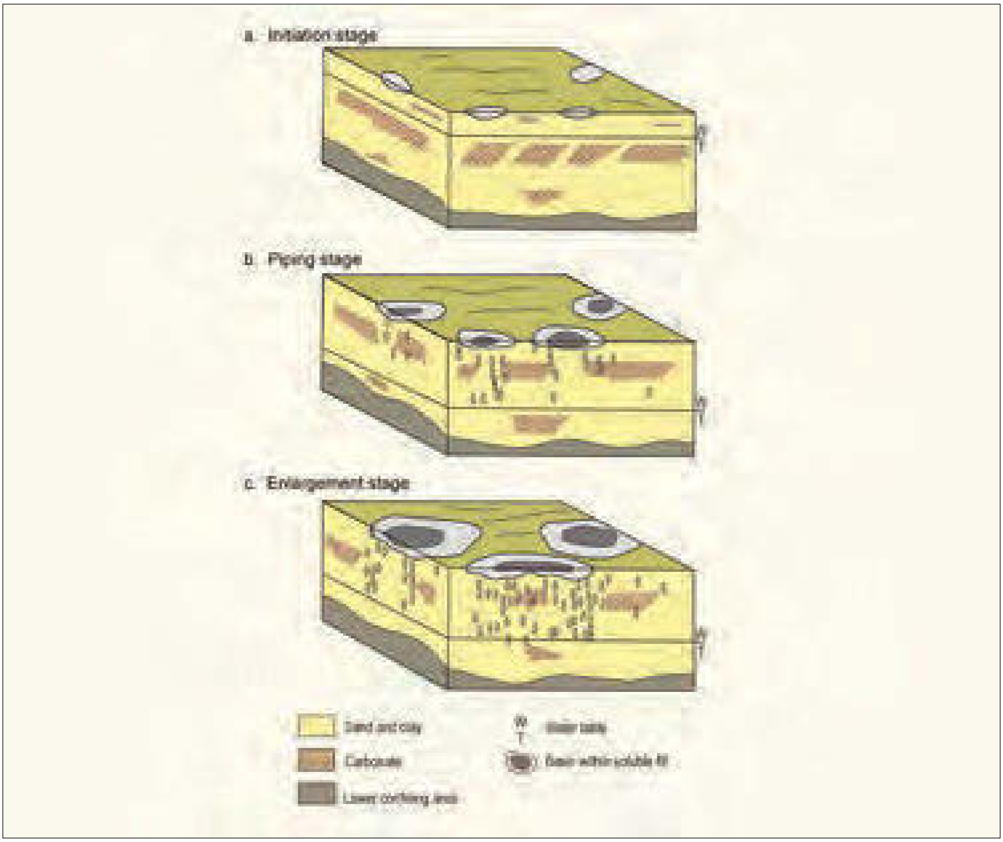

Figure 7.10. A model of pan evolution based on the High Plains of Texas and New Mexico (after Wood and Osterkamp, 1987). a) Initiation stage. b) Development of depressions by sub-surface piping due to concentration of seepage to groundwater. c) Pan enlargement by concentration of seepage from basin perimeters.

5.2 Pan chemistry

As noted earlier in the chapter, there has been a shift from the largely descriptive and classificatory studies of the 1970s to 1990s, to a more detailed process outlook. This is evident from the geochemical studies of the Makgadikgadi Basin mentioned previously. It has not, however, been matched in the studies of small pans. Most of these fall within Rosen's (1994) classification of *recharge playas* in which the groundwater table, fed in most cases by fairly small catchments, lies at depth, and there is limited interface with surface waters, which function in the interstitial zone by percolation and capillary action. The end product is usually a suite of alkaline clays, with some surface efflorescence (Figure 7.1) when desiccation occurs. This generalised picture is modified by the import and export of water, solutes and sediments over time.

A recent study of the hydrochemistry of the Mpumalanga lakes (Russell, 2008), which lie at the wet end of the climate spectrum with Lake Chrissie a semi-permanent water body (Figure 7.2), show enormous inter-pan variation in chemistry, dependent on the stage of evaporative process. Although all of the pans lie on the Na-Cl-HCO_3 pathway, total TDS in pan water varied from 100 mg/l in fresh water, to 10 g/l for saline pans and 90 g/l for evaporative pools, whilst groundwater from boreholes peaked at 1 g/l (Figure 7.11).

In hydrologically-ephemeral pans, salinity was greatest at the end of the desiccation cycle, but reduced as the dry season progressed due to deflation of solutes. A major factor in limiting the evaporative concentration potential was found to be the ratio of catchment area to pan surface area, with the most saline pans having the lowest CA:PSA ratios. Given the greater potential for evaporation and deflation in the drier parts of the region, it is probable that these characteristics will be even more pronounced.

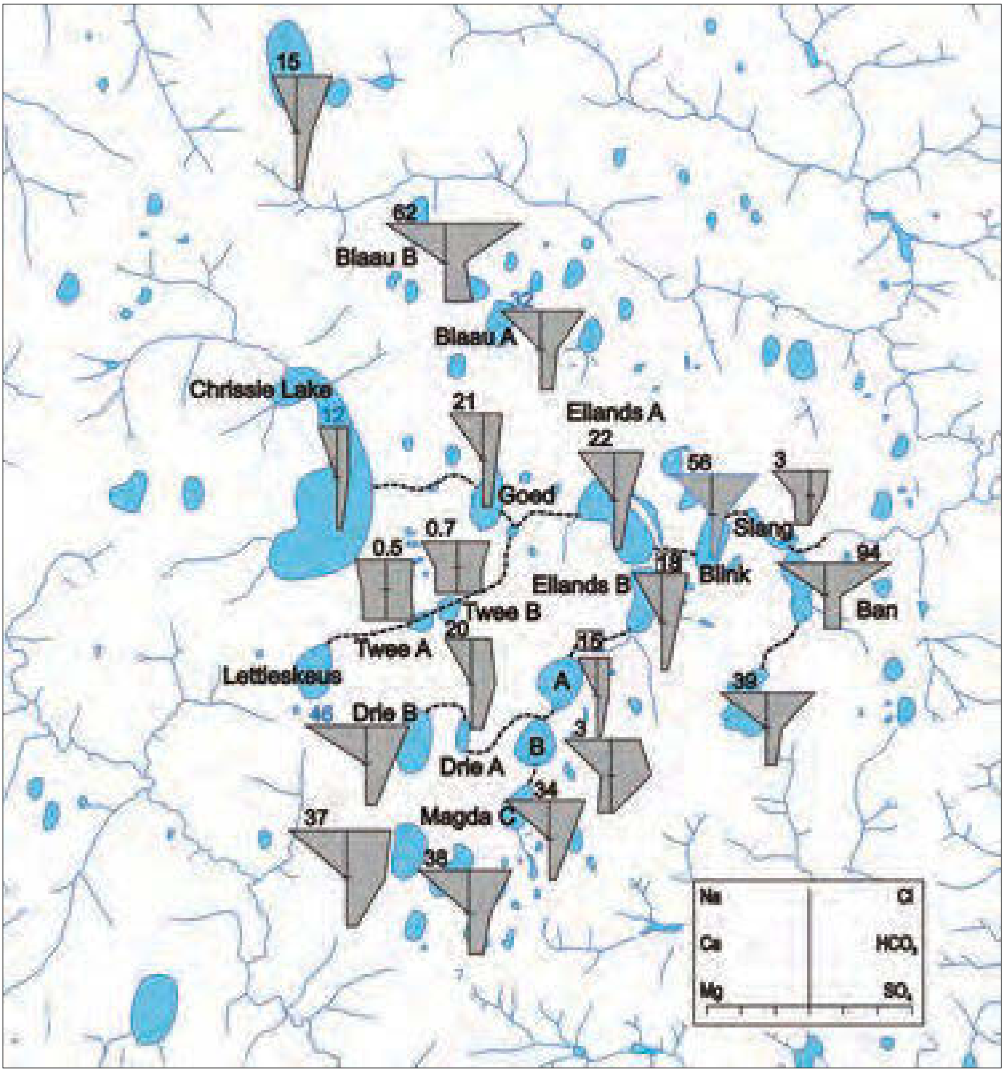

Figure 7.11. Ion concentration diagrams (Stiff Diagrams) for wet season water samples taken from the Mpumalanga Lakes. Stiff diagrams display ion concentrations (expressed in meq/l) of cations (left) and anions (right) around a vertical axis representing zero (after Russell, 2008). Numbers indicate the Na concentrations in meq/l for scale. Drainage linkages suggested by Wellington (1943) shown by stippled lines. Common water chemistry in some cases (e.g. Tweeling A and B) suggests groundwater linkages.

5.3 Pans as palaeoenvironmental tools

Following the investigations of Lancaster (1979) and the subsequent application of OSL dating to pan-margin lunettes, southern African pans have played an important role in Late Quaternary palaeoenvironmental research, especially through the dating of periods of lunette development, from pan systems spanning from Agulhas (Carr *et al.,* 2006) to Etosha (Buch *et al.,* 1992), via the Karoo (Thomas *et al.,* 2002), Free State (Holmes *et al.,* 2008) and southern Kalahari (Telfer and Thomas, 2006). The records contained in these studies are considered in the overview of southern African Quaternary environments (see Chapter 12) and are therefore not evaluated in a climatic context here. What is clear, however, is that the palaeoenvironmental interpretation of pan and lunette sediments is best achieved when it can be done in conjunction with other records of environmental change from the same vicinity, and when the full geomorphic considerations leading to landform development can be assessed.

6. Conclusion

Though much of southern Africa is a relatively dry subcontinent, and is certainly one where potential evapotranspiration exceeds precipitation on an annual basis today, the role of water in shaping the landscape cannot be overestimated and is evidenced no better by the presence of lake basins and pan depressions. In part this represents a legacy from wetter times past in the Quaternary period, though such an interpretation is proving to be unduly simplistic and more and better data on landscape evolution accrue (Thomas and Burrough, 2011). In conclusion, therefore, it is possible to draw out the following points from the ever-growing body of data on basin and pan development:

- The role of tectonics in setting up the circumstances for large basin development has to be acknowledged, but does not hold as the primary explanation of the wetting and drying of the major dryland basins in the late Quaternary.
- For the major lake basins and the smaller pans, models of alternate wet and dry conditions do not alone account for the geomorphic features present or the development of the basins as they exist today. Seasonal wetting, and interacting lacustrine and aeolian processes, have to be given credit for the existence of some features, especially on basin margins, and some sediments found within basins themselves.
- That point notwithstanding, there is significant, though not yet fully temporally-resolved, evidence that the major basins, and the pans in at least some subareas, have been host to persistent bodies of water (persistent meaning of sufficient duration to impact on morphological and sedimentary features).
- Given their size, and the size and latitudinal spread of their catchments, the Makgadikgadi and Etosha Basins are complex features that require local rainfall changes, changes in more tropical catchments, and hydrological feedbacks from the existence of large water bodies, to fully explain the waxing and waning of lakes on multiple occasions during the late Quaternary.
- Smaller pan depressions may also have complex development histories, with the factors (or balance between factors) leading to pan development potentially differing over the large latitudinal range that they occur in.
- The palaeoenvironmental archives present in individual basins should be interpreted on a case-by-case basis, as hydrological histories and preservation potentials for sediments can vary over relatively short distances. This should not be surprising given that in quite discrete areas (such as in western Zambia, the Free State or the Northern Cape) adjacent pans may be wet or dry at any one time, may have or not have fringing lunette dunes, and may have different sediment sequences preserved on basin floors.

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8

Duricrusts

Duricrusts

David J Nash

1. Introduction

Duricrusts are, by definition,

> *... near-surface geochemical crusts formed as a result of low-temperature physico-chemical processes operating within the zone of weathering which lead to the absolute or relative accumulation of minerals through the replacement and/or cementation of pre-existing soil, sediment, bedrock or weathered material (Nash, 2009:284-285).*

They have long been recognised as an important control upon landscape development in southern Africa. Duricrusts exist in a variety of forms, with gypsum-cemented *gypcrete*, calcite-cemented *calcrete*, silica-cemented *silcrete*, and iron oxide-rich *laterite* and iron oxide-cemented *ferricrete* the most widespread in the subcontinent.

Investigations in southern Africa have made major contributions to global knowledge about the origins of duricrusts. Indeed, some of the earliest accounts of duricrusts in the scientific literature were made in the subcontinent. The British missionaries David Livingstone (1857) and Robert Moffat (1858), for example, described *calcareous tufa* and *ferruginous conglomerate* during their travels in present-day South Africa and Botswana, whilst the French geographer Émile-Félix Gautier (1902) includes accounts of laterite profiles in his overview of the landscape of Madagascar. Siegfried Passarge (1904) was one of the first scientists to describe the relationship between calcrete and silcrete in his seminal work on the geology and geomorphology of the Kalahari. Arthur Rogers and Alex du Toit (1909) include a detailed description of variations in the physical properties of silcrete within their account of the geology of the former Cape Province of South Africa. Modern scientific investigations into the origin of silcrete in southern Africa have a legacy extending back to the 1930s, with key investigators including Bosazza (1936; 1939), Frankel and Kent (1938), Mountain (1946; 1951; 1980), Frankel (1952), Du Toit (1954), Goudie (1973), Smale (1973) and Summerfield (1981; 1982; 1983a-d; 1984). Research into gypsum crusts dates back to the work of Gevers and Van der Westhuyzen (1931), and Kaiser and Neumeier (1932), whilst calcrete has been described in detailed studies since the 1950s by Du Toit (1954), Mountain (1967), Netterberg (1967; 1969a-c; 1971; 1980) and Watts (1980). Ferricrete and laterite also have a long history of investigation in the subcontinent (e.g. King, 1951; Du Toit, 1954; Maud, 1965; Mountain, 1967; Helgren and Butzer, 1977).

The term *duricrust*, first coined in Australia by Woolnough (1927), suggests a material that is extremely hard. In fact, crusts can exhibit a range of consistencies including highly indurated (e.g. hardpan calcrete, nodular ferricrete), non-indurated (e.g. soft or powdery forms of calcrete) or a mixture of the two (e.g. nodular calcrete). The degree of induration is an important regulator of the landscape effects of any crust. For instance, during the earliest stages of its development, weakly cemented calcified soil such as found in parts of the Kalahari Desert (Netterberg, 1980; Watts, 1980), is relatively easily eroded and has little long-lasting geomorphological impact. In contrast, more indurated crusts, including silcrete, hardpan calcrete and the upper parts of laterite weathering profiles, can be persistent features and protect underlying rocks and sediments from subaerial weathering and erosion, thereby enabling the preservation of thick sedimentary sequences (King, 1951; Du Toit, 1954; Nash and McLaren, 2007a). Extensive silcrete and lateritic crusts (Figure 8.1a), for example, preserve palaeosurface remnants along the Cape Coastal Belt (Summerfield, 1983a; Marker and McFarlane, 1997; Marker *et al.*, 2002), and calcrete tens of metres in thickness underlies large areas of the southwest Kalahari (Figure 8.1b; Nash *et al.*, 1994a; Haddon and McCarthy, 2005).

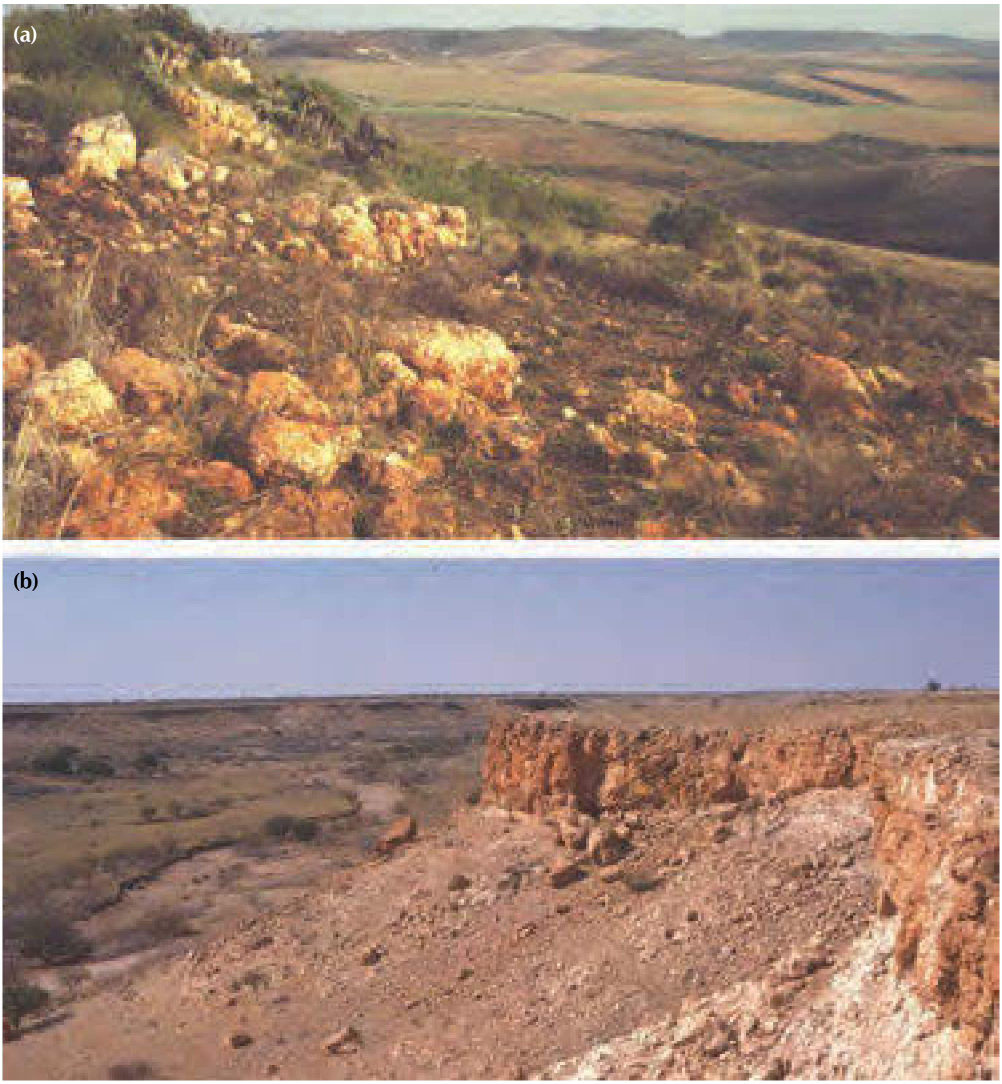

Figure 8.1. Duricrusts as controls upon landscape evolution in southern Africa: a) A partially dissected silcrete carapace preserving remnants of the African Surface Complex near Albertinia in the Western Cape Province, South Africa; b) A 10-m thick partially silicified calcrete duricrust exposed in the flanks of the Auob River in the southwest Kalahari near Kalkheuval Farm, southeast Namibia.

The overall aim of this chapter is to provide an understanding of the geomorphological significance of calcrete, silcrete, gypcrete and iron-rich duricrusts in southern Africa, and to demonstrate how recent developments have improved our knowledge of the long-term development of landscapes in the subcontinent. For each duricrust type, the general properties, classification, geomorphological context, distribution in southern Africa and latest ideas concerning formation are considered. The chapter concludes with a short review of the role of duricrusts as palaeoenvironmental indicators. A map showing the general distribution of the major types of duricrust in southern Africa is given in Figure 8.2. This figure also identifies areas underlain by a duripan (or *dorbank*), a diagnostic soil horizon that is cemented by illuvial silica into a subsurface hardpan, which is not discussed here (see Ellis and Schloms, 1982, for further information). For more detailed maps of calcrete, silcrete and laterite/ferricrete distribution, readers are referred to Summerfield (1982; 1983b), Netterberg (1985), Botha (2000), Partridge *et al*. (2006) and Fey (2010).

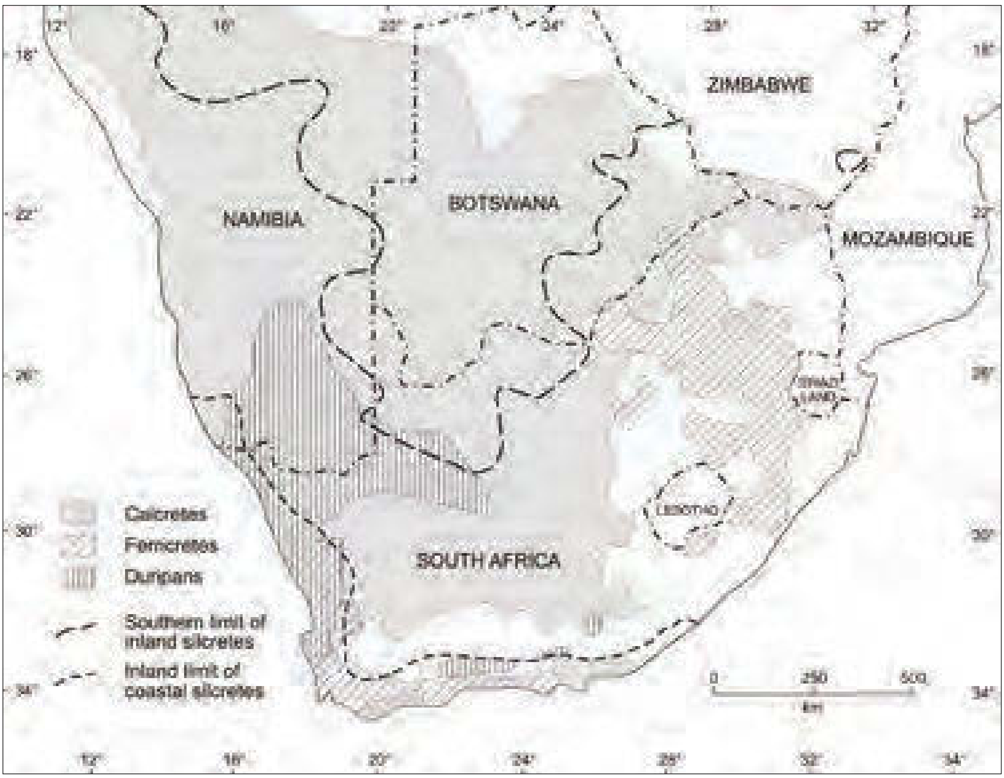

Figure 8.2. General distribution of pedogenic duricrusts in southern Africa (after Du Toit, 1954; Weinert, 1980; Ellis and Schloms, 1984; Schloms and Ellis, 1984; Netterberg, 1985; Partridge, 1997; Botha, 2000; Fey, 2010).

Throughout the chapter, the various categories of duricrust are discussed separately. However, it should be noted that silcrete, calcrete, gypcrete and laterite/ferricrete can also exist in hybrid or *intergrade* forms where, for example, diagenetic alteration has occurred due to water percolation or minerals have been precipitated within pores during the later stages of cementation. It is possible in parts of the Kalahari, for example, to find not only silcrete and calcrete, but also calcareous silcrete (or *cal-silcrete*) and silicified calcrete (or *sil-calcrete*) (e.g. Summerfield, 1982; Shaw and De Vries, 1988; Nash and Shaw, 1998). Indeed, as early as 1907, Lamplugh coined the term *silici-calcretes* to described silicified calcretes along the Zambezi River (Lamplugh, 1907). Calcretised ferricrete (*cal-ferricrete*), iron-cemented calcrete (*ferro-calcrete*) and silcrete (*ferro-silcrete*) have been identified in some environments (Nash *et al.,* 1994a; Lee and Gilkes, 2005; Ramakrishnan and Tiwari, 2006), and gypcrete and calcrete may also interdigitate (Watson, 1985; Jacobson *et al.*, 1988) to produce a complex spectrum of duricrust forms.

2. Gypsum crusts

Gypsum crusts (often referred to as *gypcretes*) are described by Watson (1985) as:

> *Accumulations at or within 10 m of the land surface from 0.10 m to 5.0 m thick containing more than 15% by weight gypsum ... and at least 5.0% by weight more gypsum than the underlying bedrock (Watson, 1985:885).*

They are the most soluble of all the duricrusts and have a restricted geographical distribution, partly due to their solubility, but also as a result of the less widespread occurrence of sources of gypsum ($CaSO_4{\cdot}2H_2O$) compared to calcium carbonate or silica. Most gypsum crusts are found in areas where mean annual rainfall is less than 250 mm (Watson, 1985), with a transition from calcretes to gypsum crusts identified in Tunisia as rainfall drops below this level (Pervinquière, 1903). There also appears to be a rainfall threshold below which gypsum crusts are less common. In the Namib Desert, the only region of southern Africa where extensive gypsum crusts have been documented, there is a gradual transition from gypsum to evaporites as rainfall drops below approximately 25 mm (Watson, 1983a, b; 1985).

Gypsum crusts often form an erosion-resistant horizon at the land surface. However, due to their susceptibility to dissolution, they rarely create the mesa-and-butte landscapes associated with duricrusts such as silcrete and calcrete. They can develop as a result of pedogenic processes, as lacustrine evaporites (Warren, 1982) or as phreatic precipitates (Kulke, 1974; Risacher, 1978). Pedogenic gypsum crusts may develop directly on unweathered bedrock (Watson, 1985, 1988), or on unconsolidated sediments such as colluvium, alluvium and dune sand (Watson, 1979). They may blanket the landscape and can form on steep slopes beyond the phreatic zone. Non-pedogenic gypcretes, in contrast, normally develop close to landscape depressions.

Three main forms of crust have been identified (Watson, 1979; 1983a; 1985). The first are horizontally bedded crusts. These generally occur at the surface and are composed of 0.05-0.1 m thick layers, each showing a gradation from top to bottom of fine-grained alabastrine material (crystal diameters $<50\ \mu$m) to coarse (>0.5 mm) crystals. The second form is a subsurface crust, which can be subdivided into two groups. These are

i. a macrocrystalline desert rose form (Figure 8.3), composed either of large lenticular crystals (>1 mm diameter) or meso-crystalline material (crystal diameters from 50 μm to 1.0 mm), which can achieve thicknesses of 5 m; and

ii. a mesocrystalline form characterised by lenticular crystals (<1 mm diameter) that is usually less than 2 m thick and often exhibits a columnar macrostructure.

The third form of gypsum crust is a non-bedded surface crust (Figure 8.3), which exists in three varieties. These are

i. an indurated microcrystalline form (crystal diameters $<50\mu$m) up to 2 m thick, which commonly has a columnar structure;

ii. an unconsolidated, powdery form up to 2 m thick; and

iii. an intermediate form comprising a powdery matrix with microcrystalline gypsum cobbles (<0.5 m in diameter).

Most surface crusts appear to be exhumed mesocrystalline crusts (Watson, 1985; 1988), the cobble form representing a stage in the degradation by dissolution and leaching of columnar crusts to a powdery residue. Desert rose crusts range in colour from white or grey to green or red, depending on the host material, whilst columnar surface crusts and the majority of other gypsum crusts are usually white or grey in colour.

Figure 8.3. Examples of gypsum crusts: a) Powdery surface pedogenic gypsum crust from the Tumas River in the Namib Desert, Namibia; b) Desert rose crystals sampled from sabkha sediments between Terrace Bay and Agate Beach along the Skeleton Coast, Namibia.

The three forms of gypsum crust are chemically and mineralogically diverse. Horizontally bedded crusts generally contain 50-80% gypsum, desert rose crusts contain 50-70% gypsum, and the mesocrystalline subsurface form and surface forms up to 90% gypsum. The other main constituents are quartz grains, calcium carbonate and iron minerals inherited from the host sediment (Watson, 1983a; 1985). Clay minerals including kaolinite, smectite, mica and palygorskite (Reheis, 1987), and traces of elements such as sodium and strontium (Watson, 1983a), may also be present. Fey (2010) contains a recent review of the mineralogy of gypsum-rich soils in South Africa.

Gypsum crusts are estimated to cover more than 30 000 km^2 of the Namib Desert (Eckardt *et al.*, 2001), one of the most extensive gypcrete deposits known. They have been described in a number of studies including Gevers and Van der Westhuyzen (1931), Martin (1963), Scholz (1963; 1972), Besler (1972), Goudie (1972), Rust and Wieneke (1973; 1976), Wieneke and Rust (1973a,b; 1975; 1976), Cagle (1975), Carlisle *et al.* (1978), Rust (1979), Watson (1979; 1981; 1983a; 1985; 1988), Wilkinson (1990), Heine and Walter (1996), Eckardt and Spiro (1999), Bao *et al.* (2001), and Eckardt *et al.* (2001). Much of the deposit in the Central Namib is obscured by stone pavement, but consists of mesocrystalline, lenticular and prismatic gypsum crystals, together with fibrous gypsum and lenses of alabastrine gypsum, that cement the upper horizons of the pavement colluvium (Watson, 1988). This crust may be up to 5 m thick (Eckardt *et al.*, 2001). A number of playas within the Central Namib and sabkhas along the Skeleton Coast contain extensive gypsum and halite deposits (Gevers and Van der Westhuyzen, 1931; Kaiser and Neumeier, 1932; Torien, 1964; Eckardt *et al.*, 2001). In places, both bedded and sub-surface desert rose type crusts are visible, forming areas of positive relief around the surface depressions within which the crusts developed (Watson, 1985). Former beaches along the Atlantic coast are mantled by surface gypsum crusts (Rust and Wieneke, 1973; 1976; Wieneke and Rust, 1973a, b; 1975; 1976) which have the appearance of exhumed and degraded desert rose crusts (Watson, 1985).

Capillary rise and illuviation mechanisms have been put forward to explain pedogenic gypsum accumulation. The illuvial model, whereby gypsum deposited at the surface is dissolved by rainwater, leached into the regolith and precipitated during subsequent desiccation, is the most widely embraced (Watson, 1979; 1983a; 1985; Chen, 1997; Eckardt *et al.*, 2001). Critics of the capillary rise model argue that the precipitation of salts from gypsum-saturated soil- or ground-water brought to the surface by capillary rise would rapidly plug desert soil profiles and prevent further evaporation. Significantly, capillary rise also requires large quantities of water to generate even a thin crust (Watson, 1979; 1983a; 1985). The evolution of the surface forms of gypsum crust is contingent upon the exhumation of illuvial crusts (Watson, 1985). Pedogenic gypsum accumulation deep within a soil may indicate a past wetter climate (Reheis, 1987).

The main sources of gypsum for pedogenic gypcrete formation are thought to be wind-blown sand, dust or aerosols (Watson, 1979; 1985a; Dan *et al.*, 1982; Amit and Gerson, 1986; Bao *et al.*, 2001). Gypsum may be deflated from playa surfaces (Coque, 1955a,b; 1962; Watson, 1985) or from existing hydromorphic gypsum crusts (Reheis, 1987; Chen *et al.* 1991a,b). Eckardt *et al.* (2001) identify both pathways as potential sources for pedogenic gypcrete accumulation in the central and coastal Namib, with the deflation and/or fluvial reworking of gypsum from inland playas and pedogenic gypsum crusts providing a source for secondary gypsum accumulations down-slope or down-wind (Figure 8.4). Fog, sea spray, biogenic sulphur and marine evaporite deposits have all been suggested as sulphur sources for gypsum crusts (e.g. Martin, 1963; Carlisle *et al.,* 1978; Watson, 1985a; Chivas *et al.*, 1991; Day, 1993; Eckardt and Spiro, 1999; Rech *et al.,* 2003; Drake *et al.,* 2004), although Eckardt and Schemenauer (1998) demonstrate that the ionic content of Namib fog is too low to act as a major source.

Various non-pedogenic models have been put forward to describe the deposition of gypsum close to the groundwater table or in lacustrine environments. Jacobson *et al.* (1988) suggested that the development of gypsum deposits beneath central Australian palaeolakes was driven by interactions between groundwater and infiltrating meteoric waters. In the Namib Desert, Watson (1985; 1988) has attributed the formation of gypcretes associated with playa lakes to two mechanisms. Bedded, lacustrine evaporites probably precipitated in shallow-water environments, whereas phreatic desert rose crusts accreted from evaporating groundwater where the water table was 1-2 m below the land surface.

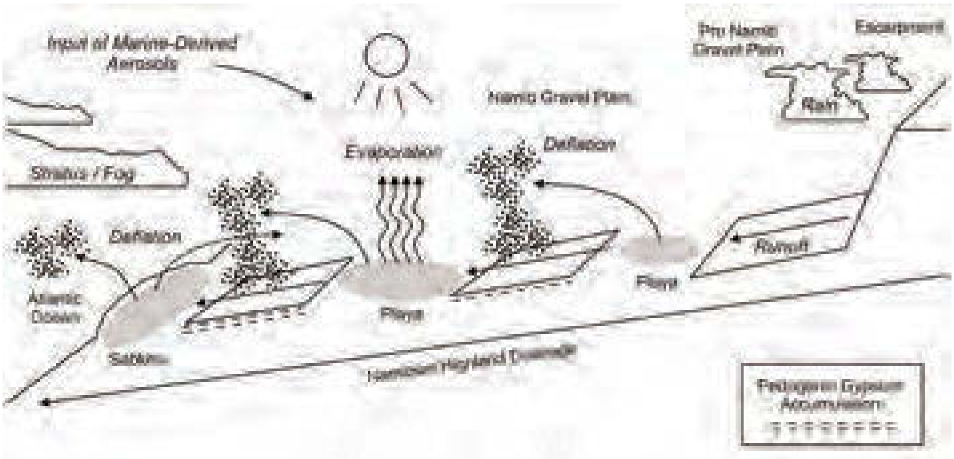

Figure 8.4. Model of pedogenic gypcrete formation for the Namib Desert. Marine-derived aerosols provide a sulphur source for the formation of primary gypsum accumulation in inland playas and sediments. These deposits are subsequently reworked by aeolian and fluvial dispersal to provide a gypsum source for gypcrete formation nearer to the coast (after Eckardt *et al.*, 2001).

3. Calcrete

Calcrete is a term that was first employed by Lamplugh (1902) to describe calcium carbonate cemented gravels in the Republic of Ireland and was applied later to geochemical sediments exposed in the vicinity of the Batoka Gorge where it forms the border between Zambia and Zimbabwe (Lamplugh, 1907). It is now used to describe

> *... a near surface accumulation of predominantly calcium carbonate, which occurs in a variety of forms from powdery to nodular, laminar and massive. It results from the cementation and displacive and replacive introduction of calcium carbonate into soil profiles, sediments and bedrock, in areas where vadose and shallow phreatic groundwaters are saturated with respect to calcium carbonate (Wright, 2007:10).*

Calcretes have been classified in numerous ways, but the most fundamental distinction is between varieties that develop as a horizon (or multiple horizons) within the vadose zone of soil profiles (*pedogenic calcretes*) and those that form around the water table or capillary fringe (often grouped for simplicity as *non-pedogenic calcretes*) (Figure 8.5). The ways in which the different types of calcrete fit within these categories are shown in Table 8.1 (after Carlisle, 1983). Pedogenic calcretes form largely as a result of illuvial-eluvial pedogenetic processes, whereby calcium carbonate is redistributed vertically within a soil or sediment profile. In contrast, non-pedogenic calcretes develop following carbonate precipitation from shallow groundwater. Recent reviews of the characteristics, formation and wider significance of calcretes within dryland landscapes are provided by Alonso-Zarza (2003), Wright (2007) and Dixon and McLaren (2009). Detailed overviews of the characteristics of southern African calcretes are provided by, amongst others, Netterberg (1969c), Watts (1980), and Nash and McLaren (2003).

Figure 8.5. Examples of pedogenic and non-pedogenic calcrete profiles from southern Africa: a) Powdery to nodular pedogenic calcrete exposed near New Hanehai, central Botswana; b) Valley calcrete exposed within a river terrace in the floor of the Okwa Valley, central Botswana.

Table 8.1. A classification of calcrete types (Carlisle, 1983).

CALCRETE CLASSIFICATION	INCORPORATED CALCRETE TYPES
Pedogenic calcrete	Caliche; Kunkar; Nari
Non-pedogenic superficial calcrete	Laminar crusts; Case hardening
Non-pedogenic gravitational zone calcrete	Gravitational zone calcrete
Non-pedogenic groundwater calcrete	Valley calcrete; Channel calcrete; Deltaic calcrete; Alluvial fan calcrete
Non-pedogenic detrital and reconstituted calcrete	Recemented transported calcrete; Brecciated and recemented calcretes

Calcrete is widespread across the arid and semi-arid regions of southern Africa (see descriptions and distribution maps in Netterberg, 1969c; 1971) and is found both within sedimentary successions and at landform surfaces (e.g. Goudie, 1973; Nash *et al.*, 1994a). The greatest areal coverage occurs in the Kalahari Basin, where the material forms an important component of the Kalahari Group sediments (Haddon and McCarthy, 2005). Calcretes are not universally exposed in the Kalahari, however, but mainly outcrop along valleys, around pans and at the margins of palaeolakes. Considerable thicknesses are exposed along the Okwa-Mmone drainage systems in Botswana, and these sequences may merge with similar deposits along the Molopo-Auob-Nossop valleys in northern South Africa and southeastern Namibia (e.g. Boocock and Van Straten, 1962; Watts, 1980; Mallick *et al.*, 1981; Nash *et al.*, 1994a; Nash and McLaren, 2003; Haddon and McCarthy, 2005; Ringrose *et al.*, 2007). Extensive calcrete deposits also outcrop around the Makgadikgadi, Ngami and Etosha Basins and the Mababe Depression (e.g. Grove, 1969; Buch and Rose, 1996; Ringrose *et al.*, 2005; 2009; White and Eckardt, 2006) in association with Quaternary shorelines, and at the margins of smaller ephemeral pans (e.g. Lawrence and Toole, 1984; Goudie and Wells, 1995). In addition to exposures in the Kalahari, calcrete is found in other parts of South Africa including the Karoo, Postmasburg and Mafikeng areas, the Springbok Flats between Pretoria and the Limpopo Province and in the coastal plains of the Western Cape Province (Goudie, 1973; Netterberg, 1969b). Calcrete also occurs around the margins of the Namib Desert, in southern Angola (Halpenny, 1957), and along the Changane River in Mozambique (Du Toit, 1954) and the South Africa-Zimbabwe border (Goudie, 1973). Calcretes are also widespread in semi-arid areas of southwestern Madagascar (Besarie, 1948).

The majority of calcretes are white, cream or grey in colour, though pink mottling and banding is common. They exhibit a variety of forms, including weakly calcified, chalky, powdery, rhizocretionary, nodular, glaebular, honeycomb, platy, laminar, stringer, pisolitic, brecciated, conglomeratic, massive and hardpan (Wright, 2007), all of which have been documented in southern Africa. In pedogenic calcretes, many of these morphologies fall within an idealised evolutionary continuum (Figure 8.6), which has been used as a tool for comparing the relative development, or *stage*, of profiles (Gile *et al.*, 1966; Netterberg, 1969b; 1980; Reeves, 1970; Bachman and Machette, 1977; Goudie, 1983; Netterberg and Caiger, 1983; Machette, 1985). In southern Africa, the scheme for describing calcrete profiles developed by Netterberg (1980) and Netterberg and Caiger (1983) has been applied widely. This recognises the geotechnical properties of calcretes at different stages of formation, ranging from calcareous soil through calcified soil, powder calcrete, nodular calcrete, honeycomb calcrete to hardpan calcrete. However, the scheme proposed by Bachmann and Machette (1977) to describe North American calcretes is now more accepted globally (Table 8.2). Under this scheme, Stage I calcified soils and chalky or powder calcretes may develop into Stage II nodular calcretes as calcium carbonate concretions increase in size. These concretions may coalesce to form Stage III honeycomb calcrete, with a Stage IV-V hardpan calcrete developing as surface horizons become plugged with calcium carbonate. Solutional degradation of a hardpan may lead to the development of a Stage VI brecciated (or boulder) calcrete. Laminar calcretes consisting of finely banded carbonate often cap hardpan and brecciated calcretes (James, 1972; Klappa, 1979; Arakel, 1982; Warren, 1983; Verrecchia *et al.*, 1995), and the whole sequence may be buried by upper soil horizons containing carbonate pisoliths (Wright, 2007). Well-developed profiles are usually between 1.0 m and 5.0 m thick. Hardpans are most commonly in the range 0.3-0.5 m thick (Goudie, 1984). Laminar calcrete zones are rarely more than 0.25 m thick (Goudie, 1983), but some forms can reach 2.0 m (Wright, 2007). This sequence represents an idealised evolutionary model. In reality, mature pedogenic calcrete profiles (Figure 8.7) often consist of a laminar zone developed on a massive hardpan overlying a zone of nodules and chalky carbonate (Netterberg, 1969c; 1971; 1980; Arakel, 1982; Goudie, 1983).

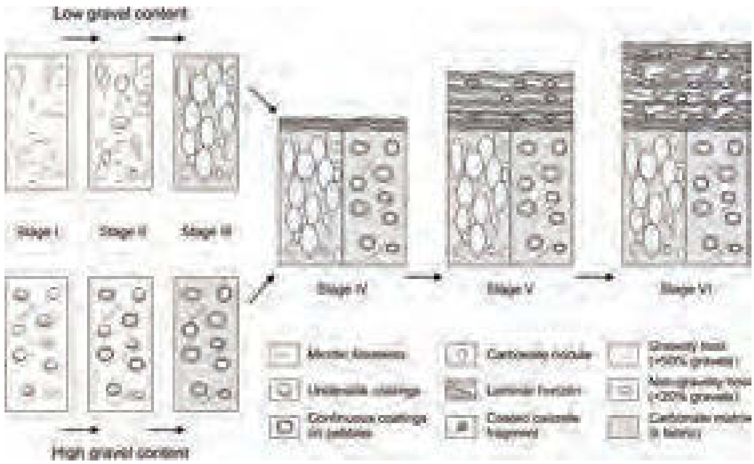

Figure 8.6. Stages in the development of a pedogenic calcrete profile developed in gravel-poor and gravel-rich sediments (based on Machette, 1985 and Alonso-Zarza, 2003).

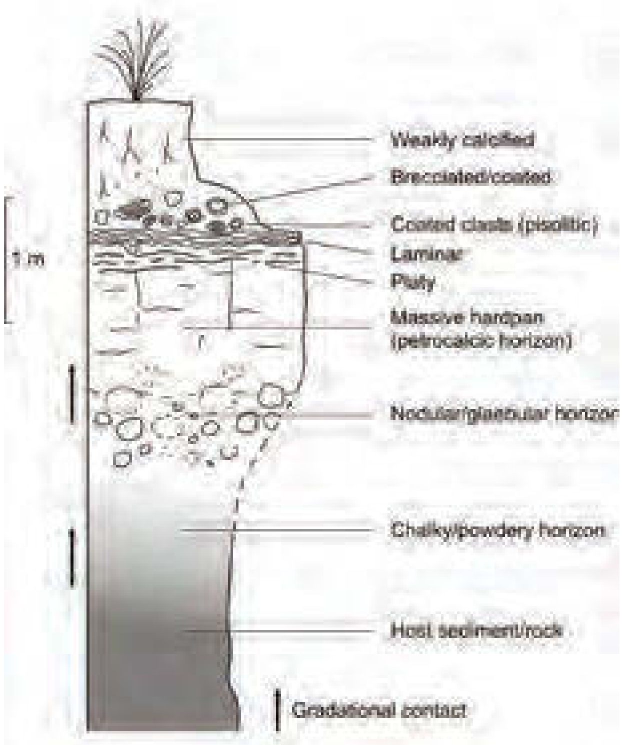

Figure 8.7. Idealised pedogenic calcrete profile showing a range of macroforms (after Wright, 2007).

Table 8.2. Stages in the morphogenetic sequence of carbonate deposition during calcrete formation (Bachman and Machette, 1977).

STAGE	DIAGNOSTIC CARBONATE MORPHOLOGY
I	Filaments or faint carbonate coatings, including thin discontinuous coatings on the underside of pebbles.
II	Firm carbonate nodules few to common but isolated from one another. The matrix between nodules may include friable interstitial carbonate accumulations. Continuous pebble coatings present.
III	Coalesced nodules in disseminated carbonate matrix.
IV	Platy, massive indurated matrix, with relict nodules visible in places. The profile may be completely plugged with weak incipient laminar carbonate coatings on upper surfaces. Case hardening is common on vertical exposures.
V	Platy to tabular, dense and firmly cemented. Well-developed laminar layer on upper surfaces. Scattered incipient pisoliths may be present in the laminar zone. Case hardening common.
VI	Massive, multilaminar and brecciated profile, with pisoliths common. Case hardening common.

Pedogenic calcretes have been described from various borrow pits for road construction and archaeological sites across southern Africa (e.g. Watts, 1980; Netterberg, 1969c; 1980). However, the majority of calcretes exposed within and around drainage features are probably non-pedogenic in origin (Nash *et al.*, 1994a) or are polygenetic, incorporating complex pedogenic and non-pedogenic carbonate signatures (Watts, 1980; see also Alonso-Zarza, 2003). Attempting to describe these types of calcrete using the Bachman and Machette (1977) scheme shown in Table 8.2 is problematic since non-pedogenic calcretes do not normally exhibit the well-organised profile structure of calcretes formed in soils, are frequently massive, and may not go through the same stages of development (Arakel and McConchie, 1982; Arakel, 1986; 1991; Jacobson *et al.*, 1988; Khadkikar *et al.*, 1998; Tandon and Andrews, 2001). Non-pedogenic calcretes also appear to develop much more rapidly than pedogenic varieties (Nash and Smith, 1998), so a densely cemented groundwater calcrete may represent carbonate accumulation over a much shorter timescale than a pedogenic hardpan of equivalent thickness and induration. In these situations, the relatively simple scheme put forward by Netterberg (1980) and Netterberg and Caiger (1983) may be more appropriate.

Goudie (1973) showed from a sample of 300 bulk chemical analyses, including many from southern Africa, that calcretes on average comprised 79.28% calcium carbonate (42.62% CaO), 12.30% silica, 3.05% MgO, 2.03% Fe_2O_3 and 2.12% Al_2O_3. Further data for the subcontinent are included within Netterberg (1969c). These averages mask considerable chemical variability, with pedogenic hardpan calcretes typically the most calcareous and powder calcretes the least. CaO levels also decline down-profile within pedogenic calcretes, with a parallel rise in the proportion of MgO (Hutton and Dixon, 1981). In contrast, CaO contents of many non-pedogenic calcretes may be remarkably homogenous throughout a profile (e.g. Nash and Smith, 1998). Carbonate mineralogy within calcretes is dominated by low-Mg calcite (Wright and Tucker, 1991) with variable amounts of dolomite present. If abundances of diagenetic dolomite are high, a duricrust may be classified as a dolocrete rather than a calcrete (see Hay and Reeder, 1978; Hay and Wiggins, 1980; Hutton and Dixon, 1981). Significant proportions of high-Mg calcite have been reported in pedogenic calcretes from various parts of southern Africa (Kalkowsky, in Passarge, 1904; Netterberg, 1969a; 1980), including the Kalahari where variations in the percentage of high-Mg calcite and dolomite appear to be related to the occurrence of authigenic silica and silicates (Watts, 1980). Quartz is the most important non-carbonate mineral in most calcretes, both in bulk samples and in the clay size fraction. Silica may also be present as opal and chalcedony, as a result of mineral replacement or precipitation during diagenesis (as noted in samples from the Kalahari;

Watts, 1980; Nash and Shaw, 1998; Nash and McLaren, 2003), or within diatoms (Netterberg, 1969c; 1971; Nash and McLaren, 2003). Aragonite is also noted in some calcretes, as are the clay minerals palygorskite, sepiolite, illite, kaolinite, montmorillonite-smectite and chlorite (Netterberg, 1969c; Watts, 1980).

The formation of both pedogenic and non-pedogenic calcrete requires a source of calcium carbonate, a mechanism for transferring this carbonate to the site of calcrete development and a means of triggering carbonate precipitation. The range of potential carbonate sources and precipitation mechanisms is common to the majority of calcrete types. Carbonate sources can include solid and dissolved carbonate introduced into the soil or sediment from above (e.g. atmospheric dust, rainfall, volcanic ash, sea-spray, surface runoff, plant materials and shells) or below (e.g. groundwater and weathered bedrock) (Netterberg, 1969c; Goudie, 1983; Cailleau *et al.*, 2004). The contribution of specific sources varies spatially and with calcrete type; atmospheric dust and rainfall inputs are most important in pedogenic calcretes (Capo and Chadwick, 1999; Chiquet *et al.*, 1999) and subsurface sources for non-pedogenic calcretes (e.g. Arakel and McConchie, 1982; Arakel, 1986, 1991; Nash and McLaren, 2003; Nash and Smith, 2003). Calcium carbonate dissolved in water may be precipitated by a variety of processes including evaporation, evapotranspiration, organic life processes, an increase in pH to above 9.0 (Goudie, 1983), a decrease in the partial pressure of soil CO_2 (Schlesinger, 1985), CO_2 loss by degassing (e.g. as temperature increases; Barnes, 1965), and the common ion effect (Wigley, 1973). Organic agency is important in the formation of some calcretes, with calcrete laminae, in particular, suggested to be of organic origin (e.g. Klappa, 1979; Wright, 1989; Verrecchia *et al.*, 1991). Termites may also fix calcite in soils, as can earthworms and slugs (Canti, 1998; Monger and Gallegos, 2000).

The mechanism by which carbonate is transferred to the site of precipitation is the main distinguishing factor in pedogenic and non-pedogenic calcrete genesis. Pedogenic calcretisation is characterised by predominantly vertical transfers of dissolved carbonate within the soil or sediment profile (the *per ascensum* and *per descensum* models of Goudie, 1983). Formation can take place either through a simple, progressive illuviation process, or a more dynamic one if carbonate build-up is interrupted by erosion (Wright, 2007). In the simple case, which applies to the majority of southern African pedogenic calcretes, formation proceeds via the downward translocation, precipitation and accumulation of carbonate over time (although there may be phases of dissolution and reprecipitation if the host soil or sediment is carbonate-rich). Profile formation may also take place by the accumulation of calcified root mats (Klappa, 1980; Wright *et al.*, 1995), although this is less common in southern Africa (see Netterberg, 1980). If the host material is carbonate-rich and well-indurated (e.g. limestone bedrock), calcretisation may take place along a progressively downward migrating alteration front. The profile may break up as carbonate progressively builds up and displacive calcite crystallisation occurs, with features such as pseudo-anticlines and brecciated cobbles forming (these are well-documented in the Kalahari; Watts, 1977; 1978; 1980). This is relatively common in non-carbonate host sediments as calcite is unable to form adhesive bonds at the molecular scale with non-carbonate materials (Chadwick and Nettleton, 1990).

In contrast, much of the carbonate for non-pedogenic calcrete formation is provided by lateral transfers of carbonate-rich surface or groundwater (e.g. within lacustrine or channel-margin settings). Less research has been undertaken into the origins of these types of calcrete and their terminology remains confused. Non-pedogenic calcretes are frequently referred to as *groundwater calcretes*, a term often used interchangeably with phreatic, channel, valley or alluvial fan calcrete. These are, however, distinct calcrete types. Groundwater calcretes (*sensu stricto*) form by precipitation in the capillary rise zone directly above the water table (Wright, 2007), although Jacobson *et al.* (1988) suggest from studies in Australia that precipitation can also occur below this level. Phreatic calcretes form as a result of cementation at or below the water table (Arakel, 1986; Wright and Tucker, 1991). Calcretes developed within drainage lines may form ribbon-like bodies extending many hundreds of kilometres in length.

Nash and McLaren (2003) distinguish between valley calcretes, which partially cement alluvium within broad, shallow, drainage courses, and channel calcretes which cement sediments within confined impermeable bedrock channels and may occupy the full channel cross-section. Channel calcretes have not been described in southern Africa. However, valley calcretes are commonly exposed as low fluvial terraces in a number of the fossil drainage systems of the central Kalahari (Figure 8.5b). These calcretes reach thicknesses in excess of four metres in outcrop (Nash and McLaren, 2003), with carbonate-cemented sediments extending tens of metres beneath the floors of some valleys (Nash *et al.*, 1994a). Detailed petrological analyses (Figure 8.8) indicate that these calcretes formed relatively rapidly at the end of the Pleistocene as a result of carbonate precipitation within and above the capillary fringe zone. The carbonate was presumably supplied by flushes of water along valleys and through valley alluvium, with calcrete development driven by evaporation. A similar origin has been suggested for calcretes exposed within strandlines around the Makgadikgadi Depression, with carbonate supplied mainly via groundwater during relatively wetter periods and precipitated in lake marginal settings during drying phases (Ringrose *et al.*, 2005; 2009).

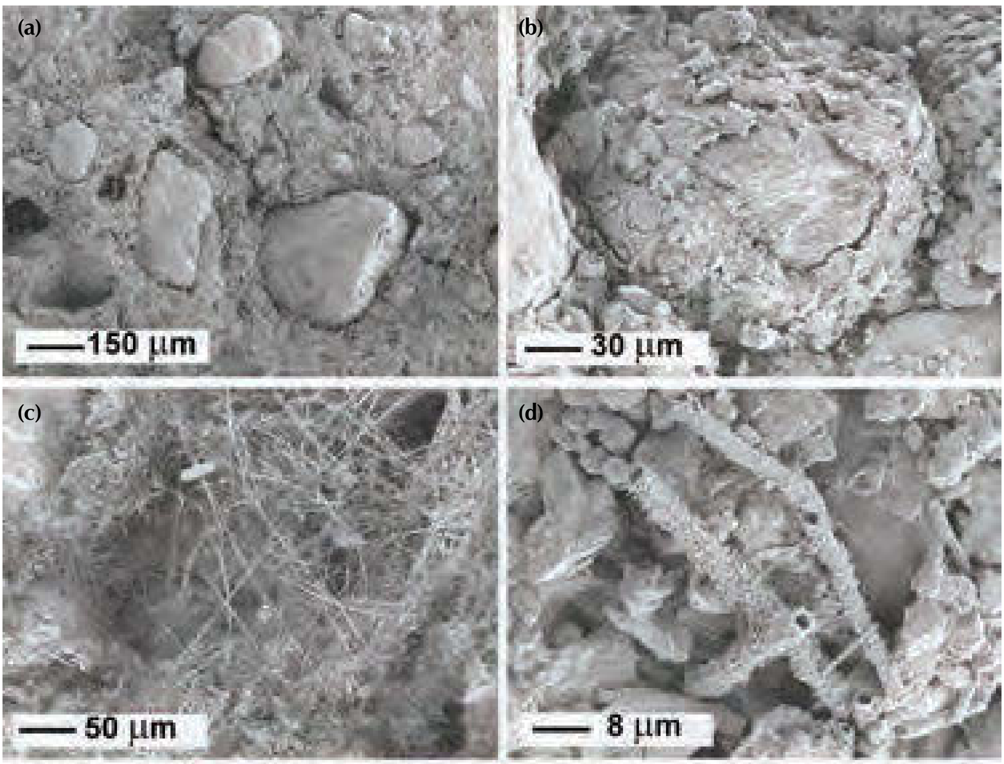

Figure 8.8. Scanning electron microscope images of cements from non-pedogenic valley calcretes exposed in the Okwa and Hanehai Valleys in the central Kalahari, Botswana: a) rounded quartz grains and shell fragments cemented by microcrystalline calcite; b) quartz grain coated by microcrystalline calcite and microspar cement; c) hollow fungal filaments encrusted with microcrystalline and small bladed and acicular calcite crystals; d) needle-fibre and calcified fungal filaments (plus possible fruiting spores) lining a void with needle-fibres crossing the void. Images (a) and (b) represent typical cements developed under evaporitic conditions whilst (c) and (d) indicate organic agency under vadose conditions.

Calcretes of differing origins can form within the same sediment profile. Nash and Smith (1998), for example, describe alluvial fan sediments from southeast Spain that contain thick groundwater calcretes in their basal sections and pedogenic calcretes on their upper surfaces. Calcretes of mixed genesis can also occur (Alonso-Zarza, 2003), with non-pedogenic forms developing pedogenic features as hydrological conditions change (Mann and Horwitz, 1979; Arakel, 1982; 1986). Evidence from many predominantly non-pedogenic Kalahari valley calcretes indicates, for example, that as valley water tables fell towards the end of the Pleistocene, vadose pedogenic and biological calcite cements and late-stage silica cements were precipitated in pore spaces and desiccation cracks to produce a polygenetic profile (Nash and McLaren, 2003).

4. Silcrete

After Australia, southern Africa contains the second greatest areal coverage of silcrete in the world. Langford-Smith (1978) describes *silcrete* as a

> *... brittle, intensely indurated rock composed mainly of quartz clasts cemented by a matrix which may be a well-crystallised quartz, or amorphous (opaline) silica (Langford-Smith, 1978:3).*

Cryptocrystalline silica and chalcedony, a mineral consisting of nanoscale intergrowths of quartz and the fibrous silica polymorph moganite (Heaney, 1993; 1995), are also important components of many silcrete cements (Nash and Hopkinson, 2004). Silcrete forms in near-surface environments as a result of the cementation or replacement of rock, sediment, saprolite or soil by secondary silica (Milnes and Thiry, 1992). Significantly, silicification proceeds via low-temperature physico-chemical processes; this distinguishes silcrete from silica-cemented sedimentary rocks such as quartzite (Summerfield, 1981; 1983a, b; Milnes and Thiry, 1992). For a recent review, see Nash and Ullyott (2007).

Given their hardness and chemical stability, silcretes are extremely resistant to erosion and often form cap-rocks on residual hills, mesas or escarpments. Exposures in low-lying parts of the landscape are relatively rare, although *in situ* outcrops have been described within valleys and around ephemeral lakes (e.g. Summerfield, 1982; Taylor and Ruxton, 1987; Thiry *et al.*, 1988; Nash *et al.,* 1994a,b; 2004). Exhumed silcretes may also outcrop as layers part way up slopes (Mountain, 1951; Young, 1978; Thiry *et al.*, 1988; Milnes *et al.* 1991). Silcretes most commonly occur as distinct horizons, but may also form a coating on rock outcrops (Hutton *et al.*, 1972; Smale, 1973; Mišík, 1996) or lenses within other duricrusts (e.g. Wright, 1963; Alley, 1977; Langford-Smith and Watts, 1978; Summerfield, 1981; 1982; Nash *et al.*, 1994a). Well-developed silcrete horizons are between 0.5 and 3 m thick, although thicknesses of more than 7 m have been documented in the Kalahari (Nash *et al.*, 1994b) and more than 15 m in the Paris Basin (Thiry and Simon-Coinçon, 1996). A variety of terms have been used to describe silcrete profiles, including massive, columnar, bulbous, nodular, glaebular and mammilated. Root casts, preserved plant material, and ant or termite burrows are also documented. The colour of silcrete is highly variable, ranging from grey, white, buff, brown and dull red to bright green (Nash and Ullyott, 2007), with all of these variations seen in southern Africa (Du Toit, 1954).

By definition, silcretes are chemically simple (Table 8.3), comprising >85% silica by mass (Summerfield, 1983a), with some exceptionally *pure* examples consisting of >95% silica (Nash *et al.*, 1994a). Minor amounts of titanium, iron and aluminium oxides and resistate trace elements make up the remainder of most samples (Summerfield, 1983a). Titanium is usually present as anatase and may be disseminated through the silcrete cement or concentrated within geopetal structures. TiO_2 may be enriched to >1% (Summerfield, 1983d; Thiry and Millot, 1987) and exceed 20% in some silcrete *skins* (Hutton *et al.*, 1972). In southern Africa, silcretes in the Cape coastal zone of South Africa have typically higher TiO_2 contents compared to inland varieties (Summerfield, 1983c), although the palaeoenvironmental significance of this difference has been disputed (Nash *et al.*, 1994b). Alumina content usually reflects the presence of illuviated, inherited or authigenic clays. Haematite and goethite may be retained during silicification (Meyer and Pena dos Reis, 1985; Bustillo and Bustillo, 1993; Armenteros *et al.*, 1995; Ballesteros *et al.*, 1997). Glauconite has been reported in some Kalahari silcretes (e.g. Smale, 1973; Summerfield, 1982) and may indicate sub-oxic, partially reducing, groundwater conditions during silicification (Nash *et al.*, 2004).

Table 8.3. Bulk chemistry of silcretes in the Kalahari and Cape coastal zone (analyses by x-ray fluorescence).

	REGION			
	Kalahari[a] Mean %	n = 48 SD[c]	Cape Coastal[b] Mean %	n = 66 SD
SiO_2	91.63	4.04	95.04	2.54
TiO_2	0.12	0.06	1.79	0.58
Al_2O_3	1.69	0.98	0.61	0.46
Fe_2O_3	0.86	0.54	1.28	1.70
MnO	0.01	0.00	0.01	0.00
MgO	0.98	0.66	0.28	0.20
CaO	0.85	1.51	0.13	0.37
Na_2O	0.36	0.35	no data	no data
K_2O	0.95	0.86	0.05	0.12
P_2O_5	0.01	0.00	0.04	0.03
SO_3	no data	no data	no data	no data
LOI[d]	2.80	1.51	0.99	0.67

Notes:

a Calculated from analyses in Summerfield (1982), Nash and Shaw (1998) and Nash *et al*. (1994b; 2004).

b Calculated from analyses in Frankel and Kent (1938), Bosazza (1939), Frankel (1952) and Summerfield (1983d).

c Standard deviation.

d Loss on ignition.

A number of classification schemes have been developed to distinguish the various types of silcrete (e.g. Goudie, 1973; Smale, 1973; Wopfner, 1978; Summerfield, 1983a, b). The most recent attempt at classification, building upon the scheme proposed by Milnes and Thiry (1992), is shown in Figure 8.9 (after Nash and Ullyott, 2007). Echoing the schemes used to classify gypcretes and calcretes, this subdivides silcrete into pedogenic and non-pedogenic categories. Pedogenic silcrete are divided into those with abundant microquartz and titania that lack clay minerals and iron oxides, and more opaline *duripans* in which clay and iron are retained (Thiry, 1999). Non-pedogenic silcretes are grouped into groundwater, drainage-line and pan/lacustrine types on the basis of their geomorphological context.

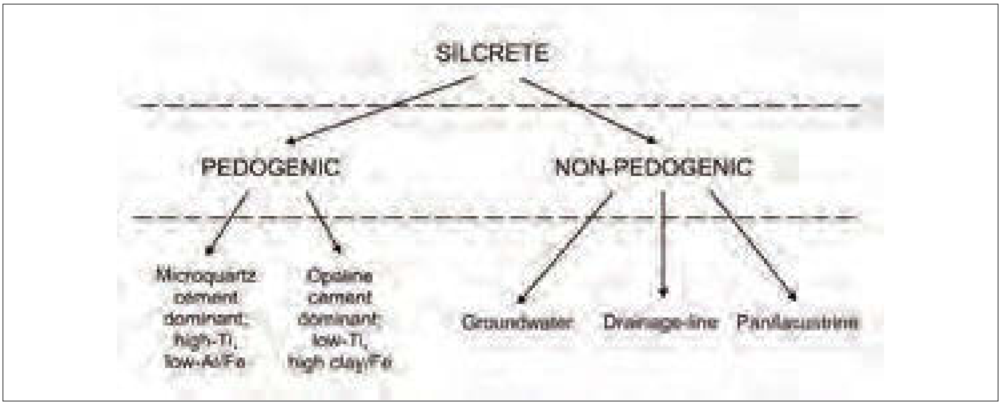

Figure 8.9. Geomorphological classification of silcrete (after Nash and Ullyott, 2007).

Silcrete occurs in two distinct sedimentary settings in southern Africa (Du Toit, 1954): as extensive cappings, 5 m or more in thickness, developed on deeply weathered and dissected remnants of the African Surface Complex in the Cape coastal zone (Marker and McFarlane, 1997; Roberts, 2003; Figures 8.1a and 8.10a), and as part of the terrestrial Kalahari Group sediments in the interior Kalahari Basin (Summerfield, 1983b; Haddon and McCarthy, 2005; Figure 8.10b-d). In the Cape coastal zone, silcrete has been documented from the Oliphants River Valley in the Western Cape to East London, but is particularly well-developed around Grahamstown, Riversdale and Albertinia (Bosazza, 1936; 1939; Frankel and Kent, 1938; Mountain, 1946; 1951; 1980; Frankel, 1952; Smale, 1973; Summerfield 1981; 1983a-d; 1984; Marker and McFarlane, 1997; Roberts *et al.*, 1997; Marker *et al.*, 2002; Roberts, 2003). It is found both overlying, and at depth within, a range of weathered lithologies, including shales, phyllites and tillites, as part of deeply weathered profiles in excess of 20 m thickness. In a few localised occurrences, silcrete consists of silicified sand (along the Oliphants River Valley) and Witteberg Quartzite (on the coastal plain south of Grahamstown; Summerfield, 1981; 1983b). Silcrete is often found in association with laterite, particularly towards the eastern end of the Cape coast (Marker and McFarlane, 1997), and more rarely with calcrete (Summerfield, 1981). Silcrete outcrops range from extensive horizontal sheets to isolated lenses, and from single layers to multiple units with up to four discrete horizons (Smale, 1973; Summerfield, 1983b). A range of morphologies is also present; many outcrops exhibit well-developed columnar structures (Figure 8.10a), whilst others are more massive and structureless. The majority of silcretes in the Cape coastal zone are early to middle Cenozoic in age (see Summerfield, 1981, and references therein). Silicified Permo-Triassic wood fragments are present within silcretes from near East London, but the fossil wood has been shown to be reworked into Tertiary fluvial debris-flow deposits that were subsequently silicified (Roberts *et al.*, 1997).

Silcretes in the Kalahari Basin have been described by a variety of authors (e.g. Passarge, 1904; Boocock and Van Straten, 1962; Goudie, 1973; Smale, 1973; Mallick *et al.*, 1981; Summerfield, 1982, 1983b,c; Shaw and De Vries, 1988; Shaw *et al.*, 1990; Nash *et al.*, 1994a,b; 2004; Harrison and Shaw, 1995; Nash and Shaw, 1998; Shaw and Nash, 1998; Nash and Hopkinson, 2004; Haddon and McCarthy, 2005; Ringrose *et al.*, 2005; 2009). Four principal modes of occurrence have been identified:

i. silica cementation of sands;

ii. silicification of terrace deposits related to a present or former drainage channel;

iii. silicification of sediments marginal to pans; and

iv. the replacement silicification of calcrete.

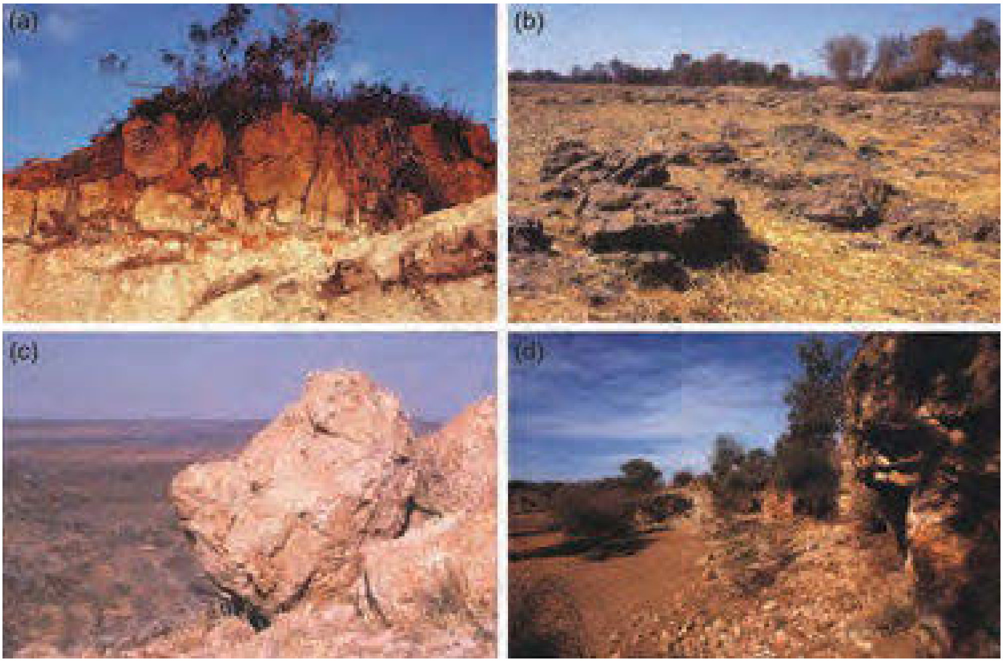

Figure 8.10. Examples of silcrete profiles from the Cape coastal zone and the Kalahari: (a) *In situ* pedogenic silcrete profile developed within deeply weathered Bokkeveld Shale bedrock, near Enniskillin, Western Cape Province, South Africa; (b) Massive drainage-line silcrete at Samedupe Drift in the floor of the ephemeral Boteti River, Botswana; (c) Silcrete caprock overlying basalt at the eastern margin of the Kalahari Group sediments, Sesase Hill, Botswana; (d) Silcrete outcropping in the flanks of the Moshaweng Valley south of Letlhakeng, Botswana.

Silcretes in the Kalahari are most commonly exposed in association with drainage-lines, lakes and pans (Figure 8.10b, d). On the basis of their structure, these silcretes are most probably non-pedogenic in origin. More rarely, silcretes occur as escarpment caprocks, for example in the Serowe area of east-central Botswana (see Smale, 1973; Summerfield, 1982; Nash and Hopkinson, 2004). Escarpment silcretes at the eastern margin of the Kalahari Basin are frequently situated on top of weathered bedrock, particularly basalt, and may be pedogenic in origin (Figure 8.10c). If this is the case, then the Kalahari may mirror the pattern of silcrete distribution found in the Paris Basin (France) and the Eromanga Basin (Australia), with pedogenic silcretes at the margin of the sedimentary basin and groundwater or drainage line silcretes towards the basin centre (Ullyott *et al.*, 1998). Kalahari silcretes are frequently found in close association with other duricrusts, notably calcrete and occasionally ferricrete. These relationships may be laterally or vertically gradational (Summerfield, 1982; Shaw and De Vries, 1988; Nash *et al.*, 1994a, b; 2004). Silcretes in the Kalahari range from those with very simple mineralogy and structure to those where there are indications of polyphase development (Nash *et al.*, 1994b). Simple silcretes occur, for example, within the Boteti River at Samedupe Drift (Shaw and Nash, 1998), whilst the Letlhakeng area (Shaw and De Vries, 1988; Nash *et al.*, 1994a, b) and Okwa Valley (Nash *et al.*, 1994b; 2004) contain a range of simple to polyphase or intergrade silcrete types. The ages of silcretes in the Kalahari remain unknown. However, silcretes from Sua Pan, Botswana (Harrison and Shaw, 1985) are possibly contemporary as they are cemented by abundant poorly crystalline opal-CT or opal-T and incorporate relatively high amounts of molecular water (Nash and Hopkinson, 2004).

Silcrete development can take place via a variety of processes, but, in common with other duricrusts, all require a silica source, a means of transferring this silica to the site of formation, and a mechanism to trigger precipitation. The most significant silica source is chemical weathering of silicate-rich rocks, particularly those containing clay minerals (Summerfield, 1983a). Silica released in this way may then be available for transport in solution, usually in the form of undissociated monosilicic acid (Dove and Rimstidt, 1994). The silica required to form the thick silcrete horizons in the Albertinia area of the southern Cape, for example, may have been generated by the desilicification of a deeply kaolinised weathering profile, with re-precipitation occurring at depth (Marker *et al.*, 2002). Highly alkaline conditions, such as those found in arid zone lakes, can also lead to the dissolution of silicate minerals. Other important sources include replacement of silica during carbonate precipitation, dissolution of volcanic and other dust, and biological inputs from silica-rich plants and microorganisms. Silica from these sources may be transferred in solution to the point of precipitation via lateral or vertical movements of groundwater, porewater and surface water, with a range of local and far-travelled silica potentially contributing to silcrete formation (Stephens, 1971; Hutton *et al.*, 1972; 1978). Silica precipitation may be initiated by a variety of factors, of which the most important are evaporation, cooling, organic processes, absorption by solids, reactions with cations, and changes in pH, particularly a shift to below pH 9.0 in alkaline environments (*see* Nash and Ullyott, 2007).

Pedogenic silcretes form as a result of cycles of downward flushing of silica-bearing water followed by silica precipitation, possibly under strongly seasonal tropical or sub-tropical climates with high rates of evaporation (Webb and Golding, 1998). This generates gravitational silica structures along water pathways, together with glaebules and rootlets, and may be associated with evidence of bioturbation (Summerfield, 1983d; Thiry *et al.*, 2006; Terry and Evans, 1994; Lee and Gilkes, 2005). The time period required for a mature pedogenic silcrete to develop is $>10^6$ years (Thiry, 1978; Callen, 1983; Milnes and Thiry, 1992). Pedogenic silcrete profiles are usually complex and exhibit two sections (Figure 8.11). The upper part commonly has a columnar structure and well-ordered silica cements and may be surmounted by a nodular or pseudo-brecciated layer, whilst the lower is weakly cemented by poorly ordered silica (Milnes and Thiry, 1992). Many of the columnar silcretes from the Cape coastal zone exhibit this profile structure (Figure 8.8a) and, by analogy, are pedogenic in origin. The suggestion that they formed during a "prolonged period of desiccation" (Partridge and Maud, 1989:429) as part of an arid zone sodic soil (Partridge, 1997; 1998) is, as such, highly improbable.

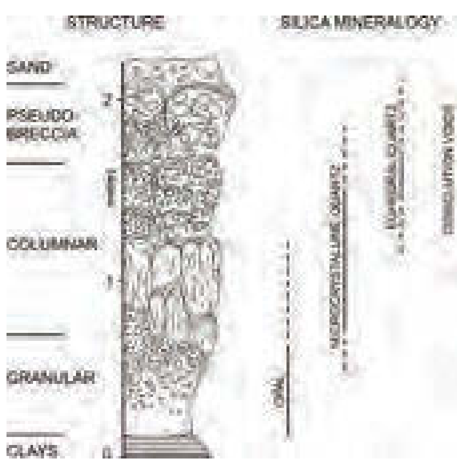

Figure 8.11. Schematic representation of a *typical* pedogenic silcrete profile, showing silcrete structures and the vertical arrangement of silica cements (after Milnes and Thiry, 1992; Thiry, 1999).

Groundwater silcretes occur typically as discontinuous lenses or sheets that may be superposed. They can form near the surface, but also develop at depths of up to 100 m, commonly underlying and sometimes overlying pedogenic silcrete horizons (Thiry, 1999; Basile-Doelsch *et al.*, 2005). They lack the organised profile of pedogenic types, and exhibit simple fabrics with good preservation of host structures (Ullyott *et al.*, 2004; Ullyott and Nash, 2006). Formation is normally related to a present or former water table, with silicification occurring under phreatic conditions (Callen, 1983) at the water table or near groundwater outflow zones (Thiry *et al.*, 1988). Water table fluctuations have also been proposed as a control (e.g. Taylor and Ruxton, 1987; Milnes *et al.*, 1991; Thiry and Milnes, 1991; Rodas *et al.*, 1994).

Drainage-line silcretes are closely related to groundwater varieties but develop within alluvial fills in current or former drainage networks as opposed to within unconfined bedrock (Nash and Ullyott, 2007). Silcretes develop at locations that are subject to seasonal wetting/drying or in zones of water table fluctuation (Taylor and Ruxton, 1987; McCarthy and Ellery, 1995). Few genetic models for drainage-line silicification exist. Silcretes exposed in the Boteti River, Botswana (Figure 8.10b), however, are suggested to have formed by the accumulation within channel alluvium of phytolithic and clastic silica from floodwater, and dissolved silica from groundwater, with silica precipitation triggered by evapotranspiration from seasonal pools (Shaw and Nash, 1998). Massive silcrete layers buried within the channel alluvium (Figure 8.12a) are suggested to have developed in response to salinity shifts associated with movements of the wetting front during flood events.

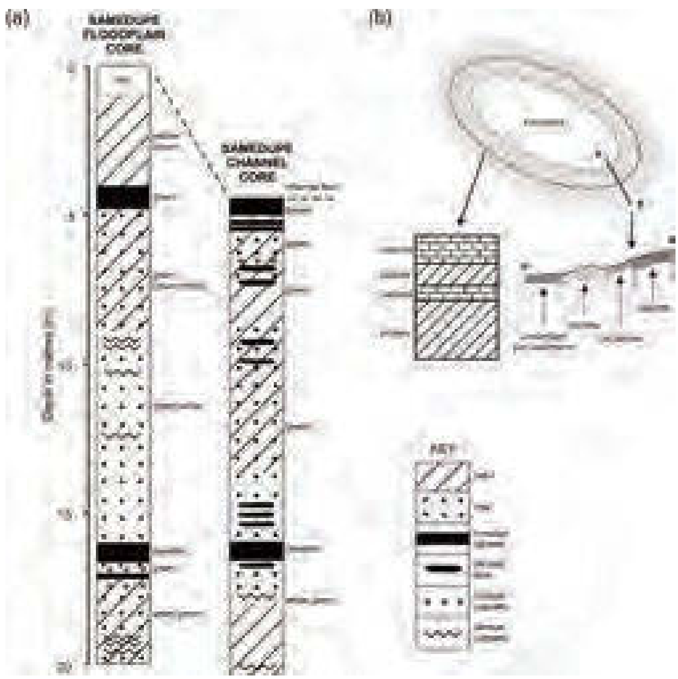

Figure 8.12. a) Cores extracted from the bed of the Boteti River at Samedupe Drift, Botswana, showing a range of geochemical sediments developed beneath the channel floor including massive and pisolithic drainage-line silcretes (after Shaw and Nash, 1998); b) Schematic representation of geochemical sedimentation patterns in the vicinity of a pan or playa (after Summerfield, 1982).

Pan/lacustrine silcretes (Figure 8.12b) develop within, or adjacent to, ephemeral lakes, pans or playas (Goudie, 1973), with phases of silica mobility and precipitation driven by changes in pH and salt concentration during annual cycles of flooding and evaporation (Summerfield, 1982). The zone of maximum silica precipitation occurs around the lake margin and immediately above the water table (Thiry, 1999). Biological fixing of silica may also occur. Sua Pan in Botswana contains silcretes developed by colonies of the silica-fixing cyanobacteria *Chloriflexus* (Shaw *et al.*, 1990). Pan/lacustrine silcretes may contain a variety of minerals in addition to silica, depending upon the chemistry of the lacustrine sediments. Glauconite-illite occurs within green pan silcretes in the Kalahari (Netterberg, 1969c; Summerfield, 1982). Silica replacement is commonplace in such environments and may lead to the development of intergrade cal-silcrete and sil-calcrete crusts (Nash and Shaw, 1998).

5. Laterite and ferricrete

Laterites and ferricretes are, in the broadest sense, iron- (and often aluminium-) rich sub-aerial weathering products that develop as a result of intense substrate alteration under tropical or sub-tropical climates. The terms *laterite* and *ferricrete* are often (incorrectly) used interchangeably and there has been considerable terminological debate as to the differences between the two materials. Following Aleva (1994), laterites *sensu stricto* are considered here to be residual materials formed by *in situ* rock breakdown that do not contain significant externally derived components. The high iron content of ferricretes, in contrast, is achieved via the net input of externally derived iron into a host rock or existing weathering profile rather than through residual enrichment. Laterites occur in two varieties: *pedogenetic laterite* that develops within vadose profiles and *groundwater laterite* that forms within the range of oscillation of the groundwater-table (McFarlane, 1983). Using Aleva's (1994) definition, laterites owe their composition to the "relative enrichment" of iron (and aluminium) and other less mobile elements during weathering, whereas the iron within a ferricrete is derived predominantly by "absolute accumulation" (D'Hoore, 1954). This process-based definition provides a useful distinction between the two duricrusts. However, as Widdowson (2007) notes, determining the origin of mineral components within a profile can be problematic; lateritic profiles may be modified by the introduction of externally derived materials and, once formed, ferricretes can be exposed to weathering processes and evolve toward more lateritic-type profiles.

Lateritic duricrusts occur typically as the uppermost layers of *in situ* tropical-type weathering profiles (Figure 8.13a). They are best developed in landscapes characterised by almost horizontal surfaces, such as those formed by long-term erosion, which limit surface runoff. When such surfaces are exposed to chemical weathering over lengthy time periods, the more mobile mineral constituents are evacuated from the weathering profile (see Widdowson, 2007 for a detailed discussion of the mechanisms involved), leaving behind residual accumulations of chemically resistant materials closest to the surface (McFarlane, 1976; 1983). This is accompanied by gradual land-surface lowering. As the weathering front extends deeper into the landscape (Figure 8.14), the residua produced at each stage of weathering effectively provide the parent material for subsequent stages, resulting in the progressive accumulation of the most highly resistant residual Fe- (and Al-) oxyhydroxides in the uppermost parts of the weathering profile (McFarlane, 1983). Many (if not most) lateritic crusts may have developed over 10^6-10^7 years and experienced phases of surface erosion and/or variations in the degree of weathering over this time (Nahon, 1986; Thomas, 1994). Ferricretes may develop as accumulations at the base of slopes, or within topographic depressions such as valleys, and may incorporate transported materials from sources beyond the immediate site of duricrust development.

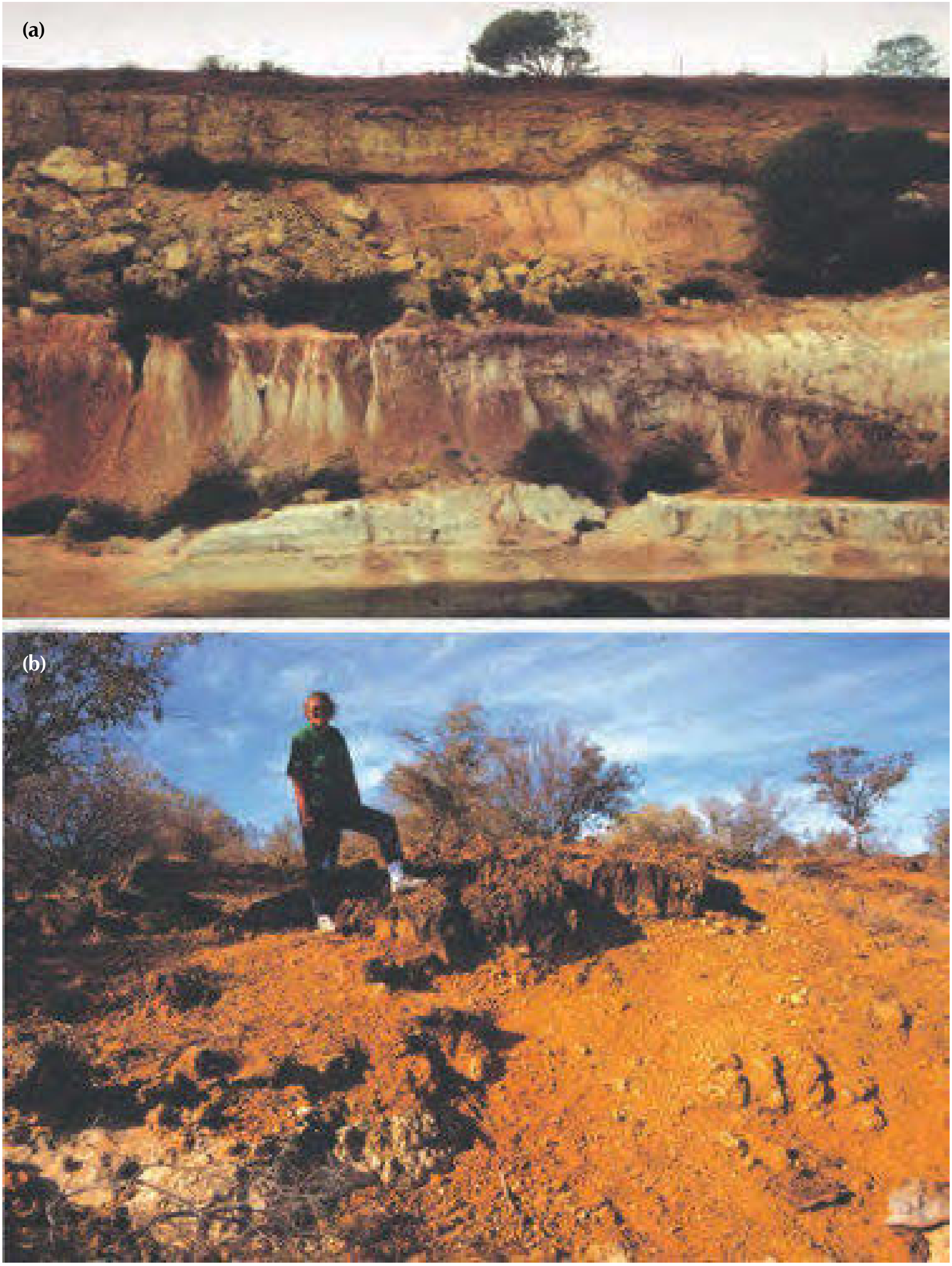

Figure 8.13. a) Deep weathering profile developed within Bokkeveld Shale north of Albertinia, South Africa; b) Vermiform ferricrete developed on top of calcrete in the flanks of the Gaotlhobogwe Valley south of Letlhakeng, Botswana.

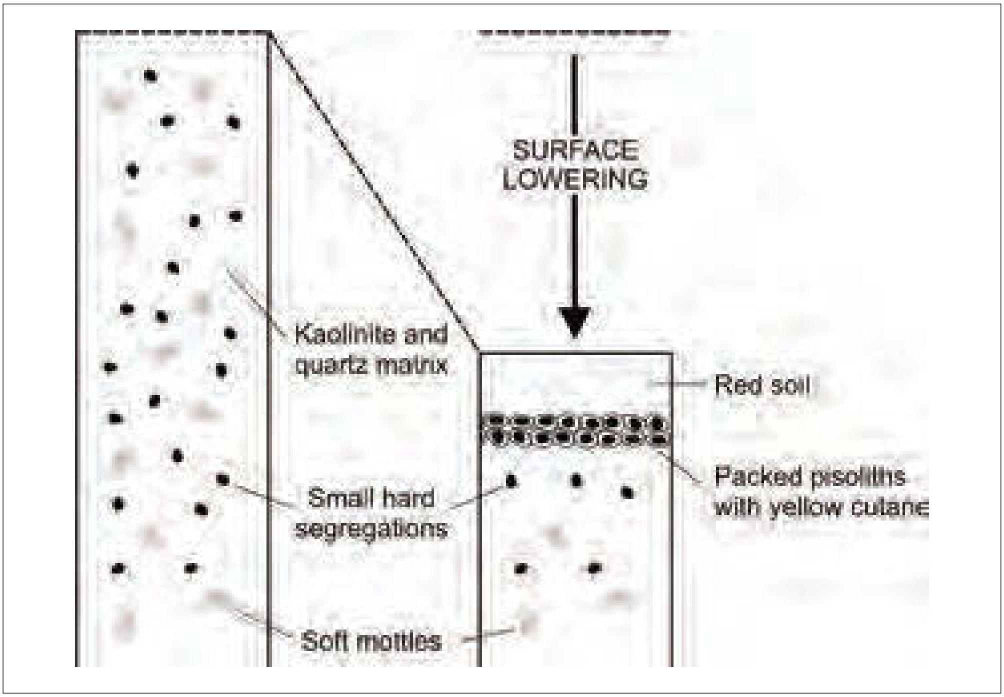

Figure 8.14. Accumulation of packed iron oxide pisoliths by the removal of quartz and kaolinite matrix as the ground surface is lowered as a result of tropical weathering. This process can lead to a continuous skeleton of ferricrete as iron is re-dissolved and precipitated between the pisoliths (after McFarlane, 1991b).

Where well-developed, laterite weathering profiles (first described in detail in Australia; Walther, 1915) display an uninterrupted series of zones, grading upwards from unaltered bedrock to indurated iron-rich duricrust, which can be distinguished in the field by changes in colour, texture and mineralogy (Figure 8.15). The nature of the various zones is described in detail by Aleva (1994) and Widdowson (2007). Unaltered bedrock at the base of the weathering profile is overlain by a saprock zone, consisting of a mixture of weathered and unweathered material. This passes upwards into a saprolite zone in which textures and structures inherited from the original bedrock may be recognised. A *mottled zone* with a microaggregated texture occurs above this, consisting of kaolinite particles and iron oxyhydroxide crystals. Upper sections of this zone may be characterised by nodular accumulations of segregated iron oxides (sometimes termed *pisoliths* or more accurately *pisoids*). Above this zone, at the top of the profile, is an indurated *vermiform* or *tubular* laterite duricrust (McFarlane, 1976). This comprises an interlocking reticulated fretwork of iron (and aluminium) oxides and hydroxides, formed by the progressive coalescence of segregated iron mottles during weathering and the concomitant removal of saprolite remnants from within and between these mottles. Pipes tend to be horizontally arranged towards the base of the vermiform duricrust, are anastomosing in the middle sections and predominantly vertically aligned in upper sections. The process by which an indurated crust develops is complex and involves dehydration and an increase in the degree of crystallinity of iron minerals such as goethite and haematite (Thomas, 1994). The general assumption is that repeated incremental increases of iron content over time, through the expansion and development of the mottled zone, lead to the development of an indurated layer. Nahon and Bocquier (1983) identify changes in Eh, either as a consequence of wetting and drying within the vadose zone or due to water table fluctuations, as a critical influence upon iron mobilisation and precipitation within the weathering profile.

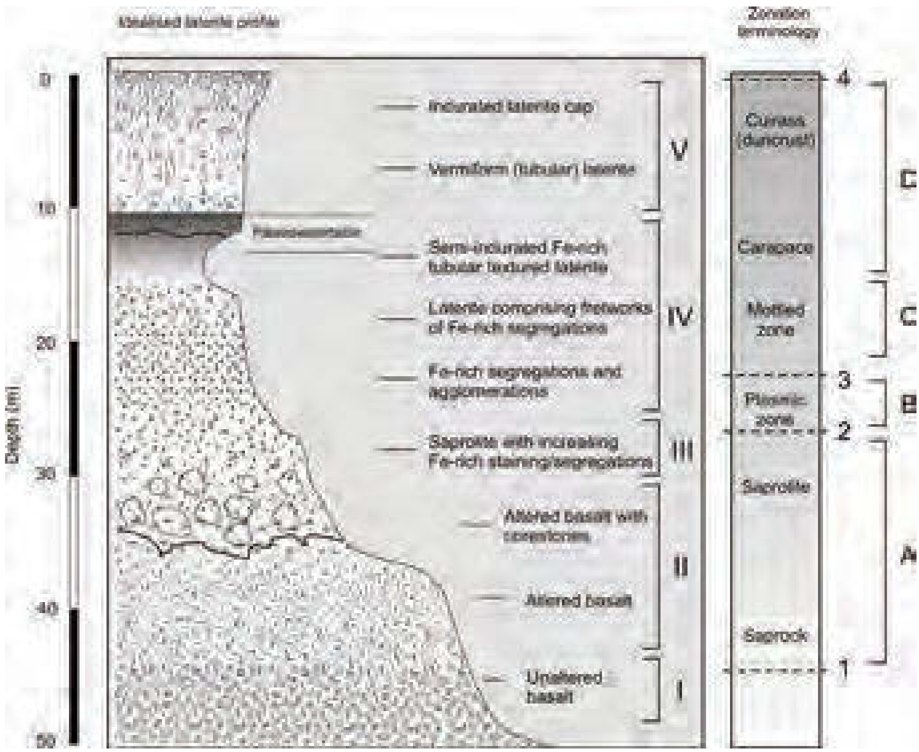

Figure 8.15. Generalised vertical section through an autochthonous laterite weathering profile, illustrating the compositional and textural progression from an unaltered (basalt) protolith to indurated laterite (after Aleva, 1994; Widdowson, 2007). A broad zonal distinction is as follows: i) unaltered basalt progressing up-profile to material displaying limited alteration of primary minerals; ii), altered basalt containing *unaltered* corestones within a soft saprolitic matrix, and progressing upward into iii), a saprolitic to weakly lateritised zone in which some primary lithological characteristics remain, but becoming increasingly obscured due to an up-profile increase of Fe-mottling and segregations; iv), moderately lateritised zone comprising iron segregations and vermiform textures, and becoming increasingly indurated up-profile; v), strongly indurated laterite with vermiform textures, becoming highly ferruginous in the uppermost levels of profile. A similar alteration progression is commonly observed on other protolith substrates. The right hand column shows the terminology of the various zones within a typical laterite weathering profile (A, Saprolitic zone; B, Plasmic zone; C, Mottled zone; D, Lateritic zone) and alteration fronts (1, Weathering front; 2, Pedoplasmation front; 3, Cementation front; 4, Surface erosional/weathering front).

Laterite weathering profiles can vary considerably in scale, ranging from a few centimetres to tens of metres in thickness (McFarlane, 1983). The precise nature of any profile is a response to climatic, geological and geomorphological conditions, with variability also introduced by rock structures and changes in the groundwater regime. A groundwater laterite may develop, for example, where iron segregations accumulate within the range of a fluctuating water table to form a closely packed pisolithic layer (McFarlane, 1983). With progressive land surface lowering and ongoing groundwater fluctuation, the packed pisoliths may be altered to form a massive vermiform layer (McFarlane, 1976). Ferricretes do not normally display a progressive weathering profile and can often be distinguished by their obvious discordance with the relatively unaltered underlying substrate (Ollier and Galloway, 1990). They do, however, often exhibit vermiform structures, formed by similar mechanisms to those in laterite profiles (Widdowson, 2007).

The mineralogy and chemistry of laterite and ferricrete is complex, with some 200 minerals having been identified in laterite alone (Aleva, 1994). The suite of minerals present is dependent upon the nature of the host rock or sediment and the weathering regime. The majority of iron rich duricrusts consist of a combination of haematite, goethite, kaolinite, gibbsite, boehmite, diaspore, corundum, anatase, rutile and quartz (McFarlane, 1983). In laterites, the iron and aluminium oxyhydroxides may form either directly from the alteration of primary minerals or via a series of pathways (Figure 8.16) involving the formation of intermediate sheet silicates and clays which are themselves broken down and converted to iron and aluminium oxyhydroxide residua (Anand, 2005; Widdowson, 2007). The type and nature of minerals may vary throughout a profile. Goethite, for example, may exist in an amorphous state within the top of the saprolite zone and in a crystalline form in upper parts of the profile. Similarly, kaolinite often becomes more crystalline towards the top of a profile. Goethite is typically the most common iron mineral within the saprolite and mottled zones, whereas haematite becomes dominant within the duricrust layer (Thomas, 1994). Ferricretes rarely display the progression of alteration minerals observed in laterites and may retain primary minerals from the host rock in addition to secondary iron and aluminium oxyhydroxides precipitated from external sources.

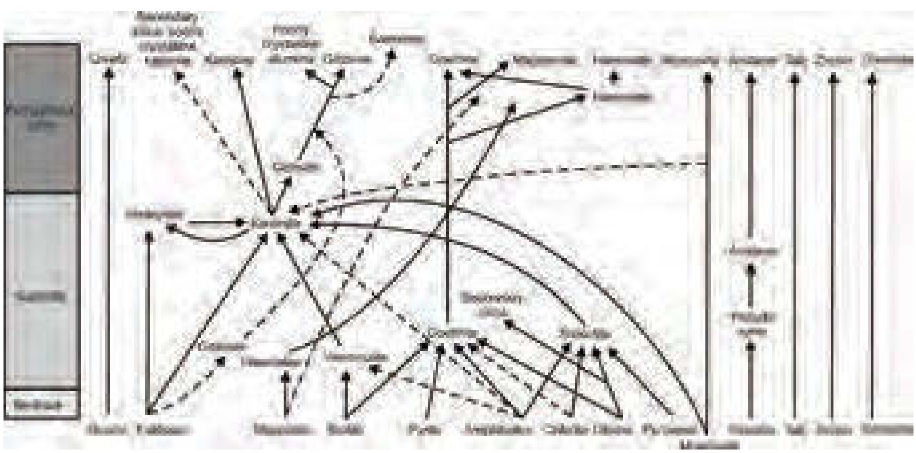

Figure 8.16. Pathways of formation of secondary minerals in lateritic weathering profiles (after Anand, 2005).

Both laterite and ferricrete are physically and chemically durable, and have played an important role in the evolution of tropical and sub-tropical landscapes, including those of southern Africa. The precise distribution of laterite and ferricrete in the subcontinent is difficult to determine since current geological and soil survey maps do not distinguish between the two materials in the manner used here (see Netterberg, 1985; Botha, 2000; Fey, 2010). Iron-rich duricrusts are found in parts of the southern Democratic Republic of Congo, Angola, Zambia, Zimbabwe, eastern Botswana, Mozambique, South Africa and Madagascar (e.g. King, 1951; Maud, 1965; Goudie, 1973; Helgren and Butzer, 1977; Alexandre and Alexandre-Pyre, 1987; Partridge and Maud, 1987; McFarlane, 1989; Marker and McFarlane, 1997; Marker and Holmes, 1999; Marker *et al.*, 2002). In South Africa, laterite is widespread in the north part of the Eastern Cape Province of South Africa, the eastern coastal hinterland of KwaZulu-Natal Province of South Africa and parts of the Highveld in former Transvaal. In Botswana, iron-rich duricrusts are relatively common at the elevated periphery of the Kalahari Sandveld (e.g. around Serowe, Kanye and Molepolole; Marty McFarlane, pers. comm.). Iron-rich crusts have also been described overlying silcrete and calcrete in the flanks of fossil valleys (e.g. Nash *et al.*, 1994a; Figure 8.13b). These are almost certainly ferricretes, since neither the calcrete nor silcrete has the potential to have acted as a source of iron.

Laterites in southern Africa are best documented where they occur in association with the African Surface Complex. Indeed, a close genetic link between the processes responsible for palaeosurface formation and lateritic weathering profile development has been identified by a number of authors (see McFarlane, 1991a; Marker and Holmes, 1999; Marker *et al.*, 2002; Widdowson, 2007). Both palaeosurfaces and lateritic weathering profiles require long periods of exposure to sub-aerial weathering, relative tectonic stability and a reasonably stable climate in order to develop (Widdowson, 1997), with rapid uplift or climatic shifts potentially terminating the lateritisation process. However, laterite development may be associated with the slow incision of low relief palaeosurfaces (Thomas, 1994) and can be polycyclic if punctuated by episodes of uplift or relatively short-lived shifts in climatic conditions. Most of the laterite profiles in southern Africa are likely to be pedogenetic in origin, the subcontinent lying well beyond the low latitude groundwater laterite belt. An exception is an immature groundwater laterite profile associated with the African Surface Complex in the Albertinia area of South Africa (described by Marker *et al.*, 2002). This profile shows evidence of polycyclic development, suggesting that there was tectonic instability during its formation, consistent with the identification of three altitudinally separate components to the African Surface in this area (Marker and McFarlane, 1997). The documentation of groundwater laterites only around the continental margin of the subcontinent implies that moisture was a limiting factor in their formation (Marty McFarlane, pers. comm.).

6. Duricrusts as palaeoenvironmental indicators

Duricrusts can be employed as palaeoenvironmental indicators if (a) their mode of origin, (b) age, and (c) the contemporary environmental factors that control their formation and distribution are known. The mode of origin of the majority of duricrusts found within southern Africa can be readily identified through careful analysis of their macro- and micromorphology and comparison with equivalent crusts found elsewhere (see Nash and McLaren, 2007b, and chapters therein). This level of analysis is essential if the types of palaeoenvironmental misinterpretation noted above for silcretes in the Cape Coastal Zone are to be avoided in the future.

Determining the precise age of any duricrust is problematic unless there is tight stratigraphic or radiometric dating control. Stratigraphical techniques provide a relatively poor age constraint since the host sediment or bedrock may be considerably older than any duricrust profile developed upon or within it (Netterberg, 1978). This has proven to be a particular problem when attempting to determine the relative age of Kalahari duricrusts (Haddon and McCarthy, 2005). The incorporation of fossils or rare stone artefacts within a duricrust may be useful as an indication of its maximum age (e.g. Netterberg, 1978). It is possible to date carbonate-rich crusts using the radiocarbon and uranium series techniques (see Netterberg, 1969c; 1978; Partridge *et al.*, 1984; Shaw *et al.*, 1992; Shaw and Thomas, 1993 for examples). However, great care is needed to ensure that only one generation of carbonate is represented in the dated sample (Netterberg, 1978; Ku *et al.*, 1979; Rust *et al.* 1984), otherwise any age estimate derived from a bulk sample will be flawed. This problem can be overcome through microsampling of carbonate phases (e.g. Candy *et al.* 2004; 2005). It is also important when using radiocarbon dating that the appropriate correction factors are applied to take into account natural variations in stable isotope levels (Salomons and Mook, 1976; Salomons *et al.*, 1978). Silcretes have yet to be dated successfully, although the development of techniques for dating diagenetic events (McNaughton *et al.*, 1999) and K-Mn oxides in weathered profiles (Vasconcelos, 1999; Vasconcelos and Conroy, 2003) offers considerable potential. Radtke and Brückner (1991) used electron spin resonance dating to estimate the ages of Australian silcretes; however, their use of bulk samples with multiple cement phases rendered the derived dates almost meaningless. Laterite weathering profiles can be dated using palaeomagnetic techniques, uranium-series and other dating methods such as K/Ar and $^{40}Ar/^{39}Ar$ (see Vasconcelos, 1999; and Widdowson, 2007 for a review). Care is needed when using isotopic dating techniques in the event that the weathering profile contains inherited muscovite since this can generate anomalously old ages.

Some generalisations can be made about the environmental factors that control duricrust formation, although these should be treated with considerable caution. Low relative relief appears to be a requirement for the development of all duricrust types other than pedogenic gypcrete. The topographic setting may also influence the ability of a host sediment or regolith to retain moisture or permit leaching. Most contemporary gypsum crusts are forming in arid areas (Watson, 1983a; 1985). Pedogenic calcretes are known to be forming today in dryland soils in regions with rainfall in the range 200-600 mm (Goudie, 1973). *Fossil* pedogenic gypcretes and calcretes may, therefore, be a useful indicator of arid and semi-arid climates respectively in the past. Analyses of stable carbon and oxygen isotopic within pedogenic calcretes can be used in the reconstruction of palaeoclimates, past vegetation and CO_2 concentrations (see Cerling, 1999; Alonso-Zarza, 2003). The oxygen isotope composition of a calcrete is directly related to that of the meteoric water from which it formed. Values of $\delta^{18}O$ lower than –5% do not occur in arid zones, and areas receiving less than 350 mm rainfall per annum have $\delta^{18}O$ values greater than –2% (Talma and Netterberg, 1983). The $\delta^{13}C$ values of soil carbonates at depths below 30 cm depend on the isotopic composition of the soil CO_2. Low $\delta^{13}C$ values are normally taken to indicate the dominance of C_3 plants whilst heavier values suggest greater proportions of C_4 and CAM plant communities (Cerling, 1999).

The environment of silcrete formation is more difficult to determine. Silcretes are only forming in a limited number of locations today (Flach *et al.*, 1969; Ellis and Schloms, 1982; Chadwick *et al.*, 1989; Shaw *et al.*, 1990; Dubroeucq and Thiry, 1994) making it difficult to identify representative modern analogues. Pedogenic silcrete is thought to form under climates with alternating humid and dry seasons or periods, although the pH range under which silicification occurs remains disputed. Palaeotopography is considered to be important for the formation of groundwater, drainage-line and pan/lacustrine silcrete, since topography controls both the water table position and groundwater flow (see Nash and Ullyott, 2007). Climate may, however, also be important; with Laser Raman and Fourier Transform Infrared vibrational spectroscopic (Raman/FT-IR) analyses of silica species within non-pedogenic Kalahari silcretes indicating that formation took place under broadly semi-arid conditions (Nash and Hopkinson, 2004). The precise climatic conditions needed for lateritisation remain uncertain (e.g. Bardossy and Aleva, 1990; Tardy *et al.*, 1991; Taylor *et al.*, 1992). However, there is broad agreement that consistently warm temperatures and high amounts of seasonal precipitation are required, making laterite a potentially useful indicator of humid tropical climates (e.g. McFarlane, 1983; Widdowson, 2007). The application of isotope geoscience to the study of laterite has provided a number of environmental tracers. For example, strontium, lithium, samarium and neodymium isotope systems have been used to identify surface aeolian inputs to lateritic profiles in India and Cameroon (e.g. Rudnick *et al.*, 2004; Viers and Wasserberg, 2004).

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9

Periglacial and Glacial Geomorphology

Periglacial and Glacial Geomorphology

Stefan W Grab, Stephanie C Mills and Simon J Carr

1. Introduction

The geomorphology of cold regions receives wide international attention from researchers, particularly in the context of past and future global environmental change (Haeberli and Beniston, 1998; IPCC, 2007). The term *periglacia*l reflects a complex set of environmental interactions, but is used here to refer to a "wide range of cold, non-glacial conditions, regardless of their proximity to a glacier, either in time or space" (French, 1996:3), and is generally defined as "the sub-discipline of geomorphology concerned with cold non-glacial landforms" (French, 2007:5). The effectiveness of periglacial processes and landform development is controlled by many environmental factors including the nature of earth materials, available ground moisture and the intensity, duration and cyclic pattern of ground (soil or rock) temperatures passing below 0 °C. Areas with less severe frost action, such as in the mountains of southern Africa, are sometimes referred to as *marginal periglacial* or *subperiglacial* zones (e.g. Lewis, 1988a; Hanvey and Marker, 1992; Hall, 1992; Boelhouwers, 1994). Such areas typically experience diurnal or seasonal ground freeze. However, high latitude and high alpine regions where mean annual temperatures are usually below 0 °C may experience more severe frost action associated with permanently frozen ground, known as permafrost, which is ground that remains at or below 0 °C for at least two years (Harris *et al.*, 1988).

The landscapes reflecting present-day glacierisation and past glaciation are probably the most readily visible element of cold climate environments at a global scale. Glaciers and ice sheets currently cover around 10% of the Earth's surface (~16 million km^2), but have covered over 30% during recent geological time. Furthermore, the influence of glaciers and ice sheets extends well beyond mapped ice limits through the supply of melt-water and debris to glacier margins, with global impacts on ocean circulation and the climate system (Benn and Evans, 2010). South of the volcanoes of east Africa, there is no present-day glacierisation of Africa, but some researchers have proposed the development of mountain glaciers during the Quaternary ice-ages on the highest areas of the Drakensberg Escarpment and associated upland regions. Further back in geological time, there is evidence of considerable late Paleozoic glaciation over southern Africa, at a time when the region was located close to the South Pole, and part of the Gondwana supercontinent (Visser, 1990).

This chapter considers periglacial and glacial phenomena, as well as their associated geomorphic processes and environmental implications, reported from southern Africa. Whilst acknowledging the substantial periglacial work undertaken on Marion Island, the sub-Antarctic islands constitute a "distinct periglacial environment" in contrast to those in southern Africa or other alpine environments (Boelhouwers *et al.*, 2003), and geographically fall outside the scope of this chapter.

2. Periglacial environments in southern Africa

Periglacial phenomena and processes have been described from three southern African mountain regions, namely the Western Cape Province (Hex River and surrounding ranges), Eastern Cape Province (Amatola, Witteberge and southern Drakensberg) and Lesotho/KwaZulu-Natal Province/northeastern Free State Province (Drakensberg) (Figure 9.1). Little is known about similar features in other high mountain environments such as the Cedarberg in the Western Cape, Karoo mountain ranges, the Winterberge in the Eastern Cape, and central Lesotho mountains. In the first instance, it is important to determine what climatic (air and ground) regimes are most conducive to establishing cryogenic landscape imprints in those areas where they have been observed.

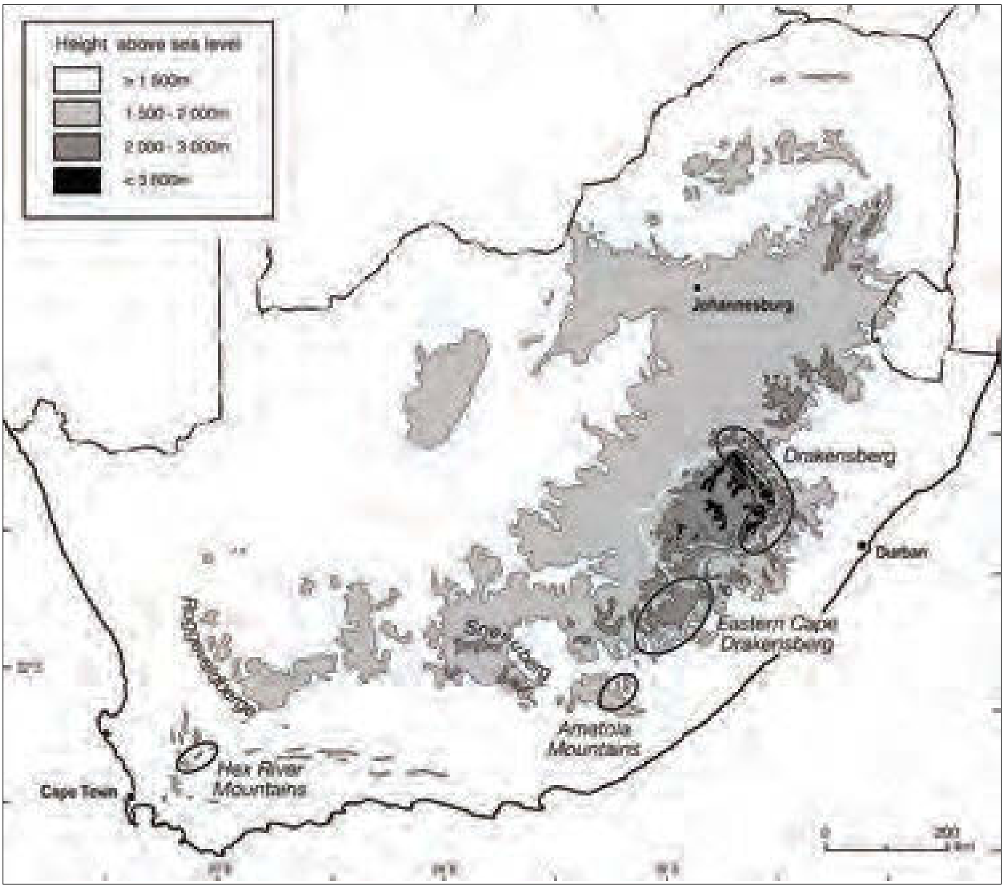

Figure 9.1. Mountain and upland regions of southern Africa and core areas of periglacial and glacial geomorphological research.

The Hex River Mountains in the Western Cape have summits between 1 600-2 000 m a.s.l., attaining a maximum altitude at Matroosberg (2 249 m a.s.l.). These mountains represent a series of folded quartzitic sandstones (Peninsula Formation), with occasional narrow argillaceous bands (Cedarberg Formation shale) (Boelhouwers, 1995a; 1998). The mountain summits have a mean annual air temperature of between 7 and 10 °C, with frost occurring from May to September (Boelhouwers, 1998; 1999a). Air temperature records suggest approximately 62 frost days (<0 °C) p.a., with absolute minimum temperatures of –6 to –8 °C occurring 2-3 times per year on the higher summits (Boelhouwers, 1998). The Western Cape mountain climate is distinctly different to that of the Eastern Cape and Lesotho-Kwazulu-Natal region, as it receives approximately 77% of precipitation during the frost season from May to September. Consequently, soil moisture is readily available in winter, and conservative estimates suggest that mountain summits receive an average of 31.4 snow days p.a. (range = 15 to 68 days p.a. between 1971 and 1989), which has important implications for ground insulation, cooling effects and moisture regime changes (Boelhouwers, 1991a; Nel *et al.*, 2009). Whilst snow may only remain on the ground for a couple of days during early winter (May/June), it sometimes lasts for up to three weeks in mid to late winter (July-September) (Boelhouwers, 1998).

Ground temperatures vary considerably, both spatially and temporally, due to inter-annual and microclimate contrasts and associated snow depth and longevity controls. In addition, underlying lithology has important implications for frost susceptibility and the consequent development of periglacial landforms (Murton and Lautridou, 2003; French and Thorn, 2006; André, 2009). For example, Waaihoek Peak (1 900 m), underlain by quartzitic sandstone, is only subject to diurnal freeze-thaw cycles and a maximum frost penetration of approximately 10 cm. In 1993/1994, frost penetrated to –1 cm on 56.5 days p.a., whilst at –20 cm it only occurred on 1.5 days per annum. In contrast, the Cederberg Formation shale at Mount Superior (1 860 m) recorded 46.5-frost days per annum at –1 cm (1993/1994), but negligible frost penetration beyond this depth, and no records of frost at –5 cm. Despite these trends, Waaihoek Peak lacks cryogenic phenomena, whilst Mount Superior hosts active miniature sorted patterned ground on the summit, and stone- and turf-banked lobes and steps on its north- and south-facing slopes (Boelhouwers, 1995a; 1998). Approximately 90% of the Western Cape mountains are underlain by coarse textured quartzitic sandstone with 4-5% fines (clay/silt), as found at Waaihoek Peak. According to Meentemeyer and Zippin (1981), an optimum efficiency composition of 12-19% of fines characterise frost- susceptible soils, conducive to producing cryogenic landforms. The Cederberg Formation shale at Mount Superior has 24% fines and is frost-susceptible under saturated conditions (Boelhouwers, 1998). The climate and lithological conditions of the Western Cape region thus largely restrict contemporary ground freeze, with geomorphically effective diurnal freeze-thaw being spatially limited to a few summits containing Cederberg Formation shale.

The mountains of the Eastern Cape vary from approximately 2 000 m a.s.l. in the Amatola Mountains to approximately 3 000 m in the southern Drakensberg (Figure 9.1), with active and relict cryogenic phenomena reported between about 1 800 and 2 900 m a.s.l. The geology primarily comprises Beaufort and Stormberg Group sandstones (Molteno, Elliot and Clarens Formations) at lower altitudes, and Karoo Dolerites and flood basalts at higher altitudes. Ambient and ground climate data from approximately 2 800 m a.s.l. at Tiffindell Ski Resort (see Kück and Lewis, 2002) suggest mean annual temperatures of 7.5 °C, with frost occurring on 40% of days between May and September. The data indicate an average of 63 frost days per annum, and an absolute minimum temperature of –13 °C; hence not dissimilar to conditions measured in the Hex River Mountains, 1 000 m lower in altitude.

Inter-annual ground frost penetration in the Eastern Cape mountains may vary considerably, depending on snow abundance and duration (Kück and Lewis, 2002). Although near surface (–5 cm) freeze-thaw days were similar in 1995 (45 days) and 1996 (34 days), days recording maximum temperatures ≤0 °C varied from 10 days in 1995 to 78 days in 1996. Ground freezing at –20 cm was recorded on 61 days in 1995 and on 114 days in 1996. Thus, contemporary frost penetration on high altitude (2 800 m a.s.l.) south-facing slopes in the Eastern Cape extends beyond –20 cm and remains (seasonally) frozen for several weeks to months at such a depth.

The Lesotho/KwaZulu-Natal Drakensberg consists of mountains and plateaus averaging 2 900-3 000 m a.s.l. for over 280 km; reaching 3 482 m a.s.l. at Thabana Ntlenyana (Figure 9.1). Although the geology is similar to that of the Eastern Cape, most reported cryogenic phenomena and process studies stem from the basaltic summit regions. Estimates of mean annual air temperatures for the high plateaus and summits vary from about 5.8 °C for the Sani Valley at 2 900 m a.s.l. (Nel and Sumner, 2008) to about 4.0 °C for the highest Lesotho summits over 3 400 m a.s.l. (Grab, 1999a). Mean seasonal temperatures vary from approximately 10 °C in summer (December-February) to 0 °C in winter (June-August) (Grab, 2005a), with ground level frost estimated on about 180 days per annum between April and October. Absolute annual minimum winter temperatures typically range between approximately –11 °C and –20 °C, depending on altitude and topographic setting (see Grab, 1997a; Nel and Sumner, 2008). In contrast to the Western Cape, 70% of precipitation over the Drakensberg falls between November and March, and less than 10% between May and August (Tyson *et al.*, 1976), which has important implications for nocturnal cooling, potential snow cover, and soil moisture availability. Estimates of total annual precipitation across the high Drakensberg vary from as much as 1 600 mm (Sene *et al.*, 1998) to less than 800 mm (Nel and Sumner, 2008). Although about eight annual snowfalls occur on the high Drakensberg, primarily between April and October (Tyson *et al.*, 1976; Mulder and Grab, 2009), their water equivalent contribution may be less than 100 mm (Nel and Sumner, 2005).

A primary outcome of numerous studies monitoring ground thermal characteristics in the high Drakensberg is that the intensity, duration and depth of freeze across the landscape is extremely diverse owing to the equally diverse settings in which such studies have been undertaken. Although factors such as percentage and type of vegetation cover, sediment characteristics, soil moisture, snow cover depth/duration, macro- and microtopography, all influence ground thermal regimes and cryogenic processes, these have received little research attention in southern African periglacial studies.

3. Active (contemporary) periglacial phenomena

Table 9.1 presents a list of periglacial landforms identified in the three core cryogenic regions of southern Africa. It is important to note that some of these landforms may not originate from an exclusively periglacial origin and are likely polygenetic, while it is not clear whether some features are *active* or *relict*. *Active* or *contemporary* periglacial phenomena are those landforms considered to be developing or being maintained under current environmental conditions.

Table 9.1. Active (contemporary) and inactive (relict) periglacial and glacial phenomena and landforms reported from the core cryogenic mountain regions of southern Africa. Please note that the identification of several of the inactive landforms is now considered incorrect, whilst others remain contentious.

CRYOGENIC PHENOMENA AND LANDFORMS	CORE CRYOGENIC MOUNTAIN REGIONS IN SOUTHERN AFRICA		
Active (contemporary)	**W. Cape**	**E. Cape**	**Lesotho/KZN | E. Free State**
Earth hummocks (thufur)		x	x
Flarks			x
Freeze/thaw weathering*	x	x	x
Frost action in soil	x	x	x
Miniature sorted patterned ground	x	x	x

CRYOGENIC PHENOMENA AND LANDFORMS	CORE CRYOGENIC MOUNTAIN REGIONS IN SOUTHERN AFRICA		
Miniature non-sorted patterned ground			x
Terraces and terracettes	x	x	x
Needle-ice ('pipkrake')	x	x	x
Needle-ice pans and cryo-deflation			x
Needle-ice: sediment movement and associated micro-landforms	x	x	x
Ploughing boulders			x
Segregation ice and 'ice lenses'		x	x
Solifluction (turf/step banked) lobes	x	x	x
Stone-banked lobes	x		x
Turf exfoliation		x	x
Turf exfoliation pans			x
Inactive (relict)			
Cryoplanation	x		x
Cryoturbation & 'periglacial' soils		x	x
Glaciers	x	x	x
Glacial erosional phenomena	x	x	x
Glacial depositional phenomena	x	x	x
Ice-wedge casts		x	
Nivation (processes & phenomena)	x	x	x
Openwork block deposits	x	x	x
Permafrost		x	x
Pronival ramparts		x	x
Rock glaciers		x	
Scree (cryoclastic)	x	x	x
Large sorted patterned ground	x	x	x

CRYOGENIC PHENOMENA AND LANDFORMS	CORE CRYOGENIC MOUNTAIN REGIONS IN SOUTHERN AFRICA		
Sedimentary 'periglacial' deposits		x	x
Solifluction (processes & phenomena)	x	x	x
Stone-banked lobes (garlands)	x	x	x
Valley asymmetry			x

3.1 Needle-ice and associated landforms

The occurrence of needle-ice has been described from all three-core cryogenic mountain regions of southern Africa and entails the "accumulation of slender, bristle-like crystals (needles) at or immediately beneath, the surface of the ground" (Washburn, 1979:92). Several environmental variables determine the potential for needle-ice development, including combinations of soil and microclimatic variables such as soil physical properties, freezing duration and cooling through the near-surface soil profile (Meentemeyer and Zippin, 1981; Lawler, 1988; Boelhouwers, 1998; Grab, 2001). Field experimental work in the high Drakensberg has previously monitored needle-ice growth and ablation cycles against air and soil temperatures (Grab, 2001), from which it was shown that previously defined environmental thresholds (Outcalt, 1971, Lawler, 1993) do not always apply to conditions recorded locally. Needle-ice lengths typically exceed 31 mm for individual nocturnal growth phases in the high Drakensberg, but their net lengths are dependent on freeze duration, wind characteristics and cloud cover (Grab, 2001). Polycyclic or multilayered needle-ice (Figure 9.2) with lengths exceeding 400 mm may develop during particularly cold periods when growth phases extend over several days and ablation is limited (Grab, 2002a). In the core southern African cryogenic environments, sediment heave and mobilisation associated with needle-ice are considered important contributing processes to stream bank erosion (Grab, 1999b; Grab and Deschamps, 2004) and in the development of landforms such as needle-ice mounds (Grab, 2002a), sorted patterned ground (Hanvey and Marker, 1992; Grab, 1997b), terracettes and turf-banked terraces (Boelhouwers, 1991b; Hanvey and Marker, 1992; Kück and Lewis, 2002), and those associated with turf exfoliation (Boelhouwers, 1991b; Grab, 2002b; 2010).

Figure 9.2. Polycyclic (multi-layered) needle-ice with heaved sediment cap.

Needle-ice induced particle movement has been measured along gravel/cobble dominated stream banks and gullies containing peat. Experimental designs include the use of painted cobbles and clasts positioned along linear transects and the placing of collection troughs to capture displaced materials (see Grab, 1999b; Grab and Deschamps, 2004). Although rain wash and stream discharge are the dominant agents for sediment mobilisation during the wetter months, the studies have demonstrated that needle-ice induced sediment mobilisation during winter averages between 0.4 mm and 2.8 mm/ day/particle, and is also an important indirect erosion agent by disaggregating earth materials for subsequent fluvial entrainment.

Soil disruption by needle-ice, desiccation and aeolian deflation are considered primary causes for *turf exfoliation* – "a denudation process active in periglacial areas which destroys a continuous ground vegetation cover by removing the soil exposed along small terrace fronts" (Pérez, 1992:82). Turf exfoliation is widespread through the Lesotho mountains and a primary agent in the development of needle-ice pans (Grab, 2002a) and alpine turf exfoliation pans (Grab, 2010, Figure 9.3). Particularly during autumn and spring, when there is sufficient soil moisture along terrace risers located on slopes or around pans and other microtopographic hollows/niches, needle-ice lifts soil material along the exposed risers; which subsequently is displaced through needle ice collapse/ablation and gravity effects. Soil material consequently accumulates at the base of risers where it is removed through surface wash and deflation. Although it is estimated that approximately 200 needle-ice events are possible per annum in the high Drakensberg where and when environmental controlling factors are favourable (Grab, 1999b), temporally and spatially, the extent and frequency of needle-ice activity may differ considerably. For instance, thermal conditions conducive to needle-ice development along various pan riser aspects in 2005 varied from 71 days on a south-facing aspect to 95 days on an east-facing aspect (Grab, 2010). The strong seasonality ranging from mild, wet summers to cold, dry winters has induced a cycle of dominating processes (fluvial and biological activity, aeolian deflation and deposition, frost action), which operate synergistically to promote turf exfoliation, and associated microlandforms (Figure 9.4) (Grab, 2002b; 2010).

Figure 9.3. An alpine pan in the high Drakensberg (dry season) with turf-exfoliated risers and micro-echo dune development along the base of some risers.

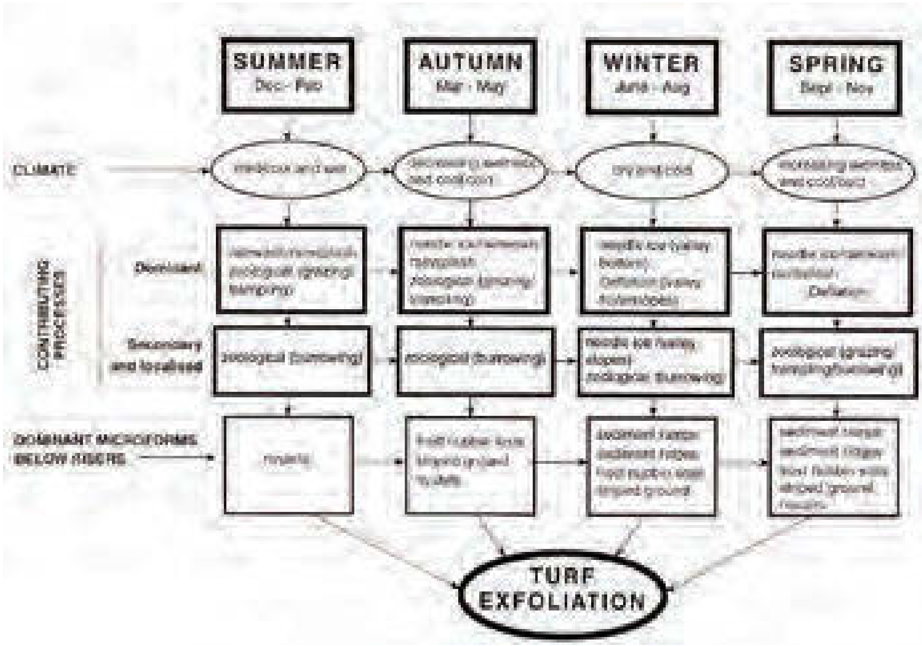

Figure 9.4. A model depicting the cyclic pattern of processes and associated microforms contributing to turf exfoliation in the high Drakensberg (after Grab, 2002b).

In recent years, considerable work has examined the interaction between needle ice processes and ecosystem functioning in the sub-Antarctic (e.g. Haussmann *et al.*, 2009a, b), which is a remaining research gap in southern Africa. Given its widespread erosional role in southern African mountain environments, and particularly so in alpine wetland ecosystems, the significance of needle-ice as a geomorphic agent cannot be overlooked in future studies.

3.2 Miniature sorted patterned ground

Patterned ground is a collective term for the "more or less symmetrical forms, such as circles, polygons, nets, steps and stripes, that are characteristic of, but not necessarily confined to, mantle subject to intensive frost action" (Washburn, 1956:824). Although sorted patterned ground in many environments does not necessarily have a cryogenic origin (see Hallet, 1990), that found in periglacial regions has been classified into sorted and non-sorted varieties. Whilst *sorted patterns* "are made prominent because of the segregation of stones and fines" (Gleason *et al.*, 1986:216), *non-sorted patterns* are characterised by "ground cover or colour variations" (Krantz, 1990:117). Miniature patterns are identified from large patterns as having a mean mesh diameter or stripe width of <20 cm (see Wilson and Clark, 1991).

Miniature sorted patterned ground has been reported from all three core southern African cryogenic regions, particularly on poorly drained surfaces with low vegetation cover, which consequently promotes the upfreezing of clasts (see Boelhouwers, 1991a; 1995a; Hanvey and Marker, 1992; Lewis 1996; 2008a; Grab, 1997b). The extent of lateral and vertical sorting of sediments into coarser and finer grained spatial units has been determined through the establishment of sorting indexes (see Ballantyne and Matthews, 1983; Grab, 1996a; 1997b). Finer varieties of loam-dominated stripes, known as *raked* patterns (Hanvey and Marker, 1992) or *wind-striped, frost-heaved soil* (Troll, 1958) are thought to originate through various environmental factors including early morning solar radiation, needle ice and wind (Troll, 1958; Hastenrath and Wilkinson, 1973). Coarser cobble-sorted stripes, approximately 13 cm wide and sorted to a depth of 7 cm, occur on Mafadi Summit (Drakensberg), where they appear to be degrading under current environmental conditions (Grab, 1996a). At Mount Superior in the Hex River Mountains, coarse stripes are typically 5-6 cm wide and sorted to a depth of 6 cm, extending several meters down slopes (Boelhouwers, 1995a). Clasts occupying coarse stripes have a pronounced vertical tilt and a long axis alignment parallel to the stripes (Figure 9.5). In the high Drakensberg, miniature sorted polygons and circles re-develop annually in dry streambeds where they consist of fine cone-shaped centres enclosed by either blocky clasts and boulders or coarse gravel and small pebbles (Grab, 1997b). The physics of small-scale patterned ground formation is still not fully understood, but it has been suggested that those in the Western Cape and Lesotho are a product of freeze-thaw processes (Boelhouwers, 1995a), differential swelling, and frost heaving (Grab, 1997b) under diurnal frost cycles (see also Van Vliet-Lanoë, 1991; Matsuoka *et al.*, 2003).

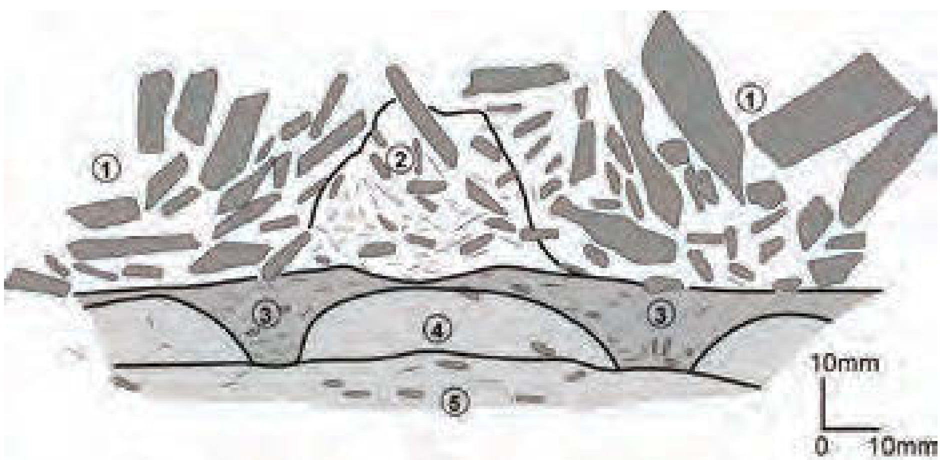

Figure 9.5. Cross section through a sorted stripe indicating sediment structures and clast orientations as found near Mt Superior at 1 850 m, Hex River Mountains. 1) Coarse stripe, clasts 15-40 mm, 2) Fine stripe, clasts 2-10 mm, coarsening upward and sideward, 3) 70-90% clasts 0.5-4 mm in matrix, coarsening down into *wedge*, 4) 10-30% clasts 0.5-4 mm in d. brown loam, 5) 30-60% clasts 0.5-4 mm d. brown loam (after Boelhouwers, 1995).

3.3 Earth hummocks (thufur)

Earth hummocks, also referred to as *thufur*, are miniature frost-induced mounds characterised as a type of *non-sorted net* common to periglacial environments (Washburn, 1956). Conditions most conducive to the development of earth hummocks include relatively deep (ca. >50 cm) frost susceptible soil, a preferential cover of lawn grasses/sedges, adequate soil moisture and at least seasonal ground freeze to several cm depth (see Grab, 2005b). Collectively, it would appear that such conditions are only met in the Eastern Cape Drakensberg (e.g. Lewis, 1996; 2008a; Kück and Lewis, 2002) and the Lesotho/KwaZulu-Natal Drakensberg (e.g. Boelhouwers, 1991b; Hanvey and Marker, 1992; Grab, 1998; 2005a,b) above approximately 2,650 m a.s.l., associated with wetland sites or ground seepage zones (Figure 9.6). The several hundred dome-shaped cryogenic hummocks measured in the Drakensberg have an average height of 20 cm and mean diameter of 49 cm (Grab, 1998), but many are elongated down the dominant slope, whilst other hummocks have complex morphologies due to the coalescence of adjoining hummocks (Grab, 2005a). The hummocks have a relatively high silt/clay (ca. 28%) and organic matter content (ca. 31%) compared to the adjoining inter-hummock areas where values are approximately 9% and 17% respectively (Grab, 1997c).

Figure 9.6. Earth hummocks which have developed around a thermal spring in Lesotho.

Field experiments in Lesotho have included the insertion of wooden pegs into hummocks to evaluate contemporary heave (Grab, 1998), and temperature probes to determine thermal characteristics through hummocks and adjoining depressions (Grab, 1997c; 2005a). These studies suggest that contemporary heave still occurs in southern African earth hummocks and that the freezing process is aspect controlled, commencing on the hummocks' southern aspects and gradually progressing towards the western and northern aspects, whilst the eastern aspects remain unfrozen throughout winter, but are subjected to near-surface diurnal freeze-thaw cycles (Figure 9.7) (Grab, 2005a). Given that some earth hummocks contain large clasts, it is possible that the *cryoexpulsion* (sub-surface clast movement towards the surface through frost push and pull mechanisms) of clasts may contribute to such hummock development (See Dionne, 1966; Grab, 2005b). More widely documented hypotheses for earth hummock development include hydrostatic, cryostatic and density-induced pressures (e.g. Hallet and Waddington, 1992), cellular circulation (equilibrium model) (e.g. Mackay, 1980; Krantz *et al.*, 1988) and differential frost heave (cryoturbation) (e.g. Van Vliet-Lanoë *et al.*, 1993; 1998). Based on the thermal data obtained from the Lesotho earth hummocks, differential frost heave is the likely dominant local mechanism operating in pre-existing hummocks (see Grab, 2005a, b).

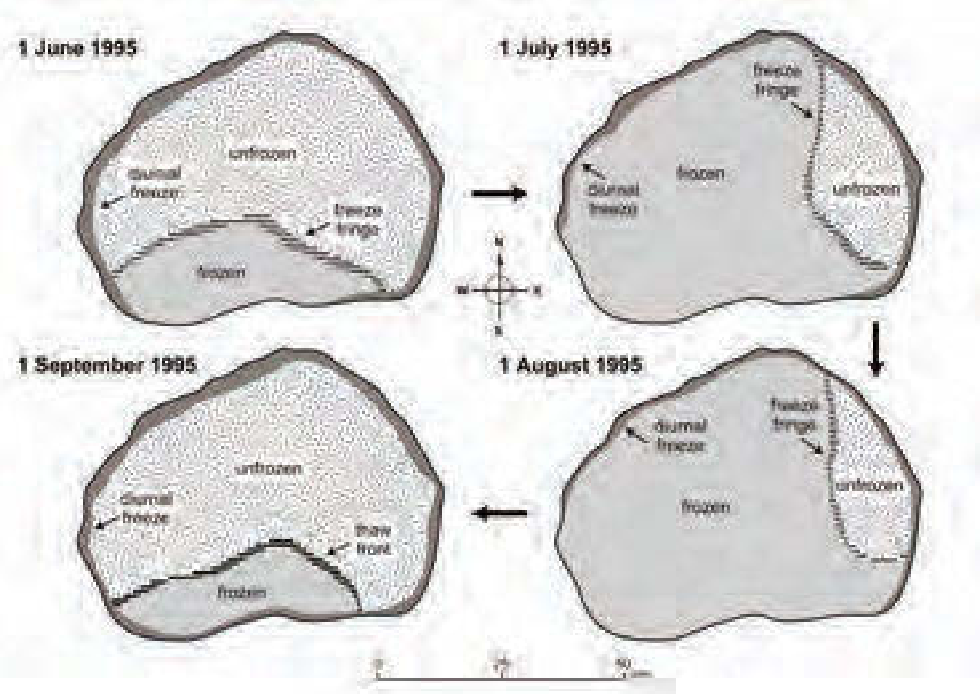

Figure 9.7. Sketched plan view of aspect-controlled freezing in the mid-portion of an earth hummock (1995 – snow-free conditions) (after Grab, 2005a).

Recent work on earth hummocks and related landforms (e.g. frost boils) have monitored soil moisture and frost heave at microsites using instruments such as reflectometry sensors and ultrasonic distance sensors attached to data loggers (e.g. Overduin and Kane, 2006; Scott *et al.*, 2008). Future southern African work could similarly include microspatio-temporal measures of soil moisture and heave associated with thermal monitoring, and perhaps more particularly attempt to establish sub-surface moisture and sediment movement patterns. Given that earth hummocks have an important impact on regulating and controlling hillslope drainage (Quinton and Marsh, 1998) and fine-scale ecology (Scott, 2006), there is considerable scope for ongoing research in southern Africa.

3.4 Solifluction landforms (stone-banked lobes, turf-banked/stepped phenomena)

The term *solifluction* was first defined as the "slow flowing from higher to lower ground of masses of waste saturated with water" (Andersson, 1906:95), and in periglacial environments includes mechanisms of frost heave and frost creep. There is a wide range of solifluction landforms identified in southern Africa. Stepped microrelief in the form of turf-banked steps and stone-banked lobes, varying in mean height between 17-93 cm and in mean lengths of 102-274 cm, have been described from the Hex River Mountains (Boelhouwers, 1995a). The bending of grass roots in the turf-banked lobes and the expulsion of dowels inserted into the fine centres of lobes suggests contemporary movement associated with differential frost heave to a depth of 5 cm, which is comparable to the depth of freeze already discussed for this region (see Boelhouwers, 1995a; 1998). In addition, a variety of turf-banked/stepped microrelief forms have been described from the Eastern Cape and Lesotho/KwaZulu-Natal mountain regions, including amongst others *terracettes* (Hanvey and Marker, 1992; Kück and Lewis, 2002), *crescentic terracettes* (Harper, 1969; Boelhouwers, 1988), *turf-banked terraces/terracettes/lobes* (Dardis and Granger, 1986; Kück and Lewis, 2002) and *gelifluction terraces* (Lewis, 1988b; 2008a; Kück and Lewis, 2002). Although the distinction between the various *terminological*-forms may not always be clear, terracettes consist of miniature step-like features in a parallel manner, having bare risers and vegetated treads which contour the slopes (see Ward, 2004), whilst turf-banked terraces have bare treads and vegetated risers (Butler *et al.*, 2004:261). In contrast, solifluction lobes are tongue-shaped accumulations of material, often with steep bulbous frontal risers; these are subdivided according to the presence or absence of vegetation into turf-banked and stone-banked varieties (Benedict, 1976; Kinnard and Lewkowicz, 2006). Whilst the terracettes described from the Eastern Cape and Lesotho/KwaZulu-Natal mountains have mean riser heights ranging from 20 to 55 cm, and mean tread widths of 70 cm (Hanvey and Marker, 1992; Kück and Lewis, 2002); turf-banked terraces have higher risers (40-100 cm) and average 169 cm in width (Kück and Lewis, 2002).

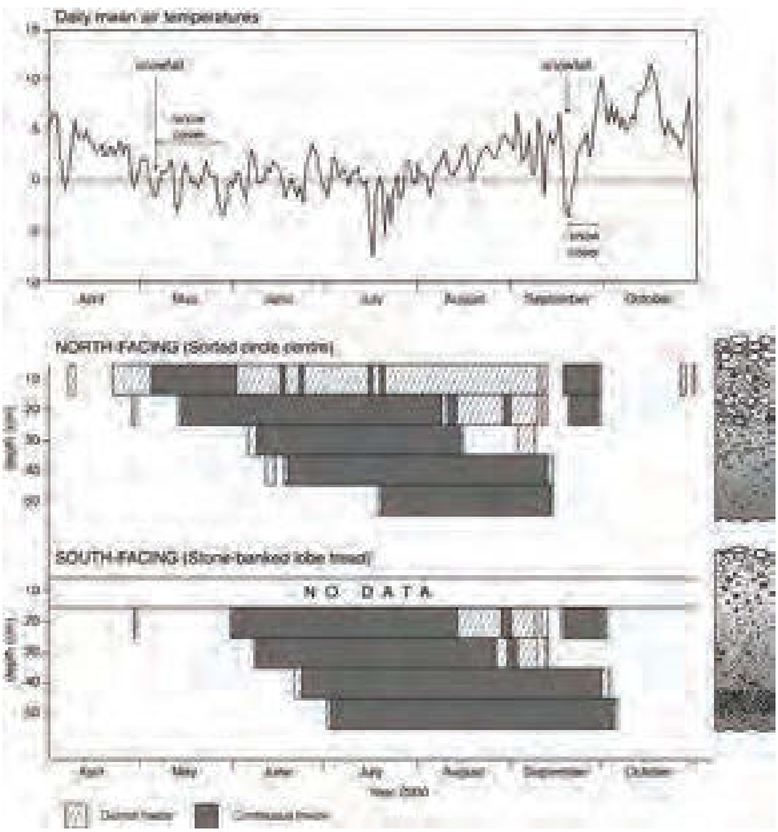

Figure 9.8. Duration of ground freeze at various depths against daily mean air temperatures during 2000. Sedimentary sketch profiles are indicated on the right (after Grab, 2004).

A study of turf-banked terraces in the Eastern Cape Drakensberg has included the monitoring of ground thermal conditions and sediment movement rates at variable microscale settings through the terraces (Kück and Lewis, 2002). Downslope movements of sediment were recorded to depths of 13 cm in winter and up to 5.8 cm downslope over an 18-month period. Although it is widely acknowledged that terracettes and/or turf-banked terraces are likely polygenetic in origin (e.g. animal trampling, fire, wind, surface runoff – see Boelhouwers, 1991b; Hanvey and Marker, 1992; Butler *et al.*, 2004), the ground thermal data and presence of ice lenses and interstitial ice suggest that cryogenic processes are at least partially responsible for current solifluction landform development and ongoing activity in the region (Kück and Lewis, 2002).

A variety of stone-banked lobe forms of variable dimensions and shapes have been described from the Hex River Mountains and high Drakensberg (see Boelhouwers, 1994; 1995a; Grab, 2000a). Smaller varieties, also referred to as *stone-banked sheets*, have lengths of 1-6 m, widths of 0.10-0.13 m and frontal heights of 0.05-0.25 m; these are known to sometimes develop in the matrix-rich sediment of large stone-banked lobe treads (Boelhouwers, 1994) or upon shallow rock scarps (Grab, 2000a). Larger varieties have total lengths up to 30 m, widths of 2-10 m and frontal heights of up to 3 m, and sometimes display a distinct raised frontal bank. Further evidence of solifluction at such sites includes frontal turf banks below stone-banked lobes and the presence of ploughing boulders (Grab, 2000a; Grab and Mills, 2011). Field measurements of stone-banked lobes have included quantitative assessments of vertical sedimentological structures, surface clast orientation and dip characteristics, and ground thermal conditions to –50 cm (Figure 9.8). Whilst sediment in the stone-banked sheets are sorted to a maximum depth of approximately 20 cm at Giants Castle below 3 260 m a.s.l., which possibly coincides with the maximum depth of local contemporary freeze (Boelhouwers, 1994), large stone-banked lobes on the south-face of Njesuthi Summit (ca. 3,400 m a.s.l.) are sorted to depths of 50-60 cm and have imbricated blocks oriented in the flow direction. Whilst it was first suggested that this may provide an indication of likely palaeofreezing depth (Grab, 2000a), a subsequent study confirmed contemporary freezing to at least 50 cm depth for >3 months during 2000 in such a stone-banked lobe tread (Figure 9.8) (Grab, 2004). It is thus imperative that detailed field monitoring be taken at particular landform sites before such landforms are classified as either active or relict, and the outcomes may well differ across the landscape between microsites.

Recent international work has focused on the application of aerial photography and GIS to map, provide morphological classifications and determine rates of movement of solifluction landforms (Ridefelt and Boelhouwers, 2006; Ridefelt *et al.*, 2009; 2010). In addition, considerable focus is now placed on physical modelling of solifluction within laboratory simulation settings, which should improve insights into the internal dynamics of solifluction processes (Harris *et al.*, 2008; Kern-Luetschg and Harris, 2008). Field surveys in the Drakensberg at approximately 2 300-2 500 m a.s.l. have identified extensive clusters of solifluction deposits, thus also providing ongoing opportunities for mapping and monitoring at lower altitudes within the Clarens Formation sandstones.

4. Inactive (relict) periglacial phenomena

Relict southern African periglacial phenomena reported in the literature are typically on a larger scale than the present day active features, and include ice-wedge casts (Lewis and Dardis, 1985), rock glaciers (Lewis and Hanvey, 1993), pronival (protalus) ramparts (Marker, 1990; Lewis, 1994; Grab and Mills, 2011), large sorted patterned ground (Grab, 2002c; Sumner, 2004a), a host of openwork block deposits (stone-banked lobes (*stone garlands*), blockstreams, blockfields, screes) (Marker, 1986; Lewis, 1996; 2008a; Boelhouwers, 1999a, b; Grab, 1999a; Boelhouwers *et al.*, 2002; Boelhouwers and Sumner, 2003; Sumner and De Villiers, 2002; Sumner, 2004b) and slope deposits of periglacial origin (Hanvey *et al.*, 1986; Boelhouwers, 1991a, 1999a; Hanvey and Lewis, 1991; Lewis and Hanvey, 1991; Grab and Mills, 2011) (Table 9.1). Additional macroscale landscape development associated with periglacial/glacial processes include *nivation cirques/niches* (Sparrow, 1964; Nicol, 1973; Dyer and Marker, 1979; Sänger, 1988; Marker, 1990; Boelhouwers, 1991a), asymmetric valleys (Meiklejohn, 1992; 1994), escarpment cutbacks (Hall, 1994) and basalt terraces (Grab *et al.*, 2005). The identification of, and palaeoenvironmental significance of such landforms has proven highly controversial and has been discussed in previous literature (Le Roux and Marker, 1990; Hall *et al.*, 1991; Hall, 1992: Boelhouwers 1995b; Sumner and Hall, 1995; Grab and Hall, 1996; Grab, 2000b; Boelhouwers and Meiklejohn, 2002) This section focuses on relict southern African periglacial landforms that have received considerable interest and debate, and provides an overview on the general palaeoenvironmental inferences made.

4.1 Rock glaciers

Rock glaciers are lobate or tongue-shaped landforms of frozen debris (French, 2007) typical of cold mountain environments, originating either as gradually creeping permafrost and ice-rich debris on non-glacierised slopes, or as debris-covered remnant glaciers in permafrost-free areas (Haeberli *et al.*, 2006) (Figure 9.9). Hence, relict rock glaciers would imply either the existence of past permafrost or glacial ice. However, it remains unclear whether rock glaciers represent a distinct cryogenic landform, or whether they should be classified as a member of a broader geomorphic continuum, with rockfall talus slopes and ice glaciers as end geomorphic members (Humlum *et al.*, 2007). Although the latitudinal-altitudinal distribution of rock glaciers generally follows the mean annual air 0 °C isotherm, recent work has argued that they are possible in dry cryogenic mountain environments where mean annual air temperatures may be as high as 0 to +1 °C (Azócar and Brenning, 2010). Some debris accumulations have been interpreted as relict rock glaciers in the Bottelnek area of the Eastern Cape (ca. 1 800 to 2 100 m a.s.l.) (Lewis and Hanvey, 1993), but have been the focus of much debate as to their origin due to a lack of diagnostic features (Boelhouwers and Meiklejohn, 2002) and the climatic conditions required for their formation. No rock glaciers have thus far been identified in the Lesotho highlands, which reach significantly higher altitudes.

Figure 9.9. An active rock glacier in the Valais, Switzerland.

4.2 Pronival (protalus) ramparts

Pronival ramparts are a product of debris accumulation below snowbanks through mechanisms including the sliding, rolling and bouncing of debris over the snowbed (Ballantyne and Kirkbride, 1986), supranival debris flows (Ono and Watanabe, 1986), snow avalanching (Ballantyne, 1987), subnival debris transport (Shakesby *et al.*, 1995) and snow- (creep/push) processes (Shakesby *et al.*, 1999). The ramparts tend to contour slopes but are usually curved, sinuous or complex in planform, and reach thicknesses of up to 10 m and lengths of up to 300 m (Shakesby, 1997). The internal composition varies from diamictons to coarse (subangular to very angular) openwork rock deposits and the ramparts are typically located within 45 m of the talus foot (Ballantyne and Kirkbride, 1986). Although it has been argued that ramparts extending more than 50-100 m from the talus foot would imply snow-bed thicknesses conducive to developing glacial ice, and would thus be a product of glacial movement (Shakesby, 1997), there has been much controversy over the correct identification of process origins for such talus foot ramparts beneath cliffs in cold environments. In southern Africa, ridges up to 298 m in length and 83 m in width, reported from approximately 2 000 m a.s.l. in the Eastern Cape, were initially identified as pronival ramparts (Lewis, 1994), but more recent investigations suggest these to be structural and fluvial erosional in origin (Hedding and Nel, 2010). Considerably smaller ridges (15-22 m in length; 4-6 m in width) at Thabana Ntlenyana (ca. 3 440 m a.s.l., Lesotho Drakensberg), have been interpreted as miniature versions of pronival ramparts with a possible snowcreep/snowpush origin (Grab and Mills, 2011). Further north, at Golden Gate Highland National Park, ridges with an apparent nival origin have been located between 2 100-2 300 m a.s.l. (Marker, 1990).

4.3 Large sorted patterned ground

Relict large sorted patterned ground has been reported from a variety of high altitude sites in the Hex River Mountains (Boelhouwers, 1999a) and Lesotho/KwaZulu-Natal Drakensberg (Boelhouwers, 1994; Grab, 2002c; Sumner, 2004a). Large sorted stripes consisting of coarse diamicts and reaching lengths of 150 m and widths of 2 m are associated with extensive openwork block accumulations in the Hex River Mountains and the mountains of Lesotho, whilst circular patterns in Lesotho generally consist of gravel to cobble dominated centres surrounded by openwork block-supported borders. Smaller varieties at Mafadi summit have mean centre diameters of 0.7 m and borders of 0.6 m, whilst those at Thabana Ntlenyana have average centre diameters of 2.5 m and borders of 2.6 m (Grab, 2002c). An important task has been the determination and characterisation of earth material sorting with depth, as this may provide indications of past freeze depths. To this end, relict sorted patterns have been excavated and detailed sedimentary profiles constructed; this entails the measurement of clast *a-b-c* axis lengths vertically and horizontally through patterns for shape and size characterisation, as well as sieving finer sediments for particle size distributions through pattern centres (see Grab, 2002c; Sumner, 2004a). Such studies have demonstrated depths of sorting varying from approximately 0.3 m to possibly >1.0 m. The ground thermal monitoring through such patterns or in environmental contexts near such patterns has added insight into contemporary cryo-thermal dynamics and assisted discussions on palaeoenvironmental implications (see Sumner, 2003; Grab, 2004). For instance, such thermal profiles have enabled zero isotherm projections to be made, indicating both potential contemporary freezing fronts and depths of freeze under depressed temperature conditions (see Figure 9.10; Sumner, 2003). Internationally, the study of large sorted patterned ground continues to focus on morphometrics and spatial modelling, particularly as these have valuable application for comparisons with extra-terrestrial forms (e.g. Mangold, 2005; Treml *et al.*, 2010).

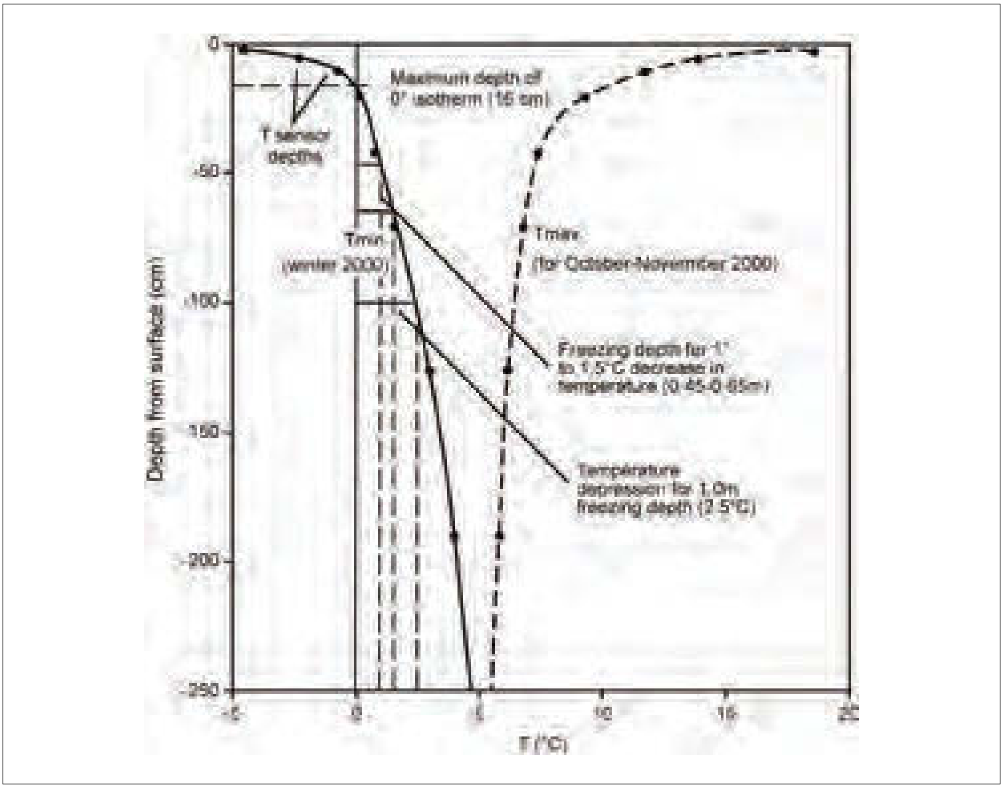

Figure 9.10. Thermal profile derived from the minimum and maximum temperatures recorded on probes between 17 May 2000 and 6 November 2000. T_{min} is representative of the 2000 winter. T_{max} underestimates summer maximum. Two scenarios for zero isotherm penetration under reduced temperatures are illustrated (after Sumner, 2003).

4.4 Openwork block deposits

Openwork block deposits are widespread across the core cryogenic mountain regions of southern Africa, as also in many uplands, mountain ranges and mid- to low-latitude regions throughout the world (French, 2007). The landforms include blockfields, blockslopes and blockstreams (sometimes also associated with large sorted patterns and/or stone-banked lobes (garlands)), which typically consist of the accumulation of sub-angular to angular boulders or blocks covering areas ranging from several hundred to thousands of square meters (Firpo *et al.*, 2006; Park Nelson *et al.*, 2007). Openwork block deposits have been classified into autochthonous (formed *in situ*) and allochthonous (formed by block emplacement; i.e. mass movement) forms.

Most relict openwork block deposits in the higher regions of the Hex River Mountains, Eastern Cape mountains and Lesotho/KwaZulu-Natal Drakensberg are allochthonous forms (Boelhouwers, 1999a; Boelhouwers *et al.*, 2002; Sumner and De Villiers, 2002), although autochthonous varieties are not uncommon to the Lesotho Highlands (Boelhouwers, 1999b). The block deposits in the Western Cape are up to 6 ha in extent (Boelhouwers and Sumner, 2003), whilst the clast-supported diamictons in Lesotho may reach 1.1 km in length, 75 m in width, and at least 4 m in depth (Figure 9.11) (Boelhouwers *et al.*, 2002). Southern African research has focused on detailed field analysis and desktop approaches to better quantify the spatial dynamics, morphology and sedimentology of openwork block deposits.

Some of these methods have included mapping openwork block units from ortho-photographs and then verifying the mapped units through ground-truthing (Sumner and De Villiers, 2002), undertaking surface profiles, determining clast size and fabrics (i.e. orientation and dip of clasts), and establishing the relative age (duration) of clast surface exposures through Schmidt hammer rebound and rock surface roughness measurements within various microtopographic/morphological settings (e.g. Boelhouwers *et al.*, 2002; Sumner and De Villiers, 2002; Sumner, 2004b). Whilst the platey angular to sub-angular clasts in the Hex River Mountains demonstrate downslope fabrics and vertical sorting to 1.5 m (Boelhouwers, 1999a), those in the Eastern Cape and Lesotho mountains are more typically sub-angular to sub-rounded, have well-developed bi- or trimodal clast orientations, but with the primary alignment of clasts being parallel to the local maximum slope gradient (Figure 9.11) (Boelhouwers *et al.*, 2002; Sumner and De Villiers, 2002; Sumner, 2004b). Whilst many of the allochthonous forms in the Western and Eastern Cape mountains originate from scarps as screes and/or rock avalanche deposits, the lobate fronts and preferred downslope clast orientations for many features in the Hex River Mountains and the mountains of Lesotho suggests they are a product of slow mass wasting under periglacial conditions.

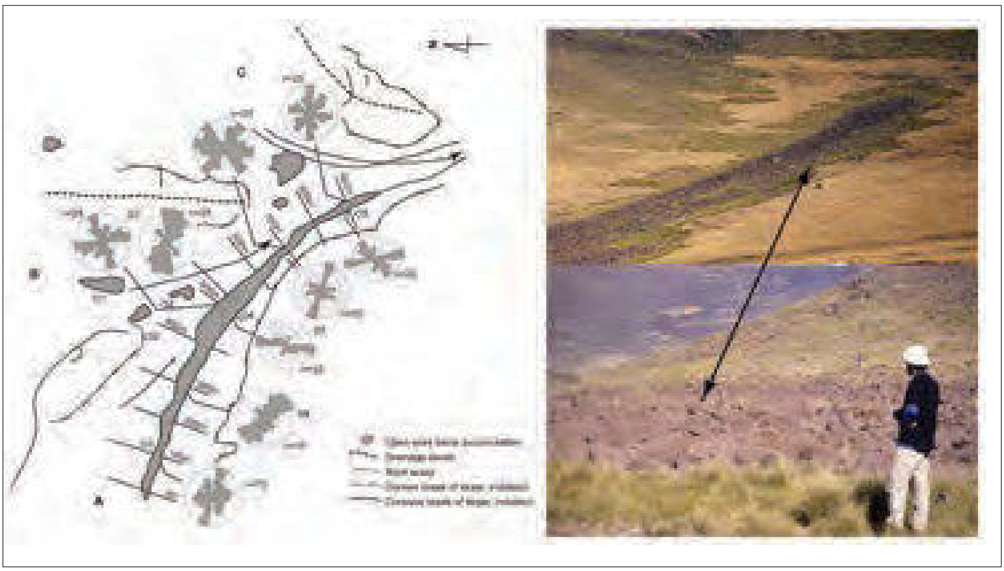

Figure 9.11. One of the longest known block accumulations (blockstream) in southern Africa, in the proximity of Sani Pass, Drakensberg. The diagram indicates fabrics for surface clasts, location of transects and rock surface relative-age dating sites (after Boelhouwers *et al.*, 2002).

Ongoing international work on openwork block deposits continues to place emphasis on mapping their spatial distribution, understanding their weathering and sedimentological contexts, and ultimately improving our understanding of their Quaternary origins (e.g. Firpo *et al.*, 2006; Park Nelson *et al.*, 2007; Goodfellow *et al.*, 2009; Ballantyne, 2010). However, equally importantly, several studies have now examined the contemporary thermal regimes through coarse blocky deposits and findings suggest that they are conducive to establishing negative temperature anomalies (Harris and Pedersen, 1998; Gorbunov *et al.*, 2004; Juliussen and Humlum, 2008), which also has implications for modelling thermal regimes during periods of known past colder periods. To this end, there is still scope to evaluate the contemporary thermal regimes through southern African openwork block deposits.

5. Palaeoenvironmental implications of relict periglacial phenomena

Macro-scale periglacial phenomena reported from southern Africa (e.g. nivation hollows, cutbacks, etc.) are arguably products originating through geological time scales of variable climates and geomorphic histories (Grab and Hall, 1996; Mills *et al.*, 2009a), and given their previous critical reviews (see Le Roux and Marker, 1990; Hall *et al.*, 1991; Sumner and Hall, 1995; Grab and Hall, 1996; Grab, 2000b; Boelhouwers and Meiklejohn, 2002), will not be discussed further in this chapter. Rather, some of the more recently (ca. last two decades) documented palaeoenvironmental inferences made are briefly summarised.

On the basis of morphological and sedimentological characteristics, it has been proposed that the openwork block deposits and large sorted stripes at the summit of Matroosberg indicate deep seasonal freeze to 2 m depth during the Last Glacial Maximum (LGM), with an associated 7 to 8 °C reduction in temperature, but would not necessarily have required permafrost (Boelhouwers, 1999a). In addition, their development during the late Quaternary would have necessitated sufficient soil moisture and thus most likely coincided with an earlier seasonal onset of cyclonic precipitation (Boelhouwers, 1999a).

Results from openwork block deposits from Lesotho and the Drakensberg suggest that the Last Glacial Maximum (LGM) experienced upfreezing of clasts and cryogenically-induced slope mobility associated with enhanced deep seasonal freeze, particularly on south-facing slopes (see Boelhouwers *et al.*, 2002; Sumner 2004a). Although, based on the occurrence of large sorted circles, late Quaternary permafrost has been proposed for high altitude interfluves in eastern Lesotho (Grab, 2002c), the more general view is that there is no unequivocal evidence of past permafrost and that most landforms probably originated during a relatively dry period with little snow accumulation, thus permitting deep seasonal freeze to depths of at least 2 m (Boelhouwers *et al.*, 2002; Sumner, 2003; 2004b). Yet, it has also been argued that large stone-banked lobes were likely active during cooler late Holocene periods (Grab, 2000a), and notably, contemporary deep seasonal freeze to depths exceeding –50 cm is still possible during some years on high (>3 400 m a.s.l.) south-facing summits (Grab, 2004).

5.1 Challenges and prospects

A particular limitation has been the inability to ascertain, with a reasonable degree of certainty, the age of relict cryogenic landforms in southern Africa. Organic material has been radiocarbon dated from a variety of landforms including apparent relict rock glaciers and pronival ramparts (Lewis and Hanvey, 1993; Grab and Mills, 2010). However, such organic horizons may not necessarily coincide with the phase of landform development. Similarly, despite many high altitude sedimentary sequences having been radiocarbon dated, providing relatively high resolution Quaternary chronologies depicting past climatic and slope process histories (e.g. Hanvey and Marker, 1994; Marker, 1994; 1995; 1998), these palaeophases cannot be adequately connected to the formative (active) stage(s) of periglacial landforms; such an attempt would be too speculative. Relative age determinations for scarp and openwork block surfaces suggest an increase in downslope age (Boelhouwers *et al.*, 2002; Sumner, 2004), yet, despite the assumption that they probably originate from the LGM (Boelhouwers, 1999a; Boelhouwers *et al.*, 2002), their precise age remains unclear. Future work could consider exposure dating of surface blocks within cryogenic landforms (e.g. cosmogenic ^{36}Cl for basalt), as has successfully been undertaken for similar landforms in Australian cryogenic mountain environments (see Barrows *et al.*, 2004). Blockstreams in the Snowy Mountains have been dated to about 22 ka and have apparently become stable since at least 17 ka (Barrows *et al.*, 2004), which poses the question: are southern African blockstreams (and large sorted patterns etc.) of possible similar age, and to what extent did past phases of periglaciation coincide inter-continentally across the southern hemisphere? Further work monitoring ground thermal conditions within relict dated landforms in a variety of topo-altitudinal settings would surely improve knowledge on contemporary ground freeze and provide a better position from which to model likely palaeofreeze.

6. Glacial phenomena

Historically, the idea that southern Africa experienced mountain glaciation during the Quaternary has proven highly controversial. Many publications have advocated Late Quaternary (Last Glacial cycle) glaciation in southern Africa (e.g. Sparrow, 1967a, b; Marker and Whittington, 1971; Harper, 1969; Dyer and Marker, 1979; Borchert and Sänger, 1981; Sänger 1988; Hanvey and Lewis, 1990; Marker, 1991; Lewis, 1994; 2008a, b; Hall, 1994; Grab, 1996b; Lewis and Illgner, 2001; Mills and Grab, 2005; Mills *et al.,* 2009a, b), but many of these studies have been criticised and it has been suggested that some evidence has been misinterpreted (Boelhouwers and Meiklejohn, 2002; Hall, 2004; Osmaston and Harrison, 2005). The perceived aridification of southern Africa during the Last Glacial cycle has been suggested by some researchers to demonstrate that there was insufficient precipitation to support mountain glaciation during the Quaternary (e.g. Hall, 2004).

Given that the mountains of southern Africa reach substantial altitudes, it is perhaps reasonable to conjecture that the highest ranges could have experienced glaciation during the Quaternary period, especially given the evidence for glaciation in other regions of the southern hemisphere at similar latitudes and altitudes (e.g. Espizua, 1993; Barrows *et al.,* 2001). However, to date there is no evidence reported to suggest repeated glaciation of southern Africa during the cyclic global glacial-interglacial phases that typify the Quaternary, and all evidence of Quaternary glaciation has been assumed to have occurred during the Last Glacial cycle. In this section the reported geomorphological evidence supporting glaciation is presented for the Western Cape, Eastern Cape and Lesotho Highlands, with some examination on the environmental implications of suggested Late Quaternary glaciation.

6.1 Western Cape mountains

Late Quaternary glaciation in the Western Cape mountains is probably the most contentious of all in southern Africa; nevertheless, it has been suggested (based largely on the interpretation of aerial photography) that the Western Cape mountains were characterised by plateau ice fields, cirque and valley glaciers during the Pleistocene. This is based on the interpretation of cirque basins, moraines, glacial polishing and outwash fans in the Matroosberg region, reaching an altitude of 2 249 m a.s.l. (Borchert and Sänger, 1981; Sänger, 1988). The suggested cirques and glacially-polished areas are of poor condition and the lack of striae is attributed to the nature of the bedrock (sandstone) which does not preserve striation and polishing for long, due to weathering of rock surfaces (Borchert and Sänger, 1981; Sänger, 1988). The suggestion of Late Quaternary glaciation in the Western Cape implies considerable shifts in palaeoclimatic conditions; Sänger (1988) proposes a temperature depression of 10 °C, resulting in annual temperatures for Matroosberg Peak of –4 °C and temperatures only rising above 0 °C in mid-summer. In addition, Sänger (1988) argues that the climate was wetter during Quaternary glaciation due to a northward shift in the westerlies.

The suggested Late Quaternary glaciation in the Western Cape has been challenged by Boelhouwers and Meiklejohn (2002), who on the basis of field observations, suggest that the moraines described by Sänger (1988) are erosional remnants of bedrock, and that steep valley heads owe their origin to structural conditions rather than representing cirque formation, whilst the outwash fans may simply reflect sub-aerial conditions during seasonal snowmelt. As such, the evidence for Late Quaternary glaciation in the Western Cape remains highly speculative.

6.2 Eastern Cape mountains

Evidence for Quaternary glaciation in the Eastern Cape mountains is based on field mapping and sedimentological investigations centred on Killmore (30º 57'S, 27º 56'E) and Mount Enterprise (31º 09'S, 27º 59'E) (Lewis, 1994, 2008a; Lewis and Illgner, 2001). Lewis (1994, 2008a) describes an apparent well-developed cirque and moraine from Killmore, between 2,000 and 2 040 m a.s.l., with a cirque backwall rising to 2 200 m a.s.l. The moraine is ~1.5 km long and up to 17 m high and was initially interpreted as a pronival rampart (Lewis, 1994). This interpretation was later revised based on the distance from ridge crest to talus slope being incompatible with those accepted for pronival ramparts (Ballantyne and Benn, 1994; Lewis 2008a, b); yet as mentioned earlier, this feature may be structural in origin.

Three ridges ~300 m long and up to 5 m high at Mount Enterprise are associated with diamictons containing striated clasts; these have been interpreted as moraines, representing a small niche/cirque glacier up to 50 m thick during the Late Pleistocene (Lewis and Illgner, 2001; Lewis, 2008b). It has been proposed that such marginal glaciation was the result of local topoclimatic influences, whereby extensive snow-blow contributed significantly to glacier mass-balance, resulting in a lower glacier equilibrium line altitude than would otherwise be expected (Lewis and Illgner, 2001). Reconstruction of the glacier by Lewis and Illgner (2001) suggests that annual temperatures may have been approximately 17 ºC lower than at present, which is a substantially greater temperature depression than that calculated from other proxy climate data for southern Africa, which typically range between 5-7 ºC for the LGM (Talma and Vogel, 1992; Partridge *et al.,* 1999; Holmgren *et al.,* 2003). Analyses of the climatic implications associated with marginal glaciation (Coleman *et al.*, 2009) highlight that localised topoclimatic factors such as enhanced snowblow significantly skew climatic assumptions such as those used by Lewis and Illgner (2001), rendering regional climatic reconstructions based on individual reconstructed glaciers unreliable.

6.3 Lesotho/KwaZulu-Natal Drakensberg

Observations of *cirque-shaped hollows* located at altitudes above 2 900 m a.s.l. in the Drakensberg led to the first suggestions for Quaternary glaciation in southern Africa (Sparrow, 1967a, b; Harper, 1969; Marker and Whittington, 1971). Morphometric analysis of these and several hundred other similar hollows was subsequently undertaken throughout the Lesotho Highlands, from which it was concluded that 27% of these show characteristics typical of glacial cirques (Dyer and Marker, 1979; Marker, 1991). Bog sequences obtained from sites within these hollows have yielded relatively high-resolution information on palaeoclimatic oscillations throughout the Holocene, suggesting that the cirques were most recently excavated during the Last Glacial cycle (Marker, 1994; 1995). Yet, 75% of the 577 hollows plotted in the highlands of Lesotho are located on the warmer north-facing slopes (Dyer and Marker, 1979; Marker, 1991), which is a problem when invoking a glacial origin, as these north-facing slopes are exposed to maximum insolation (Mulder and Grab, 2002) and are thus less likely to preserve snow and ice for long periods. An alternative view is that the north-facing hollows owe their origin to a combination of lithological and climatological factors, which would enhance weathering and subsequent transport of material (Grab and Hall, 1996), rather than a glacial erosional origin.

It has been proposed that debris ridges emanating from steep, high altitude (>3 200 m a.s.l.) south-facing cutbacks on the eastern flank of the Great Escarpment provide evidence for localised plateau and niche glaciation on higher summits during the Late Quaternary (Figure 9.12) (Hall, 1994; Grab, 1996b). A process of glacial dumping over the escarpment sidewalls to produce the moraine-like ridges has been suggested (Grab, 1996b); yet more detailed geomorphological and sedimentological investigations of these deposits are required to verify a glacial origin.

Figure 9.12. Debris accumulations below the main Drakensberg escarpment wall, which have previously been suggested to originate through glacial deposition and subsequent fluvial incision.

Figure 9.13. An example of a moraine from the Leqooa Valley, southern Drakensberg.

Based on an integrated multi-method approach, incorporating geomorphology, macro- and microscale sedimentology, glacier reconstruction and mass-balance modelling, a number of debris ridges along the high Drakensberg Escarpment have recently been interpreted as moraines and dated to the later part of the Last Glacial cycle (Mills and Grab, 2005, Mills *et al.*, 2009a, b). The described features all occur on steep, south-facing slopes at altitudes exceeding 3 000 m a.s.l., are up to 300 m in length and 16 m in height (Figure 9.13). Geomorphological and sedimentological analyses indicate that the moraines preserve evidence for both active and passive transport processes, with microscale analysis demonstrating transport in a subglacial traction zone. Detailed microscopic analysis reveals a complex depositional history for the moraines, which have undergone substantial modification under paraglacial conditions (see Mills *et al.,* 2009a, b) (Figure 9.14). The application of glacier reconstruction and mass-balance modelling to the sites indicates that it was viable for glaciers to have existed in this region, and that by contrast with glaciation in the Eastern Cape; glaciers were not dependent on snowblow (Mills, 2006; Mills *et al.,* 2009b). It is likely that the predominant factor determining the location of glaciers along the south-facing slopes of the Lesotho Highlands during the late Quaternary was reduced insolation. The greater levels of insolation received on north-facing slopes, results in higher temperatures and greater temperature variability on such slopes (Meiklejohn, 1994; Grab, 2007), which would have limited the survival of snowpatches and the development of niche glaciers. During summer, the relatively gentle north-facing slopes in the Drakensberg receive only slightly more radiation than the south-facing slopes, whilst in winter, south-facing slopes receive about 50% less radiation (Tyson *et al.,* 1976; Mills, 2006). It has been widely demonstrated that solar radiation has had an important control in the preservation or disappearance of small glacier remnants in the Pyrenees (Chueca and Julián, 2004), on Kilimanjaro (Rosqvist, 1990), Mount Kenya (Young and Hastenrath, 1987) and in New Guinea and the South American Andes (Kaser, 1999), which is also likely to have been the case in the Lesotho Highlands.

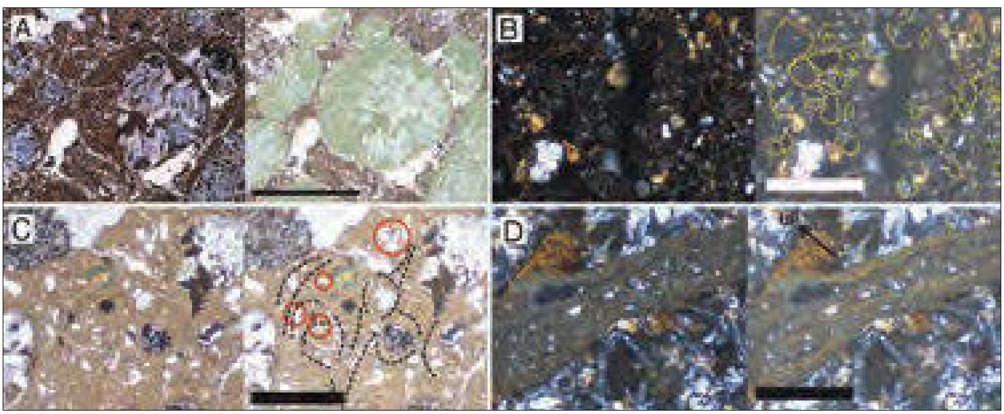

Figure 9.14. Observed micromorphological structures from samples recovered from potential moraines along the Drakensberg Escarpment (Mills *et al.*, 2009a,b). A) Intraclasts formed around *core stones*, representing reworking of the original sediment body. Plane polarised light, scale bar indicates 5 mm. Analysis of the intraclasts allows investigation of the sediment prior to re-working through earth-flow. B) Re-orientation of clays to form a *skelsepic* plasmic fabric within an intraclast. Cross-polarised light, scale bar indicates 1 mm. Within the intraclasts, a host of deformation structures may be identified, from the smallest scale, whereby clay skins around individual sand grains reflect rotation of grains in a deforming sediment body. C) Associated rotational deformation structures (turbates) and fractured basalt grains, within an intraclast. Plane-polarised light, scale bar indicates 1 mm. Both ductile deformation features and indicators of high confining stresses (fractured grains) provide strong evidence pointing towards active transport within the subglacial route through the glacial debris cascade. D) Re-orientation of clays along shear planes (masepic and unistrial plasmic fabric). Cross polarised light, scale bar indicates 1 mm. Localised shear, in this case associated with a larger grain *ploughing* through a finer matrix further indicates a complex sequence of sediment deformation, typically associated with a subglacial traction zone.

7. Palaeoenvironmental implications of former glaciation

The spatial extent of glaciation in southern Africa during the Last Glacial cycle is limited to very few locations that are dependent on specific local topographic conditions. Areas of contemporary late-lying snow provide insight into the effects of topography and the relationship between areas that would have sustained glaciers and those where periglacial processes would have dominated during past cold phases (Grab *et al.,* 2009). The spatial distribution of periglacial deposits such as earth hummocks, stone/turf-banked lobes, block deposits and sorted patterned ground coincides with topographic areas that limit snow accumulation, whilst there is a strong spatial association between areas of late-lying snow and moraines (Grab *et al.,* 2009), suggesting that periglacial processes would have dominated during cold periods, with only localised glaciation.

The demonstrable presence of Late Quaternary glaciation in the Lesotho Highlands, based on a 6 ºC temperature depression as supported by proxy data for southern Africa (Talma and Vogel, 1992; Holmgren *et al.,* 2003), suggests that an increase in atmospheric winter precipitation would have been necessary to sustain the glaciers (Mills *et al.,* 2009b; Carr *et al.*, 2010). This is in stark contrast to the suggested 30% reduction in precipitation during the Last Glacial Maximum (LGM) (Partridge *et al.,* 1999; Tyson and Partridge, 2000) and the reduced snow cover necessary to allow for deep frost penetration and the formation of the large-scale periglacial features found in the high Drakensberg, as noted earlier in this chapter. The suggested accumulation at the Equilibrium Line Altitude (ELA) of the palaeoglaciers (~1 500 mm/a), if representative of likely LGM conditions, indicates substantially increased snow accumulation at the sites and is nearly double the modern rate of total precipitation recorded close to the area (Nel and Sumner, 2008).

Evidence for increased winter precipitation during glacial periods was first proposed by Van Zinderen Bakker (1967), who suggested that winter rains expanded to ~25°S across all of southern Africa. Cockcroft *et al.* (1987) further argued that this zone expanded to 30°S over the eastern parts of southern Africa. Consequently, the pattern of precipitation in the Drakensberg region during the LGM would have been less seasonally constrained, allowing for more precipitation to fall as snow than at present. Cold fronts occur most frequently in winter when the amplitude of westerly disturbances is greatest (Tyson and Preston-Whyte, 2000). An increased frequency of cold fronts during autumn, winter and spring would have produced heavier snowfalls and in turn increased the snow cover at high altitudes (Grab and Simpson, 2000). There has been considerable debate on whether the Southern Hemisphere westerlies were displaced pole-wards or equator-wards during glacial periods (Lamy *et al.,* 1998; Wyrwoll *et al.,* 2000; Sugden *et al.,* 2005). Evidence from southern Africa supports the contention that westerlies were displaced equator-wards due to the expansion of Antarctic sea ice during the LGM; therefore the climate of southern Africa would have been more regularly and more intensely influenced by westerlies during the LGM (Van Zinderen Bakker, 1976, 1983; Cockcroft *et al.,* 1987; Stuut *et al.,* 2004; Chase and Meadows, 2007). This is predominantly evidenced from the western regions of southern Africa, however this is mainly due to the relative abundance of archives from these regions in comparison with the more central and eastern regions of southern Africa. The increased levels of snowfall required to sustain the glaciers in Lesotho strongly support an equator-wards shift in the westerlies and an associated influence on the climate of Lesotho, to the extent that more precipitation was supplied to the region in the form of snow around the LGM. Suggested greater precipitation is reinforced through a growing body of literature from a variety of proxy data which support evidence for increased precipitation during phases of the Last Glacial period in southern Africa (Butzer *et al.,* 1973; Butzer, 1984; Shaw and Thomas, 1996; Meadows and Baxter, 1999; Huntsman-Mapila *et al.,* 2006; Chase and Meadows, 2007).

8. Future prospects

It is evident from the preceding discussion that whilst often fragmentary in nature, significant research has been undertaken on cryogenic processes and geomorphology within southern Africa. It is apparent from this research that numerous active and relict features indicate that cryogenic processes have operated through the Late Quaternary, and continue to operate, at a small-scale, to the present day. Further, it seems finally that the long-running debate regarding whether southern Africa experienced Quaternary glaciation is reaching resolution, in favour of marginal glaciation during the Late Quaternary. While the idea of Late Quaternary glaciers in southern Africa is gaining acceptance, there are still key questions that future glacial geomorphological research needs to address. There are possibly other sites of former marginal glaciation along the Drakensberg Escarpment and in the Eastern and Western Cape mountains, and a systematic attempt to identify and test the geomorphological evidence is surely the next phase towards understanding past glaciation in the region. Also, a greater understanding of the nature of glaciation (temperate/warm-based or cold-based glaciers) is required: the evidence of subglacial sediment production and basal slippage from the High Drakensberg would suggest the former (Mills *et al.*, 2009a, b; Carr *et al.*, 2010), while the relative lack of observed large-scale landscape modification has been claimed to suggest the latter (Hall, 2010). However, to resolve such issues, it is clear that multiple method approaches, incorporating field, laboratory and modelling analyses are required, and that speculative and deductive approaches based on limited data collection or analysis are no longer sufficient to make significant steps forward.

A review on recent advances made in periglacial geomorphology has identified three main research thrusts in the near future; these include:

i. palaeoenvironmental reconstructions based on current processes and landform measurements as analogues;

ii. long-term monitoring of contemporary processes using automatic equipment; and

iii. modelling based on Geographical Information Systems (Humlum and Christiansen, 2008).

Given that recent periglacial research in southern Africa has already placed growing emphasis on the first two mentioned research thrusts, and is currently becoming more engaged with GIS work, this sub-discipline of geomorphology has surely come of age in southern Africa and made considerable international contributions towards expanding scientific knowledge on landforms and their associated formative processes.

The overarching priority, however, is to derive better dating to constrain the chronology of Late Quaternary cryospheric processes, using methods such as cosmogenic nuclide dating (Barrows *et al.*, 2001; Barrows, 2004), which is key to examining the local, regional and hemispheric climatic significance of past glaciation and periglaciation. At present, chronology is the weakest element in our understanding of cryospheric processes in southern Africa, preventing the current research findings from being placed in a global environmental change context. However, far from being an obscure and largely unknown element of the landscape of southern Africa, the study of cryospheric processes both past and present continues to yield useful information on the evolution of the mountain landscapes of the region, and potentially provide key data to address major spatial and temporal gaps in the knowledge required to build and test appropriate climate models at a regional and hemispheric scale.

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10

Coastal Geomorphology

Coastal Geomorphology

Andrew S Carr and Greg A Botha

1. Introduction

1.1 Preamble

Lying at the interface of the ocean, the atmosphere and the terrestrial landscape, coastlines are inherently complex and dynamic geomorphic environments. Coastal geomorphic systems receive energy from marine currents, tidal currents, wave action, wind action (the ultimate source of wave-derived energy), and in certain locations, fluvial activity. Much of this energy is concentrated over a relatively narrow vertical and horizontal zone. Dissipation of incoming marine energy is achieved through adjustment of the shoreface profile and the associated formation of characteristic coastal geomorphic landforms. These morphological products are characterised by significant cyclical (e.g. tidal) and stochastic (e.g. storms) adjustment to changing energy supplies and sediment availability. The concept of coastal morphodynamics (Wright and Thom, 1977; Cowell and Thom, 1994) recognises a continual co-adjustment of form and process (wave type, swash asymmetry, wave energy etc.) over a range of timescales. The inherent linkage of spatial and temporal scale in such a systems-based approach is conceptualised in Figure 10.1. The basic implication is that different techniques and approaches are required to study geomorphic processes at different spatial-temporal scales (Sherman, 1995; Summerfield, 2005). Research conducted at the smallest scales is often referred to as process geomorphology and broadly deals with the fundamental forces and mechanisms of sediment transport over timescales of seconds to years. Studies conducted over larger spatial scales, or which consider stratigraphic evidence of coastal change inevitably provide information about the longer-term evolution of the coastal landscape.

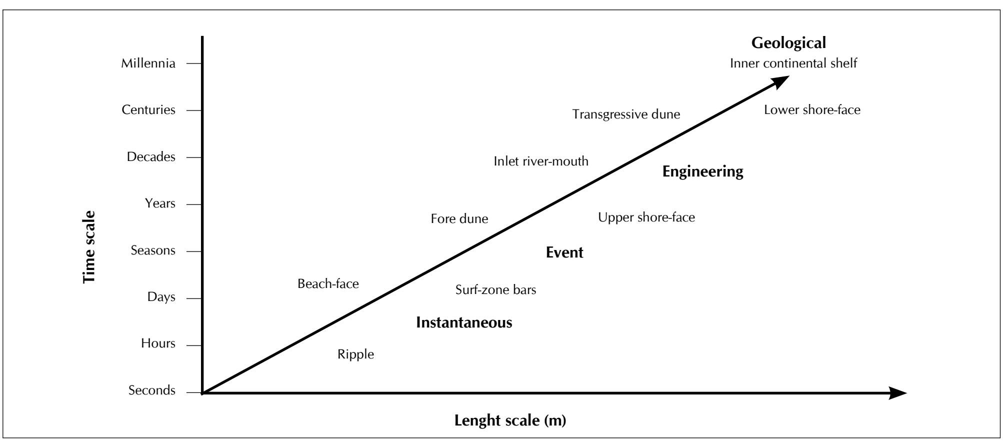

Figure 10.1. Time/Space scale relationships in coastal geomorphology (modified from Cowell and Thom, 1994).

1.2 Progress in southern African coastal geomorphic research

The following does not seek to reiterate the fundamental concepts of coastal geomorphic processes. Such information is available in general textbooks (e.g. Bird, 2000; Masselink and Hughes, 2003). Rather, in keeping with the aims of this publication we highlight major areas of progress in southern Africa since Dardis and Grindley's (1988) review. Similarly, rather than providing an exhaustive account of every coastal landform in southern Africa, we focus on those aspects of the coastal geomorphic research that have witnessed the most significant developments in our understanding. In general, process coastal geomorphic research remains relatively lacking in the southern African context and publications in the last 10 years reveal a strong bias towards studies concerning the medium to long-term evolution of coastal and near-shore landscapes. A number of process-based studies have however been carried out (e.g. Hesp *et al*., 1989; Burkinshaw *et al*., 1993; Miller and Mason, 1994; Holmes and Luger, 1996; Olivier and Garland, 2003; Garden and Garland, 2005), with a notable focus on aeolian geomorphology. For instance, studies of the Alexandria Dunefield in the Eastern Cape Province of South Africa have yielded a large body of published work (McLachlan, 1988; Illenberger, 1988; Illenberger and Rust, 1988; Hesp *et al*., 1989; Rust *et al*., 1990 Illenberger and Verhagen, 1990). Geomorphic processes relevant to coastal management and planning have also been discussed (e.g. Hellström, 1996; La Cock and Burkinshaw, 1996; Smith *et al*., 2007; Smith *et al*., 2010). Systematic study of the region's coasts has continued, and noteworthy contributions include the morphodynamic classification schemes developed for dune systems (Illenberger and Burkinshaw, 2008) and estuarine systems (Cooper, 2001). Such studies provide valuable insights into the fundamental controls on coastal geomorphic diversity in this region.

A major area of research has concerned the long-term evolution of the southern African coastline and particularly geomorphic responses to long-term (Quaternary and Holocene) relative sea level and climate change. This research has exploited developments in geochronology (e.g. luminescence dating techniques) and sub-marine geological and geophysical analyses (e.g. marine side-scan sonar). These data have provided important insights into the nature and timing of coastal landscape evolution over millennial to multi-millennial timescales and a close linkage with the region's coastal archaeological record has provided additional research impetus (Marean *et al*., 2007; Bateman *et al*., 2008; Fuchs, *et al*., 2008; Fisher *et al*., 2010). Luminescence dating is a particularly powerful tool in coastal environments as it provides direct insights into the timing of sediment deposition over time scales of decades (Ballerini *et al*., 2003) to several hundred thousand years (Bateman *et al*., 2004). Jacobs (2008) provides an excellent review of the use of this technique in coastal environments and highlights a number of southern African case studies.

2. Coastal setting

2.1 Climate and marine environments

The southern African coastline lies at the intersection of a number of key elements of the global atmospheric circulation and the oceanic global conveyor system (Lutjeharms *et al*., 2001). As a result, there is considerable variation in terrestrial climate and marine environmental conditions along the approximate 4 000 km of the southern African coastline. Two major ocean currents track the coastline; the Indian Ocean Agulhas Current, and the south Atlantic Benguela Current. The Agulhas Current brings relatively warm waters from the Indian Ocean and flows southwest, closely tracking the continental shelf break in KwaZulu-Natal Province of South Africa, before moving away from the coastline around the Agulhas Bank. Currently velocities at the seabed are sufficient to form submarine dune systems, notably along the narrow KwaZulu-Natal Shelf (Flemming, 1978).

Along much of the southern African coast the continental shelf varies between 3 and 40 km wide, (Flemming, 1981). This is much narrower than the world average shelf width of 75 km (Shepard, 1963). The continental shelf is relatively narrow (often <50 km) in southern Mozambique and KwaZulu-Natal, but widens progressively westwards along the southern Cape coast, which is characterised by the wide and shallow Agulhas Bank (Figure 10.2). Here, water depths of approximately 150 m are found as far as 150 km from the modern coast. On the west coast the continental shelf width is variable, approximating 56 km at Cape Town, but then widening between Cape Columbine and southern Namibia to a maximum of c. 190 km close to the Orange River mouth (Dingle 1973; De Decker, 1988). Further north, it widens again to nearly 100 km in the area of Luderitz. The west coast continental shelf is also relatively deep, with the outer shelf break occurring at depths as great as 400 m (Dingle, 1973). Around the Orange River the wider (ca. 100 km) and shallower (ca. 200 m) continental shelf is thought to reflect high Orange River sedimentation rates during the Cretaceous (Dingle, 1973; Dingle and Hendey, 1984); which then declined substantially from the Late Cretaceous into the Neogene (Dingle and Hendey, 1984). The west coast shelf is also cut obliquely by the Cape Canyon, the head of which lies close to Cape Columbine. This is thought to have been cut by the proto-Olifants River during the Oligocene (Dingle and Hendey, 1984).

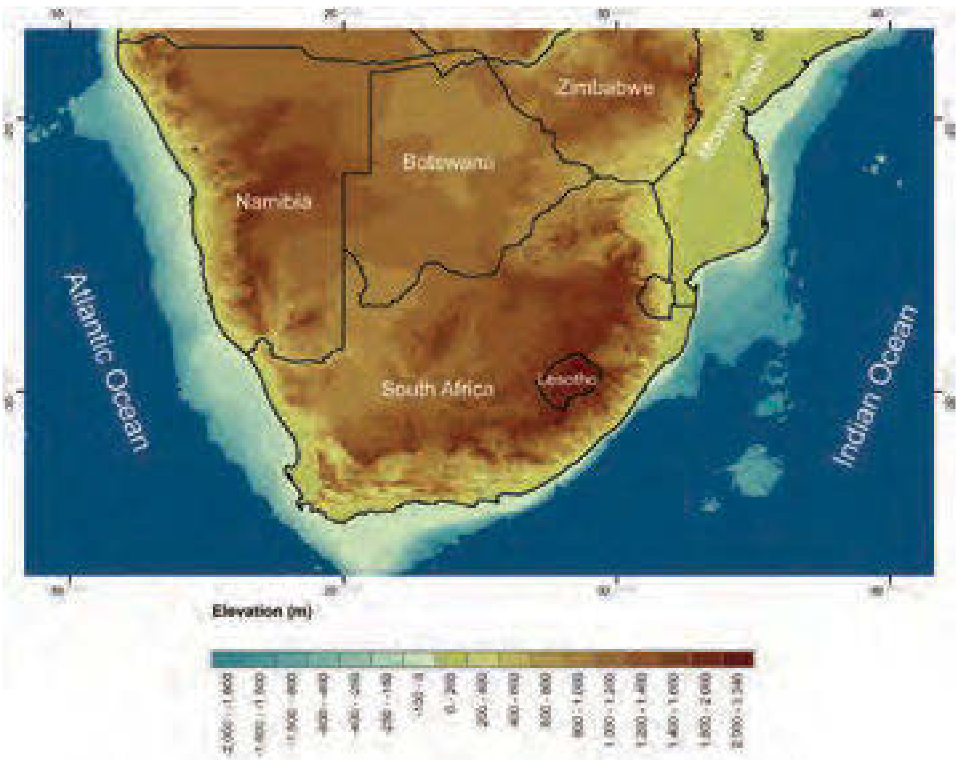

Figure 10.2(a). Topographic map of southern Africa, with continental shelf topography shown. Topographic and bathymetric data were derived from the ETOPO v2.2 dataset (US Department of Commerce, 2006).

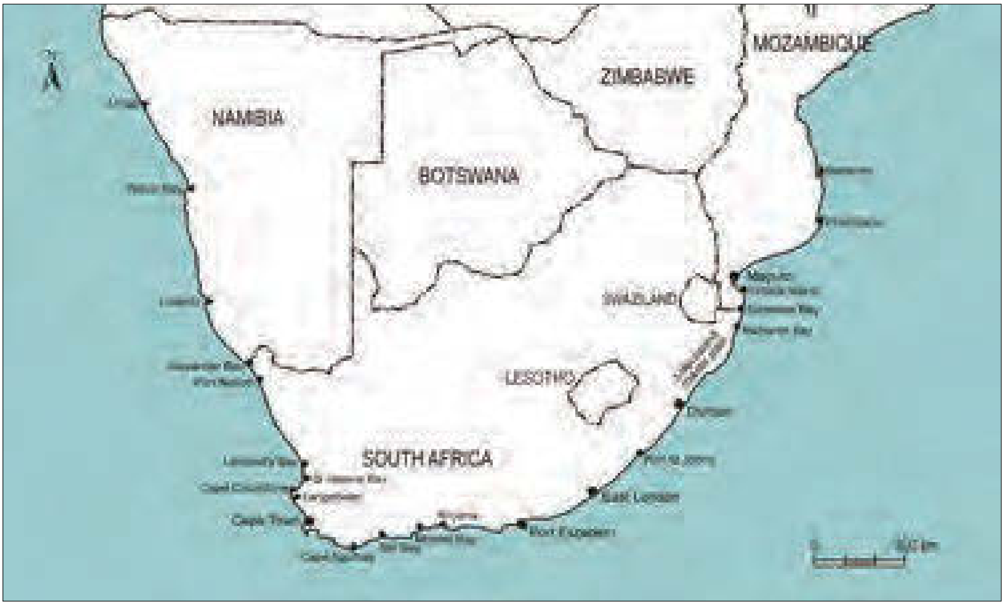

Figure 10.2(b). Locations discussed in the text.

The regional climate is dominated by the seasonal migration of the South Atlantic and South Indian Ocean subtropical high-pressure cells. The former is particularly important on the west coast, where strong south and southeast summer winds drive the Benguela upwelling cells, accounting for the region's high marine productivity and reinforcing atmospheric stability. West coast sea surface temperatures are therefore distinctly cooler than the southern and eastern coastlines. During the austral winter the subtropical high pressure systems move north and the temperate westerlies intersect the southwestern parts of the sub-continent, generating the winter rainfall zone of the Western Cape Province of South Africa and potentially driving localised south coast upwelling zones (Schumann *et al*., 1995).

The tidal range around much of the coastline is generally low and semi-diurnal, reflecting the subcontinent's position in relation to the Atlantic and Indian Ocean amphidromic systems and minimal alteration of tidal waves over a narrow and/or sufficiently deep continental shelf (Davies, 1980). The tidal range is therefore classed as microtidal (<2 m spring tidal range) to mesotidal (2-4 m spring tide range; Davies, 1980), with spring tidal ranges for much of the coastline in the region of 1.8 to 2.0 m and neap tidal ranges approximately 0.6 to 0.8 m (e.g. Cooper, 2001; De Decker, 1988). A notable exception is the macrotidal (>4 m tidal range) Mozambique Channel, where tidal ranges are up to 6.5 m (Tinley, 1985). The major wave direction on the southern African coastline is primarily from the southwest (Heydorn and Tinley, 1980; Davies, 1980), and in KwaZulu-Natal, from the south (Smith *et al*., 2010). The associated long fetch means that the coastline is dominated by swell waves. The south coast of South Africa generally experiences the highest open water wave heights (Cooper, 2001), with median wave heights east of Knysna approximately 2.5 m (Whitfield *et al*., 1983). On the west coast median wave heights progressively decline northwards from Cape Town to Namibia (De Decker, 1988), dropping from c. 2.0 m (median) at the Olifants River mouth (Morant, 1984) to 1.5-1.9 m south of the Orange River (De Decker, 1988). Storm events produce substantially higher waves, with heights in excess of 5 m regularly reached year-round on the west coast (De Decker, 1988). The March 2007 event on the KwaZulu-Natal coastline produced offshore maximum wave heights of 14 m, with wave run-up inundation at the shore reaching 10-11 m above chart datum (Smith *et al.*, 2010).

The predominant swell directions mean that average littoral drift is to the north on the west coast and to the east/northeast on the east coast of South Africa (Davies, 1980). Littoral sand drift at Richards Bay in KwaZulu-Natal is of the order 0.8 x 10^6 m^3 yr^{-1} (Schoonees, 2000), and is even larger (1.4 x 10^6 m^3 yr^{-1}) north of the Orange River (Swart 1984, cited in De Decker, 1988). Data from Algoa (Nelson Mandela) Bay suggest that volumes of sediment of the order 5-9 x 10^5 m^3 yr^{-1} may be moved by long-shore processes within individual embayments (Mclachlan, 1988). The significant fluxes in the Orange River region reflect the substantial volumes of sediment supplied by the river itself, coupled with a consistent southwesterly swell and high energy wave conditions. These are capable of transporting sediments at water depths of 30 m even under "average" wave conditions (De Decker, 1988). In the same area, a weak seasonal poleward current also transports silts and clays southwards of Orange River, forming a distinct Holocene mudbelt (Meadows *et al.*, 1997; Meadows *et al.*, 2002; Rogers and Rau, 2006; Herbert and Compton, 2007).

Given the high wave energy along much of the open coastline and the relatively narrow tidal range, southern African can be classed as a wave-dominated coastline (Davies, 1980; Davies and Hayes 1984). As such, with the exception of rocky coastlines and protected embayments, the coastal geomorphology is dominated by barrier landforms; that is wave-deposited sediments and associated landforms (Davis and Hayes, 1984; Roy *et al.*, 1994). This wave energy is concentrated over a relatively narrow vertical range and can be expected to generate a characteristic suite of landforms (e.g. Pethick, 1984; Davies and Hayes, 1984). Localised exceptions, such as the northern coast of Mozambique, produce landforms more characteristic of tidally-dominated coastlines (environments in which tidal currents are the primary sediment transport mechanism), including extensive mud flats and mangrove systems (Tinley, 1971).

2.2 Geological setting

The southern African coastline has been fundamentally shaped by the processes operating during and subsequent to the break-up of Gondwana approximately 180-115 million years ago (Watkeys, 2006). There has been much emphasis on landscape evolution subsequent to this breakup, with the sub-continent experiencing an essentially erosional history punctuated by phases of landscape rejuvenation associated with proposed Neogene uplift events (Partridge and Maud, 1987; Maud, this volume). The fragmentation of Gondwana was thus a fundamental control on the southern African coastal environment, creating a narrow platform seaward of the Great Escarpment, which can be traced around much of the continental margin (Partridge and Maud, 1987; Figure 10.2). Subsequently, this surface was subject to phases of planation, with extensive sub-aerial weathering (e.g. Summerfield, 1983) accompanied by periodic inundation of the continental margin (Le Roux, 1990; Roberts and Brink, 2002; Roberts *et al.*, 2011).

The Great Escarpment delineates a clear coastal plain on the west coast, which is up to 60 km wide. The southern part of the plain, notably between Cape Town and Lamberts Bay is known as the *Sandveld*, and is veneered by Cenozoic sedimentary deposits, the youngest of which include active dunes and stabilised Quaternary dune systems (Tankard, 1976; Chase and Thomas, 2006; Roberts *et al.*, 2011). The distinct headland at Cape Columbine is created by the presence of relatively resistant Precambrian Cape Granite Suite rocks and defines the southern limit of the log-spiral Saint Helena Bay. Further north into Namibia the coastal plain narrows, but it can be traced through to northern Namibia, where it forms the surface upon which the Skeleton Coast dune systems are deposited (Lancaster, 1982). The mantle of offshore terrigenous sediment on the west coast has been predominantly supplied by the Orange, Olifants and Berg Rivers (Dingle and Hendey, 1984).

On the south and southeast coasts, the coastal platform was cut primarily into Ordovician Table Mountain Group (TMG) sandstones, with a series of half-grabens, which formed during the fragmentation of Gondwana dividing the continental margin into distinct basins (Broad *et al*., 2006). This structural control created a coast with a distinct series of resistant TMG quartzite headlands, alternating with characteristic log-spiral bays, which are associated with post-Gondwana sedimentary infills (Marker and Holmes, 2005; Marker and Holmes, 2010). This geological control strongly influences contemporary along-shore wave energy gradients and sediment transport fluxes. The graben fills were eroded throughout the Cenozoic and formed loci of deposition for Neogene and Quaternary aeolian and marginal-marine sedimentary deposits. On the south coast these Cenozoic marine sedimentary sequences are mapped as the Bredasdorp Group (Malan, 1989; Malan, 1990). On the west coast, the equivalent-aged succession is mapped as the Sandveld Group (Rogers, 1982). The correlated deposits of the Algoa Group are found in the Eastern Cape (Le Roux, 1989). The range of elevations preserving Mio-Pliocene beach deposits bears testament to differential Neogene uplift around the South African coast. Littoral marine deposits may be found up to 90 m above modern sea-level (a.m.s.l.) on the west coast (Rogers 1982; Pether, 1986; Roberts and Brink, 2002), while boulder beach remnants are preserved at 400 m a.m.s.l. inland of Port St Johns and up to 170 m a.m.s.l. along the KwaZulu-Natal coast (Maud, 1968).

In KwaZulu-Natal and Mozambique the coastal plain widens progressively northwards, and is as wide as 440 km near to the Save River in Mozambique (Cooper and Pilkey, 2002; Armitage *et al*., 2006). The southern parts of this plain, most notably the Maputaland area between Lake St Lucia and Maputo Bay, have been the subject of considerable study (Wright *et al*., 2000; Botha *et al*., 2003; Porat and Botha, 2008). Like the western and southern coastlines they exhibit a veneer of regressional late Neogene beach and aeolian facies overlain by Quaternary dunes (Maud, 1968; Botha *et al*., 2003). A distinct whaleback ridge defines the inland margin of the palaeocoastline along the KwaZulu-Natal coast, abutting the Lebombo Mountain footslopes (Botha and Porat, 2007; Porat and Botha, 2008).

In general, and despite evidence for Neogene uplift, the southern African coast is, however, considered essentially tectonically stable and has previously been classified as an Afro-trailing edge continental margin (Inman and Nordstrom, 1971; Davies, 1980).

2.3 Sea level

Sea level is an over-arching driver of coastal geomorphic change, and adjustments in sea level can radically alter coastal form and process over a range of timescales. Relatively subtle changes in sea level may reconfigure a coastline sufficiently to adjust local wave energy gradients, long-shore sediment fluxes and the character of *soft* coastal landforms. The southern African coastline exhibits abundant evidence for relative sea-level change from the Cretaceous (Partridge and Maud, 1987) to the late Holocene (Marker and Miller, 1993). These relative sea-level changes have been driven by tectonic (regional uplift), eustatic (changes global ice volume or ocean basin volume) and hydro-isostatic (loading and unloading of the continental shelf) processes (Partridge and Maud 1987; Compton 2001; 2006; 2007; Ramsay and Cooper, 2002).

2.4 Neogene sea level

Evidence for significant Neogene sea-level fluctuations on the west coast of South Africa are well-preserved in Namaqualand where an ascending sequence of wave cut platforms and associated marginal-marine sedimentary successions has been identified (Pether, 1986). Evidence for these high stands can be found at approximately 90 m, 50 m and 30 m above modern sea-level (a.m.s.l.), and they have been associated with middle Miocene, late Pliocene and the early Pleistocene respectively (Pether, 1986; Roberts and Brink, 2002). Their associated sedimentary packages contain varied and distinct molluscan assemblages, correlation of which implies differential Late Neogene continental uplift along the west coast (Roberts and Brink, 2002). The post-mid Miocene marine regression was superimposed on pulsed Neogene uplift, which generally hinders stratigraphic correlations around the coast. On the south coast the De Hoopvlei Formation is thought to represent a late Pliocene sea-level transgression, and now lies up to 120 m above contemporary sea level (Malan, 1990). The Pliocene Nanaga (aeolian) and Alexandria (marine) Formations of the Algoa Group are now found at elevations in excess of 120 m a.m.s.l. (Le Roux, 1990; Maud, 2008).

2.5 Quaternary sea level

Perhaps more relevant to the contemporary coastal landscape have been the cyclical adjustments in global ice volume that have characterised the Quaternary period (the last 2.6 million years). Benthic oxygen isotope data indicate that such changes in ice volume were sufficient to lower sea level by an average of 125-130 m during full glacial periods (e.g. Waelbroeck *et al.*, 2002). Southern Africa represents an important region for the study of Quaternary sea level as it is located beyond the influence of glacio-isostatic (loading) effects associated with the waxing and waning ice sheets. Such "far field" locations provide important constraints on the timing and amplitude of eustatic sea-level change (Fleming *et al.*, 1998; Woodroffe and Horton, 2005). However, tectonic stability during the Quaternary also implies that onshore records of palaeosea-level will inevitably be fragmentary as there will have been great potential for the reworking of marginal-marine sediments during repeated sea-level transgressions. This leaves a terrestrial stratigraphic record biased towards the highest and most recent sea-level highstands (see Murray-Wallace *et al.*, 2001).

Evidence for Quaternary sea-level highstands has been recognised for some time (Martin, 1962; Maud, 1968; Tankard, 1976; Marker, 1987; Hendy and Volman, 1986; Ramsay and Cooper, 2002) and comprises both erosional surfaces (e.g. 15-12 m, 6-8 m, 3 m; Marker, 1987) and sedimentary deposits (raised beach deposits). Recent work has developed numerical age constraints for the latter using luminescence (Jacobs and Roberts, 2009; Carr *et al.*, 2010; Bateman *et al.*, 2011) and uranium series (Ramsay *et al.*, 1993; Ramsay and Cooper, 2002) dating techniques. These have confirmed the presence/preservation of at least two major sea-level highstands during the middle to late Quaternary. These relate to sea levels approximately 10 m and 6-8 m above modern sea levels. The 10 m high stand has been identified on the southern Cape coastline, while evidence at 8-12 m a.m.s.l. has also been reported in the Alexander Bay area (Gresse, 1988). In the former location this has been associated with the long interglacial marine isotope stage (MIS) 11, approximately 400 000 years ago (Roberts *et al.*, 2007; Jacobs *et al.*, 2011). It has long been debated whether widespread evidence for a c. 6-8 m high stand represents the Last Interglacial (MIS 5e; ca. 130-120 ka; e.g. Hendy and Volman, 1986). This has since been confirmed (Jacobs and Roberts, 2009; Carr *et al.*, 2010). On the southern Cape coast, estuarine sediments at 5.5 m a.m.s.l., interpreted as indicative of sea levels 6-8 m above present, have produced luminescence ages relating to the Last Interglacial (Carr *et al.*, 2010; Figure 10.3). *A* site on the Eastern Cape coastline associated with the renowned Nahoon hominid footprint, (Roberts, 2008; Jacobs and Roberts, 2009) reveals shelly beach deposits at 6 m a.m.s.l. that also date to MIS 5e (124.8 $\pm$5.2 ka). These probably correlate with raised beach deposits preserved locally along the KwaZulu-Natal coast, including the Durban Bluff beaches (Figure 10.3), which are described in detail by Cooper and Flores (1991). The stratigraphic evidence at some sites (Jacobs and Roberts, 2009; Carr *et al.*, 2010) is further indicative of sea-level fluctuations *within* the Last Interglacial period (i.e. 130-120 ka), which accords with recent findings from a variety of far-field locations (Hearty *et al.*, 2007). More generally, these data provide useful constraints on, and insights into, long-term coastal landscape evolution. Certainly, such high stands will have significantly altered the configuration and sediment dynamics of all parts of the coastline (Bateman *et al.*, 2011).

Figure 10.3. Last Interglacial estuarine deposits and raised beaches. a) On the eastern shore of the Groot Brak Estuary mouth. Note person to right for scale stood in front of a large sand barrier (right), which at the time (September 2009) fully closed the estuary mouth. To the left are exposures of estuarine and beach berm swash facies, relating to the Last Interglacial sea-level high stand (Carr *et al*., 2010). b) Exposure of trough cross-bedded estuarine facies on the eastern side of the Swartvlei Estuary mouth, Sedgefield (January 2005). These are overlain by heavily root-bioturbated aeolianite (top of section), which post-dates MIS 5e (Carr *et al*., 2010). c) 200 ka (MIS 7) Cave Rock Member aeolianite truncated by MIS 5e Reunion Rocks Member beach facies (Isipingo Formation) at the Reunion Canal section, Durban Bluff.

Evidence for lower (glacial) sea levels has been recognised for some time. Birch *et al*. (1978) and Martin and Flemming (1986) reported submerged dune systems along the southern Cape and KwaZulu-Natal coastlines, which are assumed to have formed during such low stands. Off Sodwana Bay, Ramsay and Cooper (2002) reported beachrock in water depths of –44 m. These usually form in the inter-tidal zone and in this case produced a uranium series age of 117 ± 7 ka (Pta-U487). Submerged cemented dune sands at Aliwal Shoal near Durban are also reported to be Pleistocene in age (Bosman *et al*., 2008). Evidence for sea levels approximately 120-130 m below present at the height of the last ice age (approximately 21 000 years ago), consistent with expectations from marine isotope records, has also been identified. Notably, *in situ* beach deposits have been reported in water depths of approximately 125 m off Sodwana Bay (Green and Uken, 2005).

The associated drop in base level during such low stand periods implies an allied adjustment in the long profiles of river systems draining the coastal hinterland. Hattingh (1996) reported evidence for this in the form of fluvial terrace sequences in the Sundays River system. Seismic data from modern coastal lagoons reveal palaeochannels 35-40 m deep, which were probably incised during the Last Glacial maximum (Birch *et al*., 1978; Maud, 2000; Wright *et al*., 2000). Recently published seismic data reveal the extension of such palaeochannels onto the continental shelf and the presence of multiple generations of subsequent transgressional sediment fills, the most recent of which relate to the post-glacial sea-level transgression (Green, 2009). This transgression beginning approximately 14 500 calendar years ago (e.g. Fairbanks, 1989) would have radically and rapidly altered the world's coastlines and would have reworked considerable volumes of formally sub-aerial sediment on the continental shelf, although as noted above, such material is still preserved in some instances (Birch *et al*., 1978; Bosman *et al*. 2008). A typical response to such a sea-level transgression would have been the re-mobilisation and landward transport of continental shelf sediment as the shoreface translated laterally and re-graded in response to the prevailing wave conditions (Bird 2000; Francheschini and Compton, 2006). Abrupt flooding of estuaries and the creation of back-barrier lagoon systems would have accompanied this.

2.6 Holocene sea level

Evidence for Holocene sea-level fluctuations is preserved at various locations and represented by a variety of sea-level markers (Järitz *et al*., 1977; Marker and Miller, 1993; Miller *et al*., 1993; Miller *et al*., 1995; Ramsay, 1995; Baxter and Meadows, 1999; Wright *et al*., 2000; Compton, 2001; Compton, 2006). Not all evidence provides sea-level index points (i.e. points of known age, elevation, and sea-level tendency). A number, such as peat deposits in estuarine environments are more appropriately considered limiting points. Others, such as beach deposits, are associated with wide *indicative ranges* and may be difficult to differentiate from storm-deposits (Miller *et al*., 1993). A significant and fairly consistent finding is the occurrence of an approximate 2-3 m high stand during the middle Holocene. Evidence from the west coast of South Africa and Namibia indicates a high stand at approximately 7 500-6 500 cal yr BP (Compton, 2001; Compton 2006; 2007). Ramsay (1995) reported a high stand of a comparable magnitude close to the Mozambique border, which occurred around 4 700-3 800 ^{14}C yr BP. This was mostly based on radiocarbon dating of beachrock cements, along with shells and coral preserved within the beachrock. Beachrock (essentially cemented inter-tidal sands) is potentially associated with relatively narrow indicative ranges, although this may not always be the case (Kelletat, 2006). A pothole infill at +2.75 m a.m.s.l. yielded a (calibrated) radiocarbon age 4 475-4 831 cal yr BP (the published un-calibrated age is 4 480 ± 70 ^{14}C yr BP; Ramsay, 1995). Similar pothole fills at 2-3 m a.m.s.l. are found along much of the Maputaland coast. Raised beach ridges around parts of the KwaZulu-Natal coastal lakes and estuaries, described by Orme (1973), also probably formed during this mid Holocene high stand. At Inhaca Island in Mozambique luminescence dating has constrained the age of a palaeotidal flat approximately 3.5 m a.m.s.l. to between 6.0 ±0.3 ka and 3.7 ±0.2 ka (Armitage *et al*., 2006).

A mid-Holocene sea level high stand is an anticipated far-field response to hydro-isostatic loading of the continental shelf and/or ocean-siphoning during the late Holocene (Woodroffe and Horton, 2005). The subsequent late Holocene regression (RSL drop) should reflect to the flow of mantle material to the continental margin from the flooded shelf (Compton, 2006). The *timing* of the high-stand peak on the west coast (ca. 7 000 BP) and the subsequent rapid regression in some records are not predicted and warrant further investigation (Compton, 2006). Overall, it is clear however that the fundamental setting of the contemporary southern African coastline was established by the mi-late Holocene.

3. Estuarine systems

Abrupt flooding of pre-existing valleys to create estuaries and back-barrier lagoon systems was a key impact of the post-glacial sea-level transgression. By definition, estuary formation is a function of the partial flooding of a river catchment by the sea, and as such they represent relatively youthful geomorphic systems. The timing of the post-glacial transgression in southern Africa suggests that most estuarine infill deposits are Holocene-aged; although the estuaries were likely shaped by numerous glacio-eustatic cycles during the Pleistocene, with some infill deposits dating to around 30 ka (Maud, 2000). Estuarine geomorphic systems are highly complex and driven by the interaction of the marine tidal prism (the volume of water moving into and out of an estuary during one tidal cycle) and the terrestrial fluvial system. They also provide areas of salt or brackish water, which are sheltered from the dominant swell waves affecting the coastline. Those along the southeast African coast north of East London are sufficiently warm and sheltered for mangrove systems to develop (Tinley, 1971; Ward and Steinke, 1982).

Globally, a wide diversity of estuary types is recognised, which are frequently classified by the extent of fluvial or tidal dominance (e.g. Dalrymple *et al.*, 1992). As an essentially wave-dominated coastline, deltaic landforms are not prominent or permanent geomorphic features on the modern southern African coastline. A notable exception is the Orange River, which has been considered an extreme form of wave-dominated delta (Bluck *et al.*, 2007). Although lacking a sub-aerial delta, the Orange River's delta front and pro-delta extend approximately 26 km out to sea and exhibit a marked seaward fining trend as the influence of wave action progressively diminishes (Rogers and Rau, 2006). The sediment load of the river, which may be delivered episodically by major flood events, is rapidly redistributed. The 1988 Orange River flood formed a substantial, but ephemeral delta extending 1.2 km from the shoreline. The flood also deposited substantial volumes of silt and clay on the normally sandy delta front (depth of ca. 40 m, which is above the wave base). This sediment was subsequently reworked by the prevailing wave systems (Bremner *et al.*, 1990). Coarser sediment delivered by the Orange River is transported northwards by the prevailing littoral drift and is a major sediment source for the Namibian shelf (Rogers and Rau, 2006).

Many southern African estuaries are fed by significantly smaller catchments, but they exhibit distinct features that reflect their climatic/wave-dominated setting. Common features include:

i. Confinement within older incised bedrock valleys (rias), which limits their lateral extent; and

ii. A microtidal setting, which limits the tidal prism and onshore and offshore tidal current velocities (Cooper, 2001). Fluvial discharge and sediment delivery are often highly variable, episodic and characterised by low-frequency/high magnitude events (Willis, 1985; Cooper, 1993; Cooper, 2002; Marker, 2004). Given the potential for waves to rapidly rework sediments, ebb tide deltas are rare, and may be rapidly re-worked after forming during major river flood events. Some exceptions, such as the St Lucia Estuary, do however exist (Wright and Mason, 1990).

Cooper (2001) developed a morphodynamic classification system for South African estuaries, which recognises a fundamental division; that of (normally) open and (normally) closed estuaries (Figure 10.4). Open estuaries are generally connected directly to the sea, whereas closed estuaries (e.g. Bot River Estuary; Willis, 1985) are typically cut off by a wave-formed bar (Figure 10.3a), which depending on its extent, may produce a perched estuary. For open estuaries in a wave-dominated environment, morphodynamic variability is generated by the relative significance of fluvial discharge and the magnitude of the tidal prism, creating the sub-classes of river-dominated and tide-dominated estuaries (Cooper, 2001; Figure 10.4). Their distribution is strongly influenced by the characteristics of the river catchment, with the former most notably associated with the short, steep and humid catchments of the KwaZulu-Natal region (Cooper, 1993; Cooper, 2001). Both types exhibit distinct facies arrangements and distinct responses to flood events (Cooper, 2002).

Tide dominated estuaries are characterised by a distinct flood tide delta (tidal prisms of the order 3-6 x 10^5 m^3; Cooper, 2002), which tends to prograde landward under average conditions, producing strong gradients in landward sediment texture within the eustuary (Reddering and Esterhuysen, 1987). This flood tide delta represents a net sink of marine sediment, but may be rapidly scoured and returned to the marine environment during river flood events (Reddering and Esterhuysen, 1987; Cooper, 2002). In river-dominated estuaries the influence of the tidal prism is lower and fluvial discharge/sediment supply deliver fluvial sediment to the mouth of the estuary (Cooper, 1993). This may result in the formation of a braided or one-two channel river system, along with muddy overbank facies and mangrove systems. This system is essentially stable between river flood events, but large floods within a bedrock-confined valley promote the rapid scouring and incision of pre-flood fluvial sediments (Cooper, 1993; Cooper, 2002). Subsequently, fluvial sediments begin to refill the scoured system, moving it back towards its stable state. Interestingly, these South African river-dominated estuary systems do not conform to the standard facies model of river-dominated estuaries proposed by Dalrymple *et al*. (1992), reflecting, in the southern African case, significant fluvial sediment delivery to the full length of the estuary, coupled with a wave-dominated coastline, which prevents seaward progradation (Cooper, 1993).

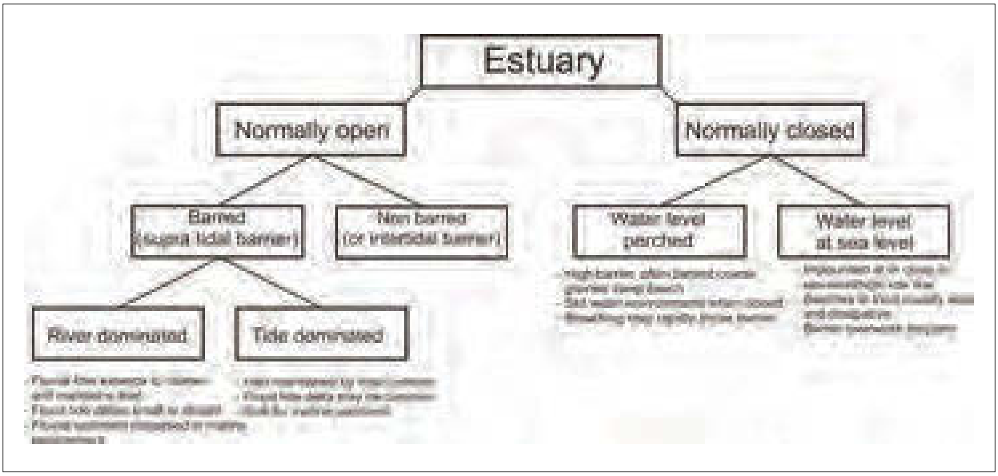

Figure 10.4. Cooper's (2001) morphodynamic estuarine classification system for southern Africa. The closure of estuaries via wave deposited bars forms the primary division within this system.

A detailed suite of studies concerning the closed system of the Bot River Estuary were published by Rogers (1985) Van Heerden (1985) and Willis (1985). Here, the roles of aeolian activity and wash-over processes are paramount in controlling the opening and closure of the estuary and the exchange of water/sediment over decadal timescales (e.g. Van Heerden, 1985). Artificial breaches are usually closed within three-four months (Rogers, 1985; Van Niekerk, 2005).

Along the hyper-arid Namibian coast, seaward draining river systems are highly ephemeral, but display some interesting interactions between fluvial systems, local aeolian dunes and the modern coastline. For instance, the Uniab River catchment (Figure 10.2) is relatively small, but on reaching the sea it forms a large low-gradient alluvial fan (Scheepers and Rust, 1999) or "terminal coastal fan" (Van Zyl and Scheepers, 1992). Its full extent is unknown due to truncation by (presumably) the postglacial-Holocene sea-level transgression, which formed an approximate 14 km long cliff system that now supplies local beaches with gravel-sized sediment. Perhaps most interesting are the hanging channels found at the cliff edge, which are accompanied by clear evidence of knick points in the channel long profiles. It is inferred that despite a regional-scale *rise* in base level during the postglacial transgression, truncation of the fan front has *locally* increased channel gradients, rejuvenating the fan and promoting incision (Scheepers and Rust, 1999). A similar scenario has been envisaged for coastal fans in southeast Spain (Harvey *et al*., 1999). Further north, near to the Hoarusib River, coastal draining river systems are associated with larger catchments and more frequent flows, but are prevented from reaching the coastline as the Skeleton Coast dune-belt progressively widens and increases in height northwards (Krapf *et al*., 2003).

4. Barrier systems

Wave dominated coastlines exhibit a characteristic suite of landforms, the large majority of which are encapsulated by the term *coastal sand barrier*; specifically defined as "elongate shore-parallel sand bodies extending above sea level and consisting of a number of lithofacies including beach, dune, shore-face, inlet and wash-over" (Roy *et al.*, 1994:124). In the southern African context these are abundant features, and are integral parts of some estuarine systems (see above). The term barrier however, encompasses a continuum for forms; from one end-member – that of an attached sandy beach – to another end-member, that of a barrier island/back-barrier lagoon system. The fundamental commonality is the dominance of wave action in delivering and depositing sediment. The positioning of a coastline along this continuum is controlled by:

i. substrate geology – structure, morphology and gradient; and

ii. coastal plan-form; for instance embayed coasts will provide points to attach features like spits (Roy *et al.*, 1994).

The largest proportion of the South African coastline is characterised by attached barriers (i.e. sandy beaches). Rocky cliffed coastlines are characteristic of parts of the Eastern Cape (e.g. Tsitsikamma and Port St Johns; Illgner, 2008).

Over shorter timescales (e.g. glacial/interglacial cycles) the evolution of barrier systems will be governed by sea-level change, wave energy and sediment supply (littoral and long-shore sediment fluxes). The interaction of sea-level change (transgression, stillstand or regression), substrate gradient and sediment supply strongly influence the character of barrier systems and their tendency to prograde or retrograde in the long-term (Roy *et al.*, 1994). Sediment supply becomes increasingly important during still-stand periods.

The southern Cape and KwaZulu-Natal coastlines exhibit multiple generations of superimposed retrogressive barrier systems, which are volumetrically dominated by aeolian sediments. Notable examples are found at Still Bay (Roberts *et al.*, 2008), the Wilderness Embayment (Bateman *et al.*, 2004; Carr *et al.*, 2010; Bateman *et al.*, 2011), and the Maputaland coastal plain (Sudan *et al.*, 2004; Botha and Porat, 2007; Porat and Botha, 2008). Some of these systems are seemingly experiencing net erosion today. Whether this relates to a reduced sediment supply under Late Holocene still-stand conditions, or more recent adjustments in coastal sediment budgets is unclear. The same issue has however been highlighted for Australian barrier systems (Roy *et al.*, 1994).

The Wilderness Embayment is characterised by back-barrier lagoon systems (impounded estuaries; Figure 10.5) and luminescence dating reveals barrier progradation throughout the middle to late Quaternary, which is manifested in the formation of two large, spatially separate barriers (Bateman *et al.*, 2004; Bateman *et al.*, 2011). In Still Bay the barrier systems are volumetrically smaller and back-barrier lagoons are lacking, but similar to Wilderness, the current coastal barrier is dominated by aeolian sediments primarily deposited during and immediately subsequent to the Last Interglacial (Roberts *et al.*, 2008). The KwaZulu-Natal coast exhibits a high, probably Neogene, barrier and localised Pleistocene and Holocene barrier remnants. The barriers are separated at large embayments such as the Natal Bay (Durban), Richards Bay, St Lucia Lakes and the Kosi Lakes. North of the coastline inflection at Mtunzini, the KwaZulu-Natal coastal barrier preserves evidence of polyphase dune accretion that stacked late middle Pleistocene (MIS 11-7), late Pleistocene (MIS 6-3) and Holocene dune systems against a deeply weathered aeolian sand core that probably dates to the Neogene (Sudan *et al.*, 2004, Botha and Porat, 2007; Porat and Botha, 2008). In all cases, relative tectonic stability has resulted in the erosion and reworking of these systems throughout the Quaternary, creating complex stratigraphic records (Roberts *et al.*, 2008; see Murray-Wallace *et al.*, 2001).

Figure 10.5. Spits and barriers. a) A SPOT 2008 image revealing spit/barrier extension across the mouth of the Amatikulu River (centre) and the capture of the Nyoni River (left). This process is driven by the high volumes of longshore sediment delivered from the southeast (left), derived from the Thukela River. b) Spit forming across the Keurbooms Estuary, near Plettenberg Bay. Note the wash-over deposits in the centre of the feature (September 2009). c) Barrier dune systems ("cordons") of the Wilderness embayment viewed looking eastwards from the Map of Africa viewpoint. The seaward barrier (formed primarily during the Last Interglacial and the Holocene) is prominent in this picture, with the main road (N2) running along it. Note the backbarrier lagoon systems of the Wilderness Lakes, Eilandvlei and Langvlei, which are visible in the upper centre of the photo (September 2009).

4.1 Barrier islands and spits

Detached barrier islands and large headland spit systems are relatively uncommon along the southern African coastline. The most notable offshore barrier islands are seen in Mozambique, at Inhaca Island and the Bazaruto Archipelago (Cooper and Pilkey, 2002; Armitage *et al*., 2006; Botha *et al*., 2008). These large and relatively complex systems are dominated by aeolian facies and attain heights more than 120 m above modern sea level. The seaward shorelines are characterised by the widespread occurrence of beachrock, while the landward shorelines exhibit the beginnings of lagoon segmentation driven via local wind waves (e.g. Zenkovich, 1959; Wright *et al*., 2000).

Cooper and Pilkey (2002) hypothesised that their large size was a function of a positive sediment supply throughout the Quaternary. Recent luminescence dating highlights to extent of Holocene dune accretion (Armitage *et al*., 2006). The barrier islands probably developed initially as spits. These grew with associated aeolian activity during major high stands (e.g. MIS 5e), with this material subsequently preserved through rapid lithification, forming the cores of the contemporary islands (*ibid*). The most recent postglacial transgression then drove extensive parabolic dune formation. Barrier island evolution was therefore episodic and focussed around an older core of material saved from marine reworking by post-depositional lithification. Further south, the coastal barrier separating the St Lucia Estuary from the sea initially comprised an embayment with an offshore barrier archipelago similar to Bazaruto during the Last Interglacial period. Later, during glacial low stands this was a dry inland basin, before forming a restricted coastal estuary, isolated along some 60 km of coast by dune accretion during the Last Interglacial and the late Holocene (Wright *et al.*, 2000; Porat and Botha, 2008). The Ponta Congolone Peninsula north of Angoche in northern Mozambique exhibits a similar history.

Mozambique is also a key locale for the occurrence of *fetch limited barrier islands* (Pilkey *et al.*, 2009) of which 337 have been mapped in the Mozambique Channel. These relatively unusual features form in very low energy environments in an absence of swell waves. They are associated with low seabed gradients, an abundance of sediment and frequently form in conjunction with mangroves and salt marshes. They are typically much smaller than open ocean barrier island systems (usually short (<1 km), narrow (<50 m) and low (1-3 m a.m.s.l.)), with storms and over-wash processes playing a relatively larger role in their formation and evolution (Pilkey *et al.*, 2009).

Spit and barrier systems are often associated with estuaries along the South African coastline and may be formed by the same wave-driven sedimentation processes that create estuarine mouth bars (Figure 10.5). The long-shore extension of such features is a function of an oblique angle of swell wave approach and thus long-shore sediment flux (e.g. Ashton *et al.*, 2001). Examples include the prominent 2.5 km long spit across the mouth of the Olifants River on the west coast (Morant, 1984) and a sand spit that forms during low flow periods at the Orange River mouth (Rogers and Rau, 2006).

A number of studies reveal the dynamism of these estuarine spit systems over short (decadal to centennial) timescales. The 1988 Orange River floods caused substantial erosion of salt marsh systems close to the river mouth and formed new spit systems on both the northern and southern shores of the estuary. The spits were initially orientated in a broadly landward-seaward direction, but subsequently their orientation changed as they were reworked by the northwest littoral drift in this area (Bremner *et al.,* 1990). Studies of the Keurbooms Estuary in Plettenberg Bay in the southern Cape (Reddering, 1983) and the Mdloti Estuary in KwaZulu-Natal (Garden and Garland, 2005) have both shown spits forming in the opposite direction to regional long-shore drift trends (which are driven by average swell wave direction). This was thought to be a function of preferential erosion by ebb tide currents (Reddering, 1983) or complex localised interactions of winds and alongshore currents (Garden and Garland, 2005).

On wave-dominated coastlines, such small barrier systems are vulnerable to reworking, particularly via wash-over processes (Figure 10.5), which transfer sediment from the shoreface into the estuary and generate barrier instability (e.g. McBride, 1995; Van Heerden, 1985). Conversely, wash-over events serve to widen a spit, and when accompanied by aeolian sediment transport from the foreshore can stabilise these features over meso-timescales. In the case of the Bot River Estuary, aeolian activity along the coastal barrier was been identified as a key factor influencing the opening and closure of the estuary mouth (Van Heerden, 1985). Additionally, as described above, high magnitude fluvial flood events may completely rework smaller spits associated with estuarine systems (Garden and Garland, 2005; Cooper, 2001; 2002).

Sequences of beach ridges, comprising low, shore-parallel ridges formed by swash and/or aeolian processes are characteristic of prograding coastlines (e.g. Otvos, 2000) and are frequently found in proximity to estuarine systems. Notable examples include the small system at Dwarskersbos in St Helena Bay in the Western Cape (Miller *et al.*, 1993), north of the Thugela River in KwaZulu-Natal (Bosman *et al.,* 2007; Olivier and Garland, 2003) and northeast of the Zambezi River mouth in the Quelimane and Moma-Angoche coastal zones of Mozambique (Tinley, 1971). All are consistent with coastline progradation associated with a net positive sediment budget. The Thugela system also appears to be highly sensitive to sediment supply from the Thugela River over the short term (i.e. inter-annually; Olivier and Garland, 2003).

4.2 Beach processes and beach-aeolian interactions

The coastline of southern Africa is dominated by sandy beaches. This reflects both its geological setting and a ready supply of sandy sediment from offshore environments and/or fluvial systems. Few studies of contemporary beach dynamics in southern Africa have been published. However, it is clear that the high-energy swell wave environment along much of the coastline promotes the formation of dissipative and intermediate surf zone morphologies (Wright and Short, 1984; Hesp *et al.*, 1989). Dissipative surf zones are characterised by wide, relatively low gradient shorefaces with multiple spilling waves breaking over a wide horizontal distance (Figure 10.6). This represents a morphodynamic end-member, adjusted to high wave energies, large wave heights, abundant fine-grained (sand) sediment and microtidal conditions (Wright and Short, 1984). Intermediate forms tend to exhibit greater alongshore-morphological variability and notable secondary morphological features such as bar-trough morphology and rip current systems (Wright and Short, 1984). The latter can occur on the southern Cape coast. The beaches of Namaqualand, Namibia and KwaZulu-Natal tend more often to be associated with lower wave heights (than the south coast) and coarser grained sediment (reflecting geological setting). As such, they are steeper and more commonly exhibit intermediate to reflective beach morphologies (Cooper, 2001). Many of the intermediate beaches are characterised by semi-continuous shore-parallel bars (Smith *et al.*, 2010).

Figure 10.6. An example of a Southern Cape beach, with a typical set of spilling waves, and a wide dissipative surf zone. This example shows the view looking southwest across Buffels Bay in the Wilderness Barrier Dunes, east of Sedgefield.

Some beaches in southern Namibia are rich in gravel reworked and transported up to 150 km northwards from the Orange River mouth. These are an economic source of alluvial diamonds (Spaggiari *et al.*, 2006). Some contemporary beaches in northern Namibia also display many characteristics typical of gravel beach/barrier systems, including sediment sorting by particle shape and size, and the formation of steep berms and prominent beach cusps (*ibid*; Buscombe and Masselink, 2006). These also provide sedimentary analogues for Neogene raised beach deposits, implying long-term morphodynamic stability (Spaggiari *et al.*, 2006).

The morphodynamic state (dissipative and intermediate) of many of beaches along the southern African coast is conducive to significant landward fluxes of aeolian sediment. It was recognised some time ago (Short and Hesp, 1982) that dissipative beach states are highly conducive to aeolian transport when wind conditions are appropriate. This reflects an abundance of fine-medium grained sands (ca. 250 μm) and gentle beach gradients with limited topographic irregularity. The latter maximises aeolian sediment fluxes by minimising the topographic disturbance of wind flow (e.g. flow separation) and allowing the full development of a saltation cascade (Short and Hesp 1982; Sherman and Bauer, 1993). Narrow tidal ranges also maximise the presence of dry sand, providing a ready supply of erodible sediment. The probability of marine erosion (e.g. foredune scarping) is also notably lower on dissipative beaches relative to reflective beaches (Sherman and Bauer, 1993). Hesp *et al*. (1989) argued that in the South African context, these are key factors serving to promote the formation of extensive foredune and coastal dune systems, and particularly the major transgressive (landward migrating) coastal dune systems.

5. Aeolian dune systems

5.1 General distribution and form

A systematic study of South African coastal dune systems was undertaken by Tinley (1985). The mapped distribution of coastal dune systems based on this work is shown in Figure 10.7 (modified from Tinley, 1985) and in South Africa and Namibia as much as 80% of the coastline comprises sandy beaches backed by dunes (Tinley, 1985). This includes the major dune systems of the Namib Sand Sea, which derive much of their sediment from the coastal zone. Recent U-Pb dating of detrital zircons indicates that the Orange River, via the coastal system, is the major sediment source for the dunes of the Namib Sand Sea (Vermeesch *et al*., 2010). On the Skeleton Coast, Lancaster (1982) noted that sand was supplied to this system at specific, appropriately orientated locations along the coastline, with sands derived primarily from the littoral reworking of sediments supplied by ephemeral, coast-draining river channels.

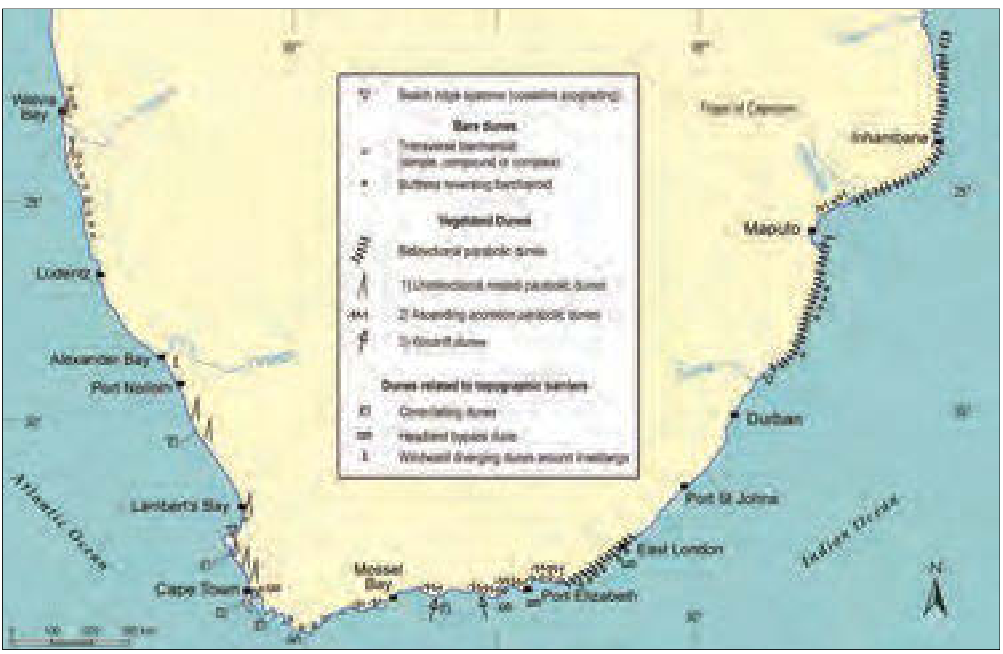

Figure 10.7. Distribution of dune types along the southern African coastline (modified from Tinley, 1985).

The proximity of relict Neogene dune systems and contemporary dunes along large tracts of the southern African coast suggests that the fundamental geographic and geologic controls on dune formation have remained conducive to coastal aeolian transport throughout the last few million years (Tinley, 1985; Rust *et al*., 1990; Franceschini and Compton, 2006). The occurrence of coastal dunes is driven by a number of key (inter-related) factors (e.g. Tinley, 1985; Hesp, 2002):

- wind regime (strength, direction, and orientation relative to the coast line)
- sediment supply (controlled by longshore drift, marine currents, wave action and proximity to river mouths)
- beach form
- coastal configuration and beach orientation
- rainfall (which controls vegetation cover and/or prevents aeolian transport)
- vegetation.

Assuming an environment generally conducive to dune formation, variability in these factors will impart variability in dune morphology. Tinley (1985) and subsequently, Illenberger and Burkinshaw (2008) developed coastal dune classification systems for southern Africa (Figure 10.8). The two major categories of dune comprise fixed dunes and mobile dunes. Fixed dunes, which do not migrate, form in the presence of persistent vegetation. This effectively traps saltating sand grains, leading to the formation of hummock dunes (akin to incipient foredunes; Hesp, 2002) and foredune systems, both of which are prone to erosion by wind and wave action over short timescales (Figure 10.9). More permanent features are *retention ridges*, which are shore-parallel features, typically beyond the reach of erosive storm waves. These may grow to in excess of 10 m and can be continuous for several kilometres (Illenberger and Burkinshaw, 2008).

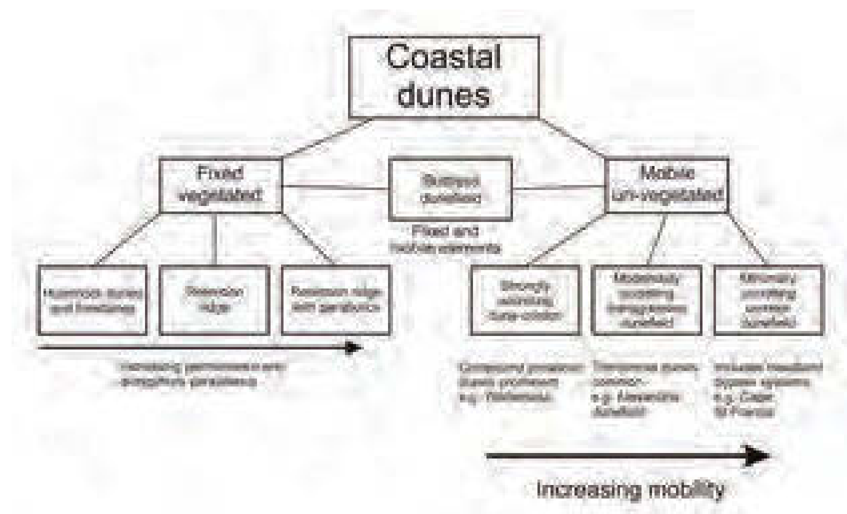

Figure 10.8. Illenberger and Burkinshaw's (2008) morphodynamic coastal dune system classification. The role of vegetation is paramount in providing a first order division of dune types (modified from Illenberger and Burkinshaw, 2008).

Figure 10.9. Hummock dunes at Elands Bay, looking south towards Cape Deseada (September 2010). Pioneer vegetation plays a key role in trapping sand and the dunes may be rapidly destroyed by the death of the vegetation or storm events.

5.2 Mobile dunes

Mobile dunes have a propensity to migrate landward. Large systems of such dunes are usually referred to as transgressive dunefields (Hesp *et al*., 1989). Apart from hyper-arid environments (e.g. Namibia) these systems usually contain some areas of vegetated sand, but high disturbance frequencies and potentially rapid burial by moving sand mean that this vegetation is unable to prevent the net landward transgression of the dune system. Within this over-arching class of dunes variability in sediment supply, climate and disturbance events influence dune form, vegetation cover and the temporal dynamics of mobile dune systems (Illenberger, 1988; Hesp, 2002). Illenberger and Burkinshaw (2008) differentiate mobile dunes on the basis of their propensity to accrete; that is, to stack vertically. Dunes within relatively humid environments tend to be well-vegetated, preventing the landward ingress of sand, which instead accumulates vertically. As dune mobility increases (due to reduced vegetation cover and/or increase erosivity (wind)) net accumulation is reduced. In minimally accreting dunefields the net sand flux is high, preventing vegetation colonisation and producing characteristic corridors of transverse dunes (Illenberger and Burkinshaw, 2008). More specific examples of this dune type are headland bypass systems, notable exemplars of which are seen at Cape Agulhas, Goukamma/Buffels Bay, Cape St Francis, and Cape Recife (Figure 10.7). These are important components of longshore sediment cells, retaining sands in the near-shore environment and nourishing beaches down-drift of the headlands. (De)Stabilisation of these systems has important implications for beach sediment supply and coastal management (Hellström, 1996; La Cock and Burkinshaw, 1996; Holmes and Luger, 1996).

Parabolic dunes have long been recognised as a key component of coastal dune systems (Pye, 1982; 1983), and are found in a variety of environments (Tinley, 1985). Parabolic dunes are essentially U-shaped features, characterised by a leeward nose of active sand transport, sourced from a distinct upwind deflation hollow. The nose migrates slowly downwind, leaving two trailing arms, which are vegetated and stabilised to varying degrees. As migration takes place from the nose parabolic dunes have a tendency to extend in the direction of transport, lengthening the trailing arms. Initiation does not require significant aridity or extensive areas of bare sand, and they form in a wide variety of coastal settings, including humid tropical (Pye, 1982) and temperate (Bailey and Bristow, 2004) environments. A common observation is that they evolve from an initial localised blowout feature (which may reflect the localised destabilisation of vegetation). These events may occur on foredunes or retention ridge systems, or within previously stabilised dune systems (Hesp, 2002; Walsh 1968). Excavation and extension of such a hollow begins as winds are accelerated through the blowout topography (Hesp and Hyde, 1996); an excellent example of morphodynamic feedback.

Three basic types of parabolic dune are commonly identified in southern Africa: parabolic, hairpin parabolic and wind rift parabolic. In regions where multiple generations of parabolic dunes have coalesced or over-ridden one another, accretion ascending or imbricate parabolic types form (Tinley, 1985; Illenberger and Burkinshaw, 2008). The high composite coastal barrier along the Maputaland coast is a good example of successively accreted transverse ridges formed from complex, ascending parabolic dunes. Hairpin or elongate parabolic dunes are indicative of the considerable landward migration of the active nose, leaving lengthy parallel trailing arms. They typically have a length to width ratio in excess of three (Pye, 1982) and may extend considerable distances landward from their initial source. Notable examples are seen in False Bay (Roberts *et al.*, 2009) and at Yzerfontein on the west coast (Tinley, 1985). Wind rift dunes comprise only the two parallel trailing arms, reflecting the erosion or breaching of the dune nose (Pye, 1982; Botha *et al.*, 2003; Illenberger and Burkinshaw, 2008). Good examples are seen at Cape Recife (Tinley, 1985) and the Maputaland coastal plain (Botha *et al.*, 2003).

Buttress dunes represent an intermediate between fixed and mobile dune systems (Figure 10.10) and are most common where winds blow parallel to the coastline. They consist of seaward-perpendicular orientated barchanoid or transverse dunes (mobile element), attached on their landward edge to a fixed and vegetated retention ridge (Illenberger and Burkinshaw, 2008). Along the northern KwaZulu-Natal coast narrow zones of buttress dunes are inclined against a pre-existing dune cordon (Tinley, 1985). Here, reversing transverse dunes lie behind the irregular foredunes, with their inland edges stabilised by a thickly vegetated coastal barrier dune (Figure 10.10). The White Sands Dunefield north of Sodwana Bay comprises transverse or buttress dunes that feed localised parabolic extensions, which ascend the coastal barrier under the influence of southerly winds blowing across the embayment.

Figure 10.10. Oblique aerial photographs of buttress dunes inclined against a pre-existing dune cordon, Maputaland, northern KwaZulu-Natal Province of South Africa. Right: Nkonyane Dune (118 m) on the vegetated coastal barrier dune that isolates coastal Lake Bhangazi, 11 km south of Sodwana Bay. Left: Buttress or reversing transverse dunes on the coastal barrier ~17 km south of Sodwana Bay.

Barchan dunes are mobile, crescentic dunes characterised by horns pointing in the downwind direction (e.g. Bourke and Goudie, 2007). They are typically associated with unimodal, high-energy wind regimes and are most abundant in the hyper-arid coastal environments of Namibia. Notable examples are found within the Skeleton Coast Erg (Lancaster, 1982; Krapf *et al.*, 2003) and on the Kuiseb Delta near Walvis Bay (Barnes, 2001; Bourke and Goudie, 2007). Localised barchan dunes are also found in the Schelmhoek dune system in the Eastern Cape (Illenberger, 1988).

5.3 Transgressive coastal dune systems

Wind regimes are bimodal or multi-modal along almost all southern African coastlines (S/NW on the west coast, W/E on the south coast, SW/NW on the east coast), but in most instances coastal dunefields display net migration/transgression in a single direction (Tinley, 1985). On the west coast, dune systems are primarily influenced by southerly winds coincident with dry summer months when vegetation is often sparse and easily overwhelmed and peak wind speeds are higher. These combine to produce characteristic plumes, which comprise a confined, wind-aligned dunefield that may transgress considerable distances landward. North of Cape Town they are fed by beaches formed in pre-existing topographic lows in the basement geology and/or in close proximity to river mouths (Roberts *et al.*, 2009). Notable examples include the Atlantis dune plume north of Cape Town (Roberts *et al.*, 2009), the Yzerfontein/Geelbek plume (Kandel *et al*, 2003; Francheschini and Compton, 2006), the Bitter-Spoeg plume north of Lamberts Bay, and the Swartlintjies River plume (Tankard and Rogers, 1978; Tinley 1985). Currently, landward migration of the Yzerfontein plume is of the order of 8-10 m a^{-1} (Kandel *et al.*, 2003), whereas over the late Holocene as a whole migration rates have been estimated to be in the region of 5 m a^{-1} (Francheschini and Compton, 2006). Individual dunes within the plumes may comprise hairpin parabolic (Tinley, 1985; Barwis and Tankard, 1983; Roberts *et al.*, 2009), along with barchanoid and transverse dunes (Francheschini and Compton, 2006; Fuchs *et al.*, 2008).

On the south coast major dune systems are primarily associated with the aforementioned log-spiral bays (Tinley, 1985), more humid climatic conditions (hence, greater vegetation cover) and a greater propensity for vertical accretion. As a result, multiple generations of accreting/imbricate parabolic dunes have created shore parallel cordon/barrier features. The orientation of modern parabolic dunes and analysis of fore-set dips preserved in Quaternary dunes reveals consistent westerly to northwesterly formative winds (Roberts *et al.*, 2008), which are associated with winter cyclonic systems. The implication is that wind strength, rather than seasonal aridity is the key factor mobilising many coastal dune systems along the southern Cape coast (see also Carr *et al.*, 2006). The dune systems of Maputaland and southern Mozambique coastal plain comprise extended (hairpin) parabolic and windrift parabolic dunes, which are also indicative of an essentially uni-modal dune-forming wind regime (Botha *et al.*, 2003; Porat and Botha, 2008). The dune systems display a generally north-south orientation across the coastal plain.

The largest active coastal dune system in southern Africa aside from the Namib Sand Sea is the Alexandria Dunefield in Algoa Bay, east of Port Elizabeth. This system is 2-3 km wide, more than 50 km in length, and is classified as an accretionary sheet dunefield (Illenberger, 1988; Illenberger and Burkinshaw, 2008). The Alexandria system is actively transgressing landward and is described by Hesp *et al.* (1989) as a tabular transgressive dunefield. The landward limit of the system is marked by an approximate 40 m high precipitation ridge, which is partially vegetated, but migrates landward at approximately 0.25 m a^{-1} (Illenberger and Rust, 1988; Illenberger and Burkinshaw, 2008). The dunefield is strongly influenced by a multi-modal wind regime and comprises a range of dune types, the most prominent of which are seasonally-reversing transverse dunes (Illenberger, 1988; Hesp *et al.*, 1989). These may be straight or sinuous crested and form oblique or transverse to the prevailing winds, seasonally adjusting their cross sectional form in response to changing wind directions. Smaller, 6-7 m high transverse dunes are found on the beach above the high tide line, which form during strong winds or storm conditions (Hesp *et al.*, 1989).

5.4 Coastal aeolian systems: long-term evolution

The Sandveld, Bredasdorp and Algoa Groups are dominated by calcareous aeolian dune sediments deposited throughout the Pliocene and Quaternary. In the Western Cape in particular, once stabilised these dunes frequently cement to form dunerock, or aeolianite (Marker, 1977; McLaren, 2007). Palaeosols (Figures 10.11) and sedimentary structures, including large-scale dune fore-sets (Figure 10.12) and prominent bounding surfaces are particularly well-preserved within these carbonate-rich aeolianites. The propensity for cementation is driven by the high marine-derived carbonate content of the source beach sands (Siesser 1970; Dingle *et al.*, 1987), coupled with a highly seasonal and semi-arid climatic setting, which promotes the dissolution and re-precipitation of carbonate minerals. Cementation processes may occur rapidly in some environments, but the general absence of cementation within Holocene dunes implies that the process broadly operates over timescales of 10^4-10^5 years. Within 1-2 m of the ground surface distinct patterns of pedogenic alteration are observed, including the homogenisation and destruction of pre-existing bedding, the intense formation of rhizoliths (root casts) that may coalesce through time, and the formation of prominent laminated calcretes (Figure 10.11**;** Knox, 1977; Roberts *et al.*, 2008). Palaeosols and calcretes exposed within these relict dune systems show large lateral variations in development and are strongly influenced by pre-existing dune topography, which alters water movement through the soil profile.

Figure 10.11. a) A well-developed palaeosol formed within aeolianites east of Still Bay. This surface formed between approximately 90 ka and the early Holocene (Roberts *et al.*, 2008). Note homogenisation/destruction of pre-existing bedding, rubification and formation of a clear laminated "hard pan" calcrete near to surface (January 2005); b) Heavily root bioturbated aeolianite at the mouth of the Swart Estuary, Sedgefield. (September 2004). This is the same section as that described in Carr *et al.* (2010).

Figure 10.12. Dune bedding structures within carbonate-rich aeolianites. a) Large-scale northeastly dipping fore-set bedding within aeolianites east of Still Bay (see Roberts *et al*., 2008 for a more detailed description). These overly a low-angle laminated facies next to the person, which contain elephant trackways dating to approximately 130 ka (January 2005). b) Prominent aeolian fore-set bedding of the Langebaan Formation preserved above lagoonal sediments, western shoreline of Langebaan Lagoon, Western Cape. These date broadly to the Last Interglacial period and are associated with hominid footprints (Roberts and Berger, 1997). The site shown here is close to Die Preekstoel (September 2009).

The Neogene to Holocene Maputaland Group on the northern KwaZulu-Natal coastal plain is dominated by non-calcareous dune deposits. Here reworking of older weathered dune deposits and a more humid environment, which increases chemical weathering rates, combine to produce dunes devoid of carbonate minerals. Calcareous dune deposits are associated with raised strandlines and dunes accreted against the coastal barrier dune core. Here, weathering is associated with *in situ* formation of secondary clay minerals, illuviation of clays and intense rubification (Botha and Porat, 2007). In humid areas of the plain near to the coast, weathering occurs with such rapidity that dunes differing in numerical age by just 1 500 years can be differentiated on the basis of soil development indices (*ibid*).

Until recently there has been limited capacity to assess the age of the southern African coastal dune systems. Palaeosol features clearly imply episodic formation and there has long been speculation as to driver(s) of this episodicity (Malan, 1990, Barwis and Tankard, 1983; Illenberger, 1996). The application of radiocarbon dating is limited by a lack of preserved soil organic matter, as well as the upper age limits of the technique. Radiocarbon ages derived from the carbonate mineral components of Holocene dune sands provide only maximum ages for the associated landforms, but interesting insights into the age of the biogenic carbonate fraction of the dune sediment (Deacon, 1966; Illenberger and Verhagen, 1990; Compton and Franceschini, 2005; Francheschini and Compton, 2006; Roberts, 2008). Thus, although not ideal for constraining the age of specific landforms, the approach provides clues as to the age and provenance of the sediments. Bulk carbonate radiocarbon ages for pioneer dunes and modern beach sediments on the west coast (16 Mile Beach) and southeast coast (Algoa Bay) have produced radiocarbon ages in the range 10-20 000 cal yr BP, and are substantially older than the expected age of the dune landforms (Illenberger and Verhagen, 1990; Franceschini and Compton, 2006). Similar insights are provided by whole-rock amino acid racemisation (AAR) analyses. Like radiocarbon dating this technique is indicative of the time elapsed since the carbonate-secreting organism's death (Murray-Wallace *et al*., 2001). When coupled with numerical (luminescence) ages for aeolianites near to Still Bay, AAR data revealed sediment ages markedly in excess of the landform ages, indicating the repeated deposition and reworking of coastal carbonate sediments throughout the middle to late Quaternary (Roberts *et al*., 2008).

Luminescence dating has been increasingly applied to the South African coastal aeolian systems and provides direct insights into the time elapsed since the sediments within the sampled landform were last transported (Roberts and Berger, 1997; Vogel *et al*., 1999; Shaw *et al*., 2001; Bateman *et al*., 2004; Sudan *et al*., 2004; Armitage *et al*., 2006; Carr *et al*., 2007; Porat and Botha, 2008; Fuchs *et al*., 2008; Roberts *et al*., 2008; 2009; Carr *et al*., 2010; Bateman *et al*., 2011; Roberts *et al*., 2011). Two study areas exemplify the application of this approach and illustrate the long-term responses of coastal aeolian systems in differing climatic and oceanographic settings.

5.5 Long-term evolution of the southern Cape aeolian systems

On the southern Cape major loci for the accretion of barrier dune and aeolianite systems broadly correspond to the mapped distribution of the Pleistocene Waenhuiskrans Formation (Malan, 1990). Notable examples are found around Cape Agulhas, east of Still Bay and within the Wilderness embayment. In recent years, more than 80 optical luminescence ages have been published for aeolianites along this stretch of coastline (Figure 10.13). The record reveals the general correspondence of aeolian activity and eustatic sea levels throughout the last 200 000 years, particularly the Last Glacial-Interglacial cycle, which most OSL ages relate to. For this period aeolian activity has been both episodic and most likely driven by the supply of sediment associated with a proximal shoreline during sea-level highstands (Illenberger, 1996). With the exception of headland bypass systems, landward migration has been limited relative to vertical accretion (Roberts *et al*., 2009). Imbricate/stacked parabolic dunes have coalesced to form larger, shore-parallel barrier structures (Roberts *et al*., 2008), which are most obviously manifested in the dune cordons of the Wilderness embayment. Illenberger (1996) argued that the volume of the Wilderness barriers implied that they must have formed during multiple interglacial high stands. Luminescence dating confirms this, revealing ages spanning the last 205 ka and demonstrating that the seaward (youngest) barrier comprises Holocene and MIS 5e – MIS 5a sediments (130-70 ka). Evidence that aeolian activity tracked the coastline southwards during sea-level regressions is found in 14 sub-parallel dune ridges at depths of –40, –50-55, –65-70 and –80-90 m (Martin and Flemming, 1986). Analysis of the timing of seaward barrier accretion in relation to bathymetric data also demonstrates the role of offshore topography in controlling coastal responses to palaeo-sea-level change. Most notably, the occurrence and duration of phases of MIS 5e and post- MIS 5e aeolian activity preserved at specific locations (Bateman *et al*., 2011).

The Wilderness Lakes formed during the post-glacial marine transgression, although their sedimentation history remains to be investigated in detail. On similarly wave-dominated, microtidal coasts such as New South Wales (Australia), relatively thick bodies of marine sediment were rapidly emplaced in back-barrier lagoons during this flooding stage (Sloss *et al*., 2006). Unpublished radiocarbon and luminescence ages from Groenvlei, near to Sedgefield reveal the rapid accumulation of sandy marine-derived sediments around 8 000 years ago, presumably prior to the lake's isolation from the sea by dune systems (Brian Chase, pers comm. 2010).

More limited studies of southern Cape headland bypass systems have been undertaken. A notable example is the system at Cape Agulhas, where luminescence ages reveal that the Holocene system was active at 5.8-5.1 ka and 4.2-4.0 ka, perhaps associated with the stabilisation of sea level in the mid Holocene (Carr *et al*., 2006; Bateman *et al*., 2008). The Holocene dunes unconformably overlie far older aeolianite, implying the existence of a comparable palaeo-by-pass system. The aeolianites date to 160-180 ka close to the surface (Bateman *et al*., 2004), but material in excess of >400 ka has also been identified (unpublished data).

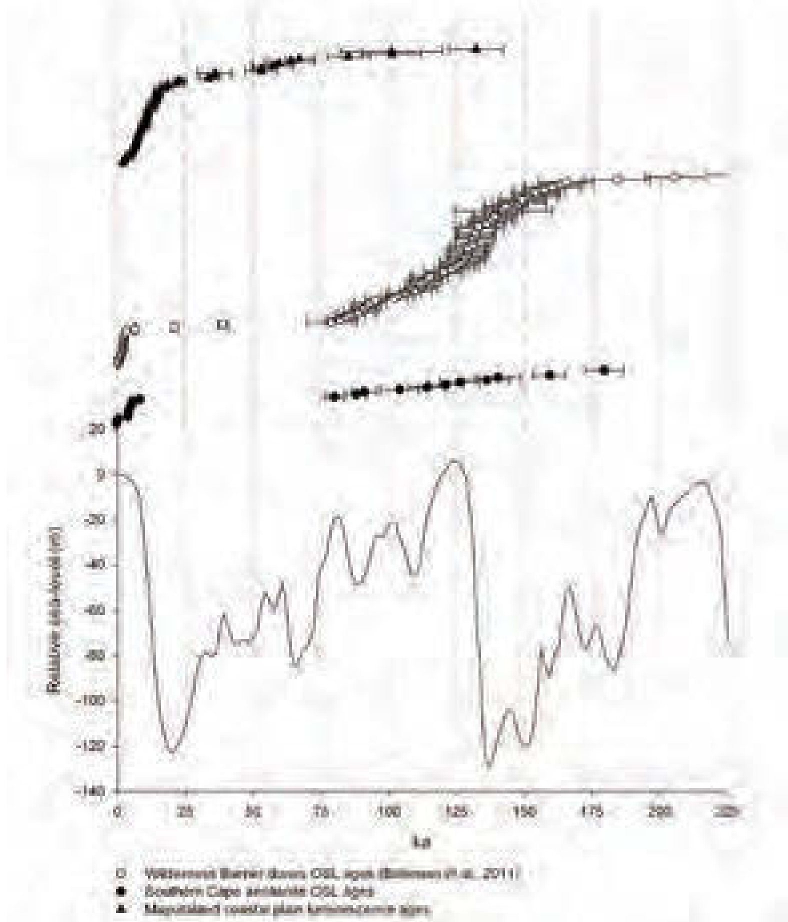

Figure 10.13. The timing of dune activity/reactivation, as derived from luminescence dating techniques, along the Maputaland coastal plain and southern Cape coastlines. The ages presented comprise: a) Data from Porat and Botha (2008) for Maputaland (filled triangles); b) Data for aeolianites and coastal dunes on the southern Cape coastline (synthesised from Jacobs *et al*., 2003; Bateman *et al*., 2004; Bateman *et al*., 2008; Roberts *et al*., 2008; filled circles). c) Ages from the Wilderness Barrier dune system (Bateman *et al*., 2011 and references therein; open circles). These are shown in relation to the relative (eustatic) sea-level record for the same period (after Waelbroeck *et al*., 2002).

5.6 Long-term evolution of the Maputaland aeolian systems

Geochronological work on the Maputaland coastal plain reveals a record of aeolian activity characterised by far greater spatial complexity in the timing of active dune accretion, and by the great antiquity of some relict dune forms (Porat and Botha, 2008). A key finding is the differentiation of aeolian activity directly associated with the coastal barrier dune system from dune activity associated with the wider coastal plain, which extends a further 70 km inland. The degraded coastal plain dune landscape comprises late middle Pleistocene to late Pleistocene Kosi Bay Formation aeolian deposits. These have been reworked and largely buried by pulses of parabolic dune migration and also form the core of the (Isipingo Formation) coastal barrier. The latter was mantled with calcareous parabolic dunes during the Last Interglacial. Like the southern Cape, the barrier aeolianites extend below contemporary sea level and coastal aeolian activity occurred periodically until 67-60 ka, only recommencing in the early Holocene (Porat and Botha 2008).

In contrast, the coastal plain dunes (KwaMbonambi Formation) reveal phases of mobilisation throughout MIS 4 and 3 (57-53 ka and 34-36 ka), separated by episodes of more humid and stable conditions. Ground penetrating radar studies of these parabolic dunes reveal the vertical stacking of trailing limbs, as well as apparent sand accretion on the western side of the dunes, belying their north-south orientation (Botha *et al*., 2003). Distinct bounding surfaces and buried dune topography clearly demonstrate the multi-phase accumulation of individual parabolic dunes (Botha *et al*., 2003). Diatomite deposits accumulated in coastal lakes and interdune wetlands during humid phases within MIS 3, with burial during later periods of sand mobility associated with a drier LGM (23-19 ka). During this period northward migration of parabolic dunes was widespread, and hairpin parabolic dunes migrated from the dry "Lake" Sibaya Basin onto surrounding areas until around 7 ka, when peat deposits imply the re-expansion of interdune wetlands (Grundling, 2004). The role of a rising regional watertable and increased vegetation cover in constraining parabolic dune activity is therefore emphasised, with the Holocene sea-level transgression simultaneously reactivating dune activity on the coastal barrier. Here, at least four pulses of parabolic dune accretion were deposited against the pre-existing barrier during the period 8-2 ka; a process driven by rising sea-levels and destabilised foredune systems (Porat and Botha, 2008).

There are thus clear contrasts in this record compared to the southern Cape. In the latter area aeolian activity was consistently confined to the littoral zone and was driven primarily by relative sea-level change. On the Maputaland Plain the coastal barrier dune record is decoupled from the coastal plain dune systems, with the activity of the latter far more sensitive to climatically-driven activation and reworking.

7. Conclusion

The last twenty years have witnessed great progress in our understanding of southern African coastal geomorphology, most notably in terms of the long-term evolution of the continental margins and the dynamism and evolution of the sub-continent's estuarine/barrier lagoon systems. Aeolian geomorphology has also been a key area of focus, reflecting the globally-significant scale and antiquity of the southern African coastal aeolian systems. As in many branches of geomorphology, there is an increasing treatment of coastal environments as systems, epitomised by the development multiple morphodynamic classification schemes (Cooper, 2001; Illenberger and Burkinshaw, 2008). These recognise the interaction between form and process, and provide genetic (i.e. explanatory) classifications of coastal landform diversity (see also Cowell and Thom, 1994). Long-term studies have also demonstrated the importance of landscape inheritance in affecting geomorphic system responses to Quaternary eustatic sea-level change. Research into the long-term evolution of various barrier dune systems and associated estuaries exemplifies this issue. Key to this understanding has been an ability to develop chronologies of landscape change.

From such data the episodicity/periodicity of geomorphic change is increasingly recognised. The widespread dune remobilisation along large parts of the southern African coastline over both long (glacial-interglacial) and short (late Holocene) timescales highlights the dynamic nature of this high energy coastal zone and implies potential sensitivity to future global climate changes and land-use change (e.g. Hellström, 1996). The recent coastal erosion event in KwaZulu-Natal during March-April 2007 (Smith *et al*., 2010) occurred due to the coincidence of high spring tides (i.e. greater coastal inundation) and large swells, revealing a scenario that could be played out more frequently as global sea-level rises (Theron and Rossouw, 2008). Similarly, recent studies have demonstrated the dynamism and sensitivity of estuarine systems to major terrestrial (fluvial discharge) and marine storm events. The former may be significantly influenced by human activity within the wider river catchment (Marker and Holmes, 2005). An increased understanding of the dynamism of southern African coastal geomorphic systems clearly has implications for the future management of the coastal zone. In the context of this, the review presented above also highlights the need for greater study of southern African coastal geomorphic process at the micro- to mesoscales.

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11

Soil Erosion and Land Degradation

Soil Erosion and Land Degradation

John Boardman, M Timm Hoffman, Peter J Holmes and Giles FS Wiggs

1. Introduction

Concerns about land degradation have long been expressed in southern Africa (Anon., 1923). In the twentieth century the problem was often linked to the perceived decline in vegetation quality and, therefore, the impact on grazing quality and quantity. More recently, concern has focused on loss of biodiversity, invasive species and the provision of adequate supplies of good quality water. Indeed, considerably more data are available for South Africa than for other areas of southern Africa.

South Africa has a long history of research into land degradation and soil erosion. Key themes in the land degradation debate have been summarised by Dean *et al*. (1995), Hoffman *et al*. (1999a), and Hoffman and Ashwell (2001), while a comprehensive review of soil erosion research in South Africa is provided by Garland *et al*. (2000). For the southwestern Cape, Meadows (1998) listed the forms of land degradation as vegetation clearance, invasive alien organisms, soil erosion by wind and water, physical and chemical deterioration of soils, loss of wetlands and riverine ecosystems, surface and groundwater quantity and quality. Many of these concerns were linked by issues of fire, water scarcity, population pressure and climate change. Land degradation, therefore, exists in a context much wider than the concerns about loss of soil *per se*.

This chapter reviews what is known about soil erosion, as this is the most important aspect of land degradation in the region. Inevitably, most focus is on South Africa and the emphasis is on erosion by water. Wind erosion has been little studied though Talbot's pioneering work took in both the water driven erosion of the Swartland and the wind eroded areas of the Sandveld (Talbot, 1947). Fluvial (riverine) and coastal erosion are alluded to elsewhere in this volume (Rowntree; Carr and Botha). Of primary concern are the development of badlands and gullies (dongas), although on susceptible soils piping plays a leading role. The historical picture is important because the role of land use under European style farming systems in the initiation and development of erosion is both contested and crucial.

2. Overview: Pleistocene and Holocene soil erosion

There is considerable evidence in the form of cut and fill sequences for hillslope erosion in the late Pleistocene. This comes largely from work by Botha and colleagues in eastern South Africa. It is supported by radiocarbon, archaeological and thermoluminescence dating of paleosols and calcretes within colluvial sequences. The majority of dated colluvial deposits are from arid stages of the Late Quaternary with deposition continuing into the Holocene at some sites in KwaZulu-Natal Province of South Africa (Botha and Partridge, 1988; Clarke *et al*., 2003). Older deposits seem to have been removed by erosion. Botha *et al*. (1994) make the point that age differences between deposition and burial of the same unit at different sites suggests site-specific responses to changing environmental conditions and the operation of local geomorphic threshold factors. The association of colluvial deposition (and, therefore, hillslope erosion) with arid phases is likely due to a decrease in vegetation cover under arid conditions. In wetter conditions slopes were more stable and soil formation occurred (Figure 11.1).

Figure 11.1(a). Three-phase donga erosion displaying cut and fill sequences, KwaZulu-Natal.

Figure 11.1(b). Donga erosion into colluviums with paleosols, northern KwaZulu-Natal.

Similar late Pleistocene ages have been obtained from colluvia in Swaziland (Price-Williams *et al.*, 1982). The origin of colluvium is fully described by Botha and Partridge (1988) and clearly relates to many forms of erosion and weathering. In the Sneeuberg Range, Great Karoo, extensive colluvial sediments are of Holocene age with dated paleosols marking periods of relative stability (Holmes *et al.*, 2003). Present day gully systems represent a single phase of incision of unknown, but probably recent age (Boardman and Foster, 2008). Compton *et al.* (2010) show that there has been a recent tenfold increase in the mud flux of the Orange River compared to mean Holocene rates and they suggest that this reflects increased soil erosion from heavily cultivated areas in the eastern catchment and grazing lands in the southern catchment. Due to the potential for sediments to be stored between erosion sites and the rivers, they suggest that these figures imply a maximum increase in soil erosion of a hundredfold. This accords with present-day estimates of the contrast between rates of erosion on valley-side badlands and deposition in valley-bottom reservoirs in small catchments in the Sneeuberg (Boardman and Foster, 2008). Dardis (1990) also records "rapidly accelerating erosion during the latter part of the past 1 ka" in Swaziland.

Colonial settlers, and in particular their introduction of sheep, appear to have induced significant overgrazing, the loss of grass and invasion of shrub. Deterioration of grazing in the Cape in the eighteenth century was noted (see Sparrman, 1977; Van der Merwe, 1945). Shortage of pasture around Graaff-Reinet in the 1810s and 1820s was related to overgrazing. Movements north through the Sneeuberg by trekboers were the result of a constant search for good pasture and water supplies. Many early writers noted these problems (see Beinart, 2003). At Wellwood, in the Sneeuberg, during the late nineteenth century, overstocking led to erosion and in 1920 Richard Rubidge noted that the vleis which had been an important water resource "were unhappily all [...] washed out into deep dongas' and that most had been formed within his recollection" (Beinart, 2003:311).

3. Overview: Twentieth century and after

The early "official" or government focus on land degradation was on the arid parts of the region (Anon., 1923). The observed decline in vegetation cover and associated soil loss was attributed primarily to the impact of land use and, in particular, to overgrazing. Acocks (1953) continued with this theme and produced a series of influential maps showing the extent of land degradation in South Africa and how it had changed over time. He emphasised the movement of more arid-adapted shrublands into mesic grassland environments in the eastern parts of the Karoo (Meadows, 2003a) and argued that overgrazing was the primary reason for this phenomenon. Significant state intervention ensued, including the introduction of stock reduction programmes and an improvement in farm infrastructural development (e.g. fencing) and livestock management systems (Hoffman *et al.*, 1999a). These interventions, as well as an increase in environmental awareness of the farmers themselves played an important role in slowing and even reversing the trend in land degradation in much of the Eastern Karoo (Hoffman and Ashwell, 2001).

Because of the political fragmentation of South Africa, no national review of the land degradation problem was undertaken until 1999. Early assessments for the country largely ignored conditions in the former homelands and "self-governing territories". The urgency to present a national perspective was given impetus when South Africa ratified the United Nations Convention to Combat Desertification (Anon., 1995) in 1997. The first national review of land degradation in South Africa (Hoffman *et al.*, 1999b, Hoffman and Ashwell, 2001) rejected Acocks' view of the Eastern Karoo as the major focus for land degradation and argued instead that it was the communal areas, particularly those in the eastern and north eastern part of the country, where soil erosion and vegetation degradation were greatest. This assessment was based on qualitative information derived from 34 regional workshops attended primarily by agricultural professionals such as resource conservation technicians and agricultural extension officers.

The focus of the report was on the 367 magisterial districts, which existed in South Africa at the time and included an assessment of water, soil and land degradation. It also presented an analysis of the primary social and biophysical correlates of soil and vegetation degradation (Figure 11.2).

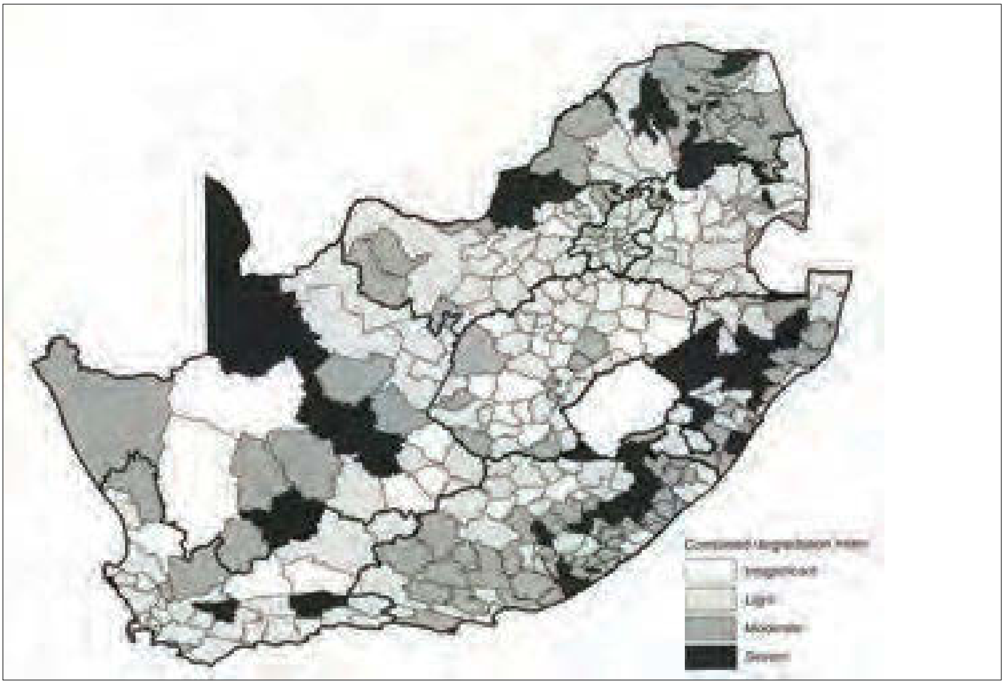

Figure 11.2. Land Degradation index (after Hoffman and Ashwell, 2001).

This report and ensuing publications (Hoffman and Todd, 2000; Hoffman and Ashwell, 2001; Meadows and Hoffman, 2002) suggested that differences in land tenure accounted for much of the pattern in land degradation in the country. Communally-managed areas of the former homelands were generally perceived to be more degraded than privately-owned, commercial areas although there were many exceptions to this pattern. Hoffman and Todd (2000) also provided an analysis of the key biophysical and socio-economic parameters associated with both soil and vegetation degradation and Hoffman and Ashwell (2001) present a spatially-explicit map of the most degraded magisterial districts in South Africa. They suggested that soil and vegetation erosion was greatest in districts with a large and poor, rural population where soils were steep, mean annual temperatures high and which contained a large proportion of easily-erodible soils. Magisterial districts of greatest concern were concentrated in the former homeland areas of the Eastern Cape, KwaZulu-Natal and Limpopo Provinces of South Africa, particular those communal areas situated along the foothills of the Drakensberg escarpment. Hoffman and Todd (2000) concluded that while socio-economic, biophysical, climatic and land use factors were all-important determinants of land degradation they differed in importance from area to area and generalisations were difficult.

Recent studies, which are based on more detailed quantitative estimates of land degradation from remotely-sensed data, have generally supported the patterns outlined in Hoffman and Ashwell (2001). For example, in their study of the Limpopo Province, Wessels *et al*. (2007), found similar trends of increasing land degradation in the communal areas of the province although there were many exceptions to this general pattern. Le Roux *et al*. (2008) also reported higher soil loss rates associated with erodible soils on steep terrain in the communal grazing lands of the eastern part of South Africa. However, another recent national review of land degradation, undertaken by Bai and Dent (2007) supports some, but not all of the findings of Hoffman and Ashwell (2001). Bai and Dent (2007) used remote sensing techniques to derive a quantitative assessment of land degradation that, in one measure, is reflected as a reduction in net primary production. They agree that much of the west of the country has shown increased production in the last 23 years, while the north east of South Africa in particular has experienced a decrease in primary production estimates over this period. Bai and Dent (2007) find some evidence for increased land degradation in the communal areas, but the relationship was far weaker than suggested previously. In their analysis high levels of degradation appeared closely associated with formerly cultivated fields, an idea which is strongly supported by the work of Kakembo and Rowntree (2003) and Keay-Bright and Boardman (2006) in the Eastern Cape. Historical factors such as land abandonment are also important determinants of the extent of land degradation and should be distinguished from an analysis of current trends (Bai and Dent, 2007).

Figure 11.3. Badlands developed on land cultivated until the 1980s, Compassberg Farm, Sneeuberg, Eastern Cape Province of South Africa.

Several new, quantitative approaches to land degradation assessment have recently been developed for South African landscapes (e.g. Thompson *et al*., 2009). In addition, accurate, large-scale measures of land cover change (Coetzer *et al*., 2010) and new developments in the conceptual approach to desertification research (Reynolds *et al*., 2007; Scholes, 2010) have also emerged. Together, these advances hold much promise for the assessment and monitoring of land degradation in the region. However, field determinations and the involvement of local land users will always be necessary to understand the full extent of the problem. Furthermore, the involvement of different stakeholders is critical for finding solutions to the negative effects of land degradation.

4. Soil erosion by water

The effects of runoff are most clearly seen in the development of degraded land, badlands and gullies. The 1951 *Report of the Desert Encroachment Committee* referred to dongas having become "a symbol of the countryside, not least in the Karoo" (quoted in Beinart, 2003:368) and Acocks (1975) claims that the "conversion of 32 200 km^2 of grassveld into eroded Karoo can only be regarded as a national disaster" (Beinart, 2003:370). The degree of degradation is clearly open to question. Indeed, the very term is problematic and value-laden as noted by Hoffman and Todd (2000). Roux and Vorster (1983) recognised five stages in the progression from pristine veld to *desert*. Less controversially, the end product of the process may be regarded as badlands, though whether the process is reversible seems open to question (Dickie and Parsons, 2010).

Badlands are almost certainly more widely distributed than is reported in the literature. There is a tendency to overlook them as a *normal* part of the landscape. However, they are reported from the Eastern Cape by Kakembo and Rowntree (2003), Kakembo *et al*. (2009), Boardman *et al*. (2003a), Boardman and Foster (2008) and Keay-Bright and Boardman (2009); from KwaZulu-Natal by Watson (2000) and Clarke *et al*. (2003); from Swaziland by Price-Williams *et al*. (1982) and Dardis (1990).

Figure 11.4. Badlands on footslopes, Sneeuberg, Eastern Cape Province of South Africa.

Badlands are located on colluvium, which has accumulated on footslopes and valley bottoms (Figure 11.3). They comprise intricate networks of gullies with little of the original ground surface remaining. Depth to bedrock tends to control local relief on badlands with erosion rates slowing as gullies reach bedrock (Figure 11.4). On the unstable, eroding slopes vegetation tends to be limited to shrubby remnants with grasses concentrated on areas of sediment accumulation in channels. In some areas, badlands have been colonised by invasive species such as *Pteronia incana* (Peddie District, Eastern Cape, Kakembo *et al*., 2007).

Figure 11.5. Badlands on formerly cultivated land and subsequence invasion of *Pteronia incana*, Peddie, Eastern Cape Province of South Africa.

Badlands have been associated with the abandonment of cultivated land (Kakembo and Rowntree, 2003) (Figure 11.5). The reasons for abandonment in this area are unclear. In other areas, the abandonment of cultivated land has also been associated with the development of erosional features (Sonneveld *et al.*, 2005; Marker, 1988). *Palaeobadlands* which have been buried by subsequent cycles of colluvial deposition are described from KwaZulu-Natal (Botha *et al.*, 1994). Their existence suggests that severe erosion may be driven by shifts in climate, for example to aridity as noted above. It may not always result from human mismanagement of the land.

In the Sneeuberg, badlands have developed on land formerly cultivated for dryland wheat and fodder. These farming practices have ceased since the 1980s and, at some sites, at earlier dates. On marginal land subject to drought and erosion the costs of cultivation outweighed the benefits. Keay-Bright and Boardman (2009) show that at one site cultivation ceased around 1925 and by 1945 an extensive area of badland had developed with little change since in terms of the areal extent of the badland. During the period of cultivation we assume (though there is limited evidence for this) that erosion was substantial, but gullies were filled in by farmers before the planting of the next crop. However, not all formerly cultivated areas have become degraded and not all badlands have developed on formerly cultivated land (Keay-Bright and Boardman, 2006). In areas where badlands have developed in the absence of cultivation – and this must include many areas beyond the Sneeuberg – the explanation seems to be overgrazing together with the influence of erodible soils, drought and fire. In the Karoo, stock numbers peaked around 1930, but had been high since the mid-nineteenth century. Rates of erosion and depths of incision on badlands in the Sneeuberg support the idea that they have developed since the introduction of European farming into the area (Keay-Bright and Boardman, 2009).

Rates of erosion at ten sites on badlands in the Sneeuberg have been monitored for almost 10 years using arrays of erosion pins (Figure 11.6). Annual average loss from eroding pins is between 7.6 and

17.3 mm. Net loss which takes into account accumulation of eroded sediment at some pins for two sites on the farm Good Hope is approximately 3 and 5 mm y^{-1} respectively. This translates to 51 and 85 t ha^{-1} y^{-1} using a measured bulk density of 1.7 g cm^{-3} (Boardman, 2010). This may be compared with sediment export from the Good Hope badlands of 54.8 t ha^{-1} y^{-1} based on sediment sampling of runoff (Keay-Bright and Boardman, 2009).

Figure 11.6. Erosion pin site, Compassberg Farm, Sneeuberg, Eastern Cape Province of South Africa.

Rowntree and Foster (in press) report surface lowering rates using erosion pins at Ganora Farm in the Sneeuberg for a 20-month period of 5.8, 4.7 and 9.1 mm for crests, midslopes and channels respectively in badland topography.

Runoff from badland areas in the Sneeuberg has been observed in response to rainfall events as low as 10 mm (see, for example Keay-Bright and Boardman, 2006). As might be expected, there is a positive correlation between amounts of sediment eroded and numbers of rainfall events >10 mm in monitored time periods (Figure 11.7). Amounts of material available for erosion and transport are also influenced by weathering history, in particular wetting and drying and frost action.

Gullies (dongas) are widespread in southern Africa. They have been reported from Lesotho (Rydgren, 1988; Showers, 2005), from Swaziland (Price Williams *et al*., 1982; Dardis, 1990; Felix-Henningsen *et al*., 1997; Mushala *et al*., 1997; Morgan *et al*., 1997), from the Swartland (Talbot, 1947), from the Peddie area of Eastern Cape (Kakembo *et al*., 2009), from the Sneeuberg, Eastern Cape (Holmes, 1991; Boardman and Foster, 2008), from KwaZulu-Natal (Botha *et al*., 1994) and from Namibia (Eitel *et al*., 2002). Photographs of spectacular gullies in the Hofmeyr, Herschel and Middelburg districts of the Eastern Cape are to be found in Beinart (2003; Figure 17), Hoffman and Ashwell (2001:31) and Boardman and Foster (2008; Figure 5) respectively (Figures 11.8 and 11.9).

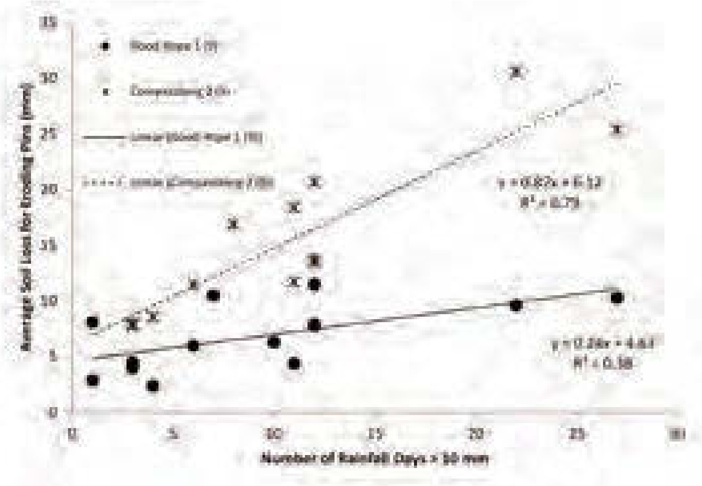

Figure 11.7. Average soil loss at Good Hope 1 and Compassberg 2 erosion pin sites for periods of time of varying length, plotted against rainfall for each period. Rainfall data from Compassberg Farm, Sneeuberg, Eastern Cape Province of South Africa (after Boardman, 2010).

Figure 11.8. Severe gully erosion, Herschel, Eastern Cape Province of South Africa (after Hoffman and Ashwell, 2001).

Ascertaining the age of the gullies continues to be a problem. Rienks *et al*. (2000:12) note the "common belief that change of land use from Iron Age pastoralism to increasing arable agriculture in the eighteenth to late nineteenth centuries, and subsequent mechanisation and intensification of subsistence farming led to an acceleration of gully erosion" (Berjak *et al*., 1986; Showers, 1989; Viljoen *et al*., 1993; Watson, 1996; 1997). An alternative view suggests that, "variation in vulnerability of parent materials, combined with certain hydrological and climatological conditions, predisposes an area to donga formation" (Rienks, 2000:12). The most convincing evidence for the latter view is the cut-and-fill sequences in colluvial sediments of late Quaternary age described by Botha *et al*. (1994).

In Lesotho, Showers (2005) suggests that accelerated erosion was rare before the coming of Europeans in the 1830s and became noticeable in the late nineteenth century. By 1935 a report suggests that 10% of arable land is threatened by existing dongas (Showers, 2005:145). The imposition of inappropriate "conservation structures" from the 1940s to the 1990s exacerbated the situation (Showers, 2005).

Dramatic recent gully development is described by Sidorchuk *et al*. (2003) from Swaziland. The gully was absent on 1961 photographs and by 1998 was 440 m long. For the first 30 years it grew at an average rate of 14 m per year. The area is heavily overgrazed and some gully development is along cattle paths. In other areas of Swaziland, Goudie (pers. comm.) suggests early nineteenth century deforestation for iron smelting as a trigger for gully growth. High rates of gully headward retreat in recent years are also reported from Swaziland by Morgan and Mngomezulu (2003). Overgrazing exacerbated by land tenure arrangements appears to the cause of much Swaziland gullying (Mushala, 1997).

Figure 11.9. Gully, Compassberg Farm, Sneeuberg, Eastern Cape Province of South Africa.

In the Otjiwarongo region of northern Namibia in the mid-nineteenth century, Holocene soils were buried by slope wash and the landscape was then affected by rill and gully erosion, indicating increased runoff in the late nineteenth century and early twentieth century probably related to intensification of cattle farming by European settlers (Eitel *et al*., 2002).

In the Sneeuberg, gully systems seem little changed in dimension since 1945 air photographs. They are active in the sense that frequent floods move sediment from the hillslopes including formerly cultivated lands and badlands along the gullies, but with little change in form (Keay-Bright and Boardman, 2006; Boardman and Foster, 2008). In the Peddie area of the Eastern Cape, gullies are closely associated with the abandonment of cultivated fields on communal lands and have developed predominantly since the mid-1970s (Kakembo and Rowntree, 2003; Kakembo *et al*., 2009).

In classic work in the Swartland of the Western Cape Province of South Africa, Talbot (1947) mapped from air photographs the distribution of gullies and degraded land in 1938 (Figure 11.10). Remapping of the same area has shown a dramatic decrease in the gully density and the area of degraded land (Morel, 1998) (Figure 11.11). In the absence of significant changes in rainfall, this rehabilitation of the Swartland landscape is attributed to a "broadly successful" series of conservation interventions supported by government acts and extension services admittedly directed specifically at a white farming electorate (Meadows, 2003b).

Figure 11.10. Erosion near Durbanville, Western Cape Province (after Talbot, 1947).

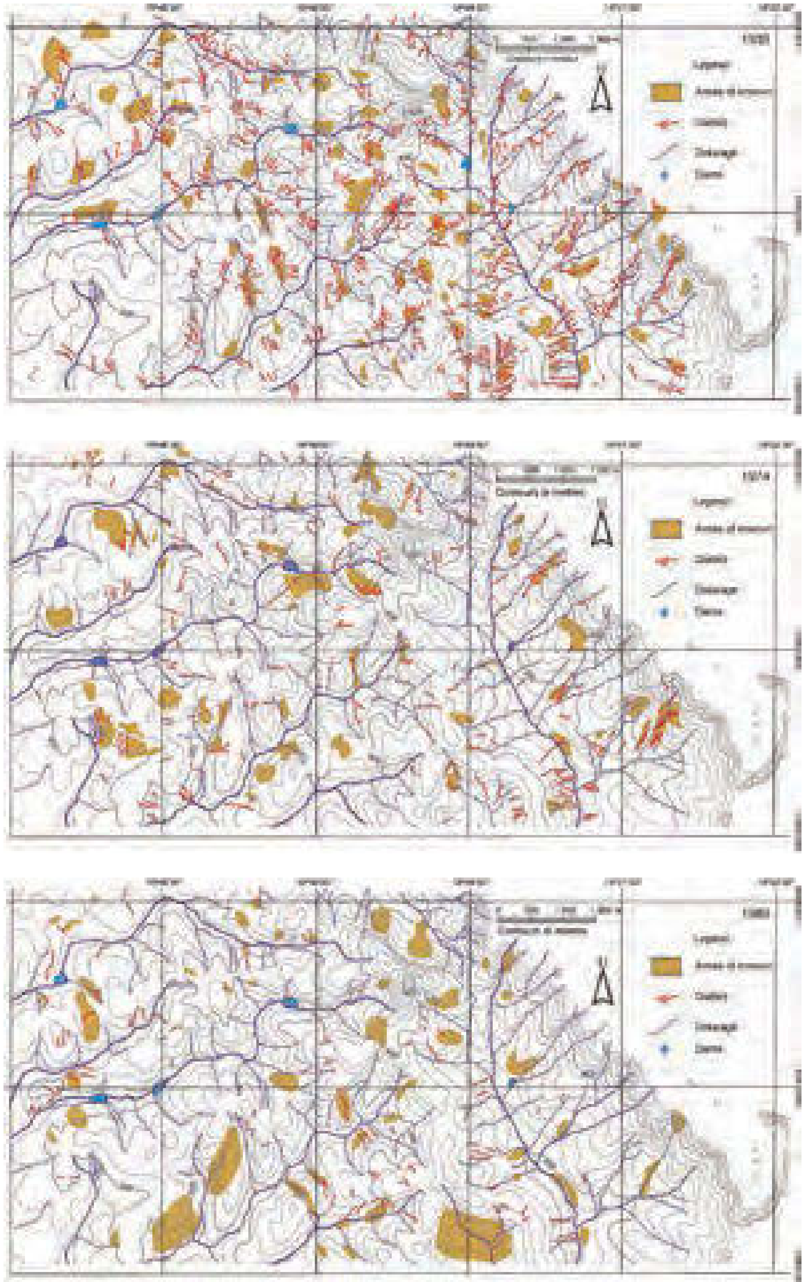

Figure 11.11. Erosion in the Swartland, Western Cape Province (after Morel, 1998).

The final problem concerning gully development is the role of piping and sub-surface flow. Beckedahl (1998) has shown that piping is frequent in southeastern Africa from Swaziland to the Eastern Cape. The development of gullies is intimately associated with piping on formerly cultivated land in the headwaters of the Tinana River, north of Mount Fletcher, Eastern Cape, which suggests that the piping here is a recent phenomena (Foster, pers. comm.). However, a detailed analysis of different measures of susceptibility to dispersion points to difficulties in using such measures and suggests that piping is probably not the main cause of gully formation at the particular site in KwaZulu-Natal (Rienks *et al*., 2000).

5. Soil erosion by wind

In this section, wind erosion refers to *accelerated* erosion, or erosion linked to human activities on grazing and arable land. For a discussion of aeolian processes, particularly in the context of dust emission and dune building on a regional or sub-continental scale, and climatically-driven environmental change, the reader is referred to Chapters 6, 7 and 12. Beckedahl *et al*. (1988) pointed out that, while soil erosion by wind is important in southern Africa, it had not previously received adequate attention. This remains true today, with very little work having been undertaken to either identify where wind erosion is prevalent, or to quantify the nature of wind erosion. Wind erosion is essentially a function of erodibility (the tendency of soil to erode, or its vulnerability to erosion by wind) and erosivity (the propensity or power of the wind to erode); see Wiggs (1997) in this regard.

Erodibility is a function of soil type, surface properties such as soil crusting and roughness, vegetation cover, soil moisture, agricultural activities and soil management and conservation practices. Both natural (inherent soil properties) and anthropogenic (farming practices) play a role. Erosivity is determined by wind speed and direction (relative to, for example, the direction of ploughing on agricultural land) as well as wind characteristics such as gustiness.

Pioneering work (see also above) on wind erosion in South Africa was undertaken by Talbot (1947), who mapped the extent of erosion by both water and wind in the Swartland and Sandveld regions of the Western Cape. Morel (1998) provided a re-evaluation of the extent of land degradation in these areas, which included an assessment of the nature and extent of wind erosion. However, detailed process studies from southern Africa remained lacking.

By way of contrast, wind erosion studies are commonplace in the northwest and Southern High Plains of the USA where agricultural activity combines with semi-arid environments, highly erodible soils and seasonally active strong winds (e.g. Sharratt *et al*., 2007; Zobeck and Van Pelt, 2006). Indeed, in the Great Plains region of the United States of America the strong scientific paradigm resulting from the dust bowl years of the 1930s encouraged the continuous development of wind erosion equations such as the RWEQ (Fryrear *et al*., 1998) and WEPS (Hagen, 1991) in order to facilitate conservation programmes and agricultural management techniques aimed at reducing the loss of productive soil to wind erosion. Whilst the calibration of these predictive equations is uncommon outside of the USA, similar scientific rigour as applied to wind erosion is also evident in Europe (Böhner *et al*., 2003; Funk *et al*., 2004), Australia (Webb *et al*., 2009) and China (Xu *et al*., 1993; Li *et al*., 2004). These studies have established the primary associations between local and regional scale environmental controls on wind erosion activity and their relationship with agricultural management techniques. They have also provided fundamental data quantifying the degree of wind erosion and aeolian dust activity in current environmental conditions. Such quantification is an important step forward in the prediction of future likely wind erosion and dust dynamics in regions of potential changing climate, including southern Africa, and human activity.

Only recently have such detailed data become available, and then only from one study, from an area of fairly intense farming activity on erodible soils in a semi-arid (the area is bisected by the 500 mm isohyet), and likely future drying, climate. This study was undertaken by Wiggs and Holmes (2011) on arable land in the west-central Free State Province of South Africa (Figure 11.12). They demonstrated the vulnerability of sandy soils of the Free State to wind erosion, resulting in significant quantities of dust being generated from agricultural land, particularly in spring. There is also evidence from the area of historic, but recent, aeolian erosion in the form of extensive fence line dunes and barbed-wire fences buried under sand (Holmes *et al.*, 2012). Further studies are currently underway to quantify dust emission from point sources such as pan surfaces in the western Free State. The lunette dunes downwind of Free State pans are frequently capped, unconformably, by sand, which typically dates by optically stimulated luminescence (OSL) from the period subsequent to the movement of European settlers into the area (Holmes *et al.*, 2008). Wind erosion in southern Africa is an under-recognised facet of soil erosion in general, and a topic requiring investigation on agricultural land in other regions of southern Africa.

Figure 11.12. Wind erosion in the Free State, showing (a) fence line sediment accumulation, (b) erosion monitoring on ploughed land and (c) dust blowing off Deelpan near Bloemfontein.

6. Impacts of soil erosion

The long-term impacts of soil erosion are well known, but often ignored because of lack of data. Erosion results in the loss of the nutrient-rich A horizon and often the exposure of infertile or compacted subsoil materials. These losses inevitably result in a loss of productive capacity be it for the growth of arable crops, forests or grazing. Without the use of expensive fertilisers productive capacity will decline. Stock numbers in the Karoo have declined from a peak around 1930, too much lower figures at the present time (Hoffman *et al*., 1999a, Figure 16.4). There could be several reasons for this decline such as a greater environmental awareness of farmers, but it seems reasonable to suggest that in many areas a landscape showing signs of degradation is not capable of sustaining the previously high numbers of stock. Recommendations for stock numbers are now much below maximum values in order to protect the veld and prevent erosion (Keay-Bright and Boardman, 2007).

The short-term impacts of erosion are easier to quantify. By far the most important concerns the siltation of reservoirs and dams (Figure 11.13). Reliable water supplies depend on storage reservoirs and recent estimates suggest that South Africa already allocates 98% of its available water resources. Many reservoirs are accumulating sediment at rates that make water provision unsustainable in the future. The country has remarkably good data on sedimentation in major reservoirs (Msadala *et al*., 2009). The challenge is to relate sediment yield to reservoirs to erosion data, the elusive, temporally and spatially variable Sediment Delivery Ratio (SDR), being the key (or chimera?). Small catchment studies suggest that maximum erosion rates (e.g. on badlands) can be two orders of magnitude greater that sediment yield to nearby small farm dams. But within such small catchments erosion rates vary with soil type, slope angle and vegetation cover. System connectivity is an important explanatory factor.

Small- or medium-scale studies of the factors affecting reservoir sedimentation are lacking. The reasons for enhanced sedimentation and the areas contributing to sediment yields need to be known. An exception is the study of sedimentation in the Hazelmere Dam, KwaZulu-Natal, with respect to the role of the exceptional floods of 1987. The reasons for the high erosion potential in the catchment are shown to be the large-scale population increases, vegetal degradation and road construction. The dam was built in 1975 and Msadala *et al*. (2009) report annual average sediment yields of 713.69 t km^2 y^1. By 1993 the dam had lost 25% of its total capacity partly due to the impact of the 1987 floods (Russow and Garland, 2000).

Figure 11.13. Small farm dam, built 1980, Sneeuberg, Eastern Cape Province.

The role of small farm dams as important stores of sediment has recently been highlighted (Boardman *et al*., 2009; Boardman and Foster, 2011). Mapping of the distribution of dams in a sample area of the Sneeuberg together with volumetric estimates of stored sediment, suggests a density of 1 dam km^{-2} and a conservative estimate of sediment storage of 2 million m^3 100 km^{-2}, equivalent to a store of 27 000 t km^{-2}. Of the 95 dams mapped in the sample area, 46 were full of sediment and 28 of the dams were breached and therefore potential sites of sediment loss. The breaching of unmanaged small farm dams constitutes a new potential source of sediment that will reduce the working life of downstream storage reservoirs.

The impact of soil erosion on river channels has been neglected. In the Bell River catchment, Eastern Cape, evidence suggests that increased channel instability is due to sediment input from sheet and gully erosion caused by poor veld management (overgrazing) since European occupation in the 1870s (Dollar, 1992; Dollar and Rowntree, 1995; Rowntree and Dollar, 1996).

7. The future of erosion in southern Africa

Future changes in land use in southern Africa are of great political import. They will be driven not merely by the economics of farming, but by a desire for an equitable redistribution of land. This, however, in an area susceptible to erosion needs careful planning. Settling of peasant farmers on uninhabited but susceptible land in KwaZulu-Natal led to a 25-fold increase in actively eroding areas (Watson, 1996). With this experience in mind, Watson and Ramokgopa (1997) warn of the risks of redistribution of unsuitable land in the Land Reform Programme (see also Watson, 2000).

Climate change also poses a threat of increased erosion. Meadows and Hoffman (2003) emphasise the risk to former homeland states where erosion is already a major problem. Lower soil-moisture availability due to higher temperatures may result in crop under-production or loss of vegetative cover on grazing lands. They also mention a predicted decrease in the reliability of seasonal and annual rains and an increase in the frequency of extreme events. Mason *et al*. (1999) examined changes in daily rainfall across the whole of South Africa in the periods 1931-1960 and 1961-1990. They found significant increases in the intensity of extreme rainfall events over 70% of the country. Similar trends have been found in an analysis of magnitude and frequency of daily rainfall in the Karoo uplands. At Cranemere on the Plains of the Camdeboo in the Eastern Cape, there is a reduction in the recurrence interval of the 80-year daily rainfall pre-1950 to a recurrence interval of ~5 years post-1950 (Foster *et al*., in press). Under future climate warming scenarios, there is likely to be an increase in the amount of rain per rainday rather than the number of raindays (Fowler and Hennessy, 1995).

As a specific example of potential climate change impact on erosion, Meadows (2003b) applies a regionally downscaled climate model for monthly rainfall under a doubled carbon dioxide scenario to the Swartland. This suggests a likely 100% increase in May rainfall. This is potentially serious for winter wheat farming in that ground cover is at a minimum at this seed-sowing time and, therefore, fields are vulnerable to erosion. This is precisely the same scenario of risk as for autumn-sown winter wheat in southern Britain.

8. Conclusion

All accelerated erosion is a result of unwise management practices with the location and rates controlled by surface vegetation, soil character and slope. Thus the ultimate drivers of erosion are socio-economic factors that influence what the land manager grows/produces and where it is grown and what practices are used (Stocking and Murnaghan, 2001:25; Boardman *et al*., 2003b). A classic example of large-scale erosion driven by economics is the impact of the Wheat Importation Act (1930), which encouraged high wheat prices and led to cultivation on many unsuitable hillslopes. The erosion effects in the Swartland are described by Talbot (1947) and Meadows (2003b).

Hofmann and Ashwell (2001) suggest that there is evidence for a slow down or even reversal of erosion in recent years in the Eastern Cape due to state interventions, farm level changes, such as fencing, and a growth in environmental awareness by farmers. Similarly, Garland and Broderick (1992) describe a decrease in eroded areas in the Tugela catchment, KwaZulu-Natal, between 1944 and 1981. This they suggest may be part of a regional trend and most likely related to changes in rainfall rather than any human influence. In a small catchment in the Sneeuberg, Eastern Cape, Keay-Bright and Boardman (2006) record a 15% decline in degraded areas 1945 to 2002. This they suggest is due to decline in stock numbers. However, rates of erosion from badlands in the catchment are still high probably driven by an upward trend in average rainfall per rain day over the last 100 years (Boardman and Foster, 2008).

The need for an accurate and reliable national survey of erosion is implicit in much of the literature. The review of technologies and achievements by Le Roux *et al*. (2007) is rather depressing. They show that most models are unvalidated, operate at inappropriate scales, ignore relevant processes or require far too much information which is not available: "there exists no methodological framework, or 'blueprint', to assess the spatial distribution of soil erosion types at different regional scales in South Africa." The National Biodiversity Institute approach (Garland *et al*., 2000) is regarded as being "limited" due to the lumping of data in magisterial districts and the subjectivity of the judgments. Hope seems to reside in the availability of high-resolution imagery (SPOT 5) and a GIS approach. However, their preference for emphasis on soil erodibility as against slope gradient is highly suspect if vegetation cover is to be ignored.

Clearly one of the big questions regarding the distribution of erosion in South Africa remains unanswered. We have already referred to conflicting evidence about the concentration of erosion in the former homelands. In the semi-arid, grazing-dominated Karoo rather limited evidence from the Sneeuberg suggests that that most serious erosion was probably associated with a century or so of overgrazing peaking in the 1930s, but that once established, gullies and badlands continue to function as transport systems and sources of erosion for decades if not centuries. This is partly because severely overgrazed landscapes require decades to recover. In the Sneeuberg, the little recovery apparent in the last 50 years suggests that reduction in stock numbers has little effect over these timescales (Boardman and Foster, 2008). In several areas the importance of abandoned, cultivated land as a focus of erosion has been demonstrated. While researchers acknowledge the potential importance of fire (managed and natural), and droughts, in the erosion story the evidence for their importance is mainly circumstantial and from small-scale studies.

Swaziland and Lesotho are acknowledged problem areas in terms of land degradation and in particular the prevalence of gullies. On the basis of rather limited evidence (WMS Associates, 1988; Mushala 1988, 1997; Morgan *et al*., 1997; Chakela and Stocking, 1988; Showers, 2005), it has been suggested that they fall into the World's top 10 hotspots of erosion (Boardman, 2006).

Future erosion risks concern both changes in land use particularly those related to the Land Reform Programme and to the vulnerable former homelands and to climate change risks associated with rainfall variability, extreme events and increases in rainfall intensity. Alternatively, De Wit and Stankiewicz (2006) posit considerable decreases in South African rainfall by the end of the twenty-first century, which would certainly increase the risk of wind erosion in the western part of the country.

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12

Landscapes and Environmental Change

Landscapes and Environmental Change

Michael E Meadows

1. Geomorphology and change: is there a climatic geomorphology?

The concept of change through time is fundamental to an understanding of landscapes and the processes that form them. Indeed, landforms – both individually and in groups – are manifestations of the complex set of relationships between underlying geological structure and a range of temporally variable environmental processes. Thus, geomorphology is more than the simple description of contemporary landscapes and processes; as a science it must seek explanations as to how changing environmental conditions have influenced processes over time and left their imprint on the landscape. This chapter explores the evidence for such changes, the legacy of geologically recent climate changes on the southern African landscape and the relevance of this to an understanding of how geomorphology may respond to future environmental changes, more especially those invoked through human activity. Without doubt, the answer to questions as to how landscapes will adjust to climate changes of the future lies in a more thorough understanding of the past, and southern Africa is no exception.

As noted by Slaymaker *et al*. (2009:4): "Climate's role in landscape change has long been of interest in geomorphology" and climate change is arguably the most important element of environmental dynamics in the geomorphological context. Of course, landforms are products of a wide range of factors, including underlying geological composition and structure, tectonics, time and human activity. Climate influences a wide range of geomorphological processes and it is possible – certainly at the global scale – to identify broad sets of landscape features that are, to a greater or lesser degree, determined by climate. The concept of climatic geomorphology has a rich history. William Morris Davis (1899) recognised the distinctive nature of landforms found in different climatic regions and went so far as to formalise this in theorising the *normal*, *arid* and *glacial* cycles of erosion. This developed into an important philosophy in geomorphology, culminating in the recognition of morphoclimatic zones based on the two variables of mean annual temperature and precipitation (Peltier, 1950). Although the limitations of such a scheme are many, the intuitively attractive idea of relating large-scale landform suites to climatic characteristics remains entrenched. Thus, commonly used terms such as *tropical* or *periglacial* geomorphology testify to the perceived importance of climate in determining landscape characteristics. Moreover, we may be approaching a more fundamental theoretical understanding as to the factors determining landforms; Perron *et al*. (2009) have modelled how valley spacing is a function of the ratio between channel incision and hillslope transport modulated by climate and structure.

There are, of course, considerable complexities in the set of relationships that characterise the climate – process – geomorphology linkages. As Schumm (1991) has noted, there are "ten ways to be wrong" in interpreting past and predicting future geomorphological processes and forms on the basis of observations of the present. For instance, there are problems of scale, both temporal and spatial, problems of cause and effect (such as equifinality, for example in relation to glacial landforms, further discussed below) and problems of system response (e.g. sensitivity and thresholds, for example Marker and Holmes, 2005) all of which render the interpretation of southern African past environments on the basis of geomorphic features a fraught exercise (Meadows, 2001).

2. Geomorphological inheritance and the Quaternary legacy

The evolution of landforms over time influences their contemporary morphology and geomorphological features may be inherited from former climatic regimes. For instance, distinctively shaped valleys are carved by glaciers that leave behind both the morphology and the sedimentary deposits emanating from a glacial period. Thus landform shapes might indicate a particular climatic condition that no longer actually prevails at the site of the landforms themselves. This is the issue of geomorphological inheritance and is prominent in savanna and semi-arid landscapes in southern Africa, particularly those that are marginal to the deserts, where the inheritance of arid zone features (in localities that are no longer strictly speaking arid) is common (Meadows and Thomas, 2009). Some important questions are raised as a result; for example, are the deeply weathered regoliths and duricrusts that underlie large areas of southern African products of contemporary processes or of past conditions or, perhaps, a combination of both? Such are the complexities of climatic geomorphology; however, there is little doubt that changes in climate certainly influence geomorphic process and that, in turn, there is evidence in the landforms shaped by those processes that may be used to reconstruct the climates of the past.

The concept is developed in Thomas (2004) in respect of tropical areas using examples from the Quaternary as a model as to how landscapes respond to rapid change. Kniveton and McLaren (2000) ponder the cumulative effects of changing climate on geomorphological and sedimentological processes and note that spatial and temporal scales influence the answer. While larger scale landforms typically respond to longer-term evolutionary processes, a notable exception involves the linear dunefields of the Kalahari which comprise mesoscale landforms markedly impacted by short-term climate changes of the late Quaternary and also potentially strongly altered by human activity both within the last few decades and into the future (Thomas *et al*., 2005). Indeed, there are certain entire landscapes that are more obviously a product of Quaternary, most importantly late Quaternary events; these include the aeolian features of much of the arid and semi-arid parts of southern Africa. Inheritance is, therefore, an important legacy in regard to the relationship between geomorphology and ecology and in terms of landscape sensitivity to extreme events such as droughts and floods; of course it is also a key element of the explanation of the present distribution of features in the savannas. As Thomas (2006:122) notes, "over long timescales, landscapes become more complex palimpsests of repeated imprints of climate, extreme events and tectonics."

3. The geomorphic evidence for Quaternary climate change in southern Africa

Due to a combination of relative tectonic stability and predominantly semi-arid climates, the climatic history of southern African has resulted in landscapes that are replete with evidence of past changes, although their elucidation and interpretation have proved a significant challenge. There is a vast array of geomorphological features and processes that have imprinted themselves on the southern African landscape and these include morphological and sedimentary evidence of wind and running water erosion and deposition, lacustrine, glacial and periglacial processes, weathering, pedogenesis and sea level changes. It is not possible to comprehensively review all the manifestations of Quaternary climate change in the region but the following sections introduce some of the more obvious, widespread or interesting features in the landscape, identified according to their major mode of formation. Several of these landscape elements are explored in their own geomorphological context elsewhere in this volume.

3.1 Glacial and periglacial features

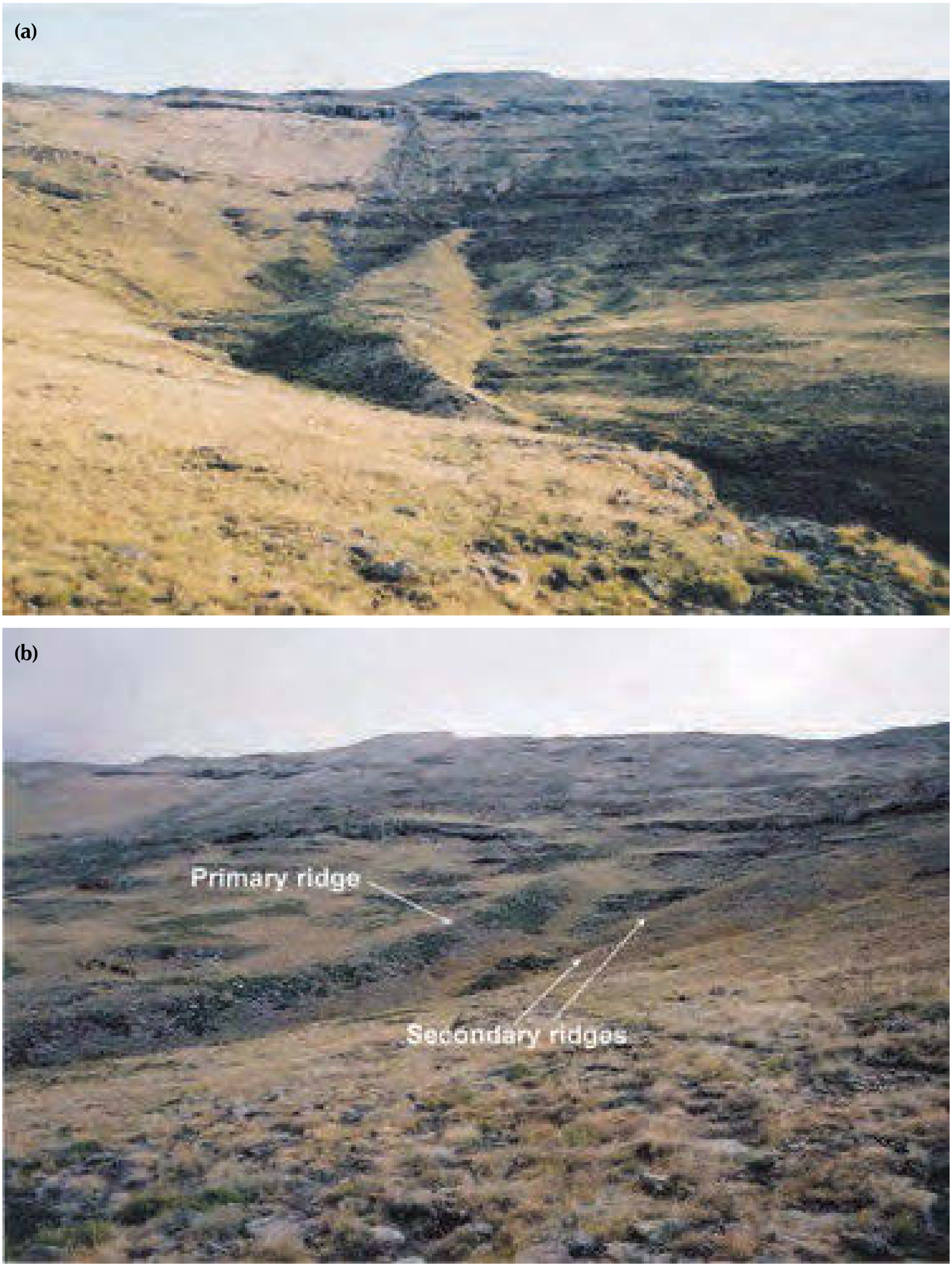

Figure 12.1. Ridges in the landscape of the Lesotho Highlands. These are from the Leqooa district and have been interpreted by some authors as indicative of niche glaciations (after Mills *et al.*, 2009).

Intuitively, one might assume that glacial and periglacial features would be indisputably associated with particular climatic features. After all, the change in physical condition of water from liquid to solid must have major implications for the mode of erosion, transport and deposition and their geomorphic expression. It is all the more surprising then that the evidence for glacial and periglacial conditions in southern Africa over the later part of the Quaternary has been the subject of such vigorous debate. The principle reason for this lies in the thorny old problem of equifinality – that similar features can be produced by different processes. A secondary reason is that views of some of the scientists involved have become polarised because the question, "Was there glaciation in southern Africa around the time of the global Last Glacial maximum (LGM)?" appears to prompt an obvious answer, namely. "yes" or "no". Certainly, there is widespread evidence of both active and inactive periglacial features, such as earth hummocks, stone or turf-banked lobes, block deposits and patterned ground in the Drakensberg (see Grab *et al*., this volume). Examples include those discussed by Boelhowers (1994) and Grab (2002), but the extent to which such features can be used to reconstruct minimum threshold temperatures during the coldest part of the Last Glacial Maximum (LGM) is complicated by the fact that palaeoprecipitation may have been overestimated in some models (see the critique by Nel and Sumner, 2008) and the poor understanding of environmental controls on frost action (Boelhouwers and Meiklejohn, 2002). Similar problems arise in the interpretation of what some believe to be clear evidence of glaciation in the high Drakensberg, including debris ridges and moraines (see Lewis and Illgner, 2001; Mills and Grab, 2005; Mills *et al*., 2009 and Figure 12.1). Lewis (2008) has gone as far as to suggest a regional terminology for the putative Drakensberg glaciation (the Bottlenek Stadial) but it must be noted that the evidence remains equivocal and, therefore, disputed not only because the features themselves may be difficult to identify but also because they are incompatible with other reconstructions of regional palaeoenvironments (see Hall, 2004).

3.2 Fluvial, alluvial, colluvial sediments and palaeosols

The erosion and deposition of sediments under the influence of running water has been extensively studied in southern Africa, although it proves difficult to interpret such features in the context of climate such that their use as palaeoclimatic proxies may be limited. Southern Africa has some major river systems, not least the Orange (Gariep), Limpopo and Zambezi as well as the extraordinary Okavango that dissipates into the sands of the Kalahari in northwestern Botswana without ever reaching the ocean. Rivers, more especially large ones, leave a palaeoenvironmental legacy in the form of suites of terraces controlled by a range of variables incorporating uplift and sea level change, although "... their cyclic formation ... almost invariably seems to have been a response to climatic fluctuation" (Bridgland and Westaway, 2008:285; and see Figure 12.2 for an example from the Sundays River Valley). Helgren (1979) made a classic study of the Vaal River terraces and documented a series of arid and humid periods in the late Quaternary corresponding to phases of terrace excavation and construction. The impressive staircase of terraces in the lower Sundays River of the Eastern Cape Province of South Africa is shown to have evolved in response to complex patterns of regional uplift, sea-level fluctuation and climate change over Pliocene-Pleistocene timescales (Bridgland and Westaway, 2008). These conclusions find support in the work of Compton and Wiltshire (2009) who identify cycles of terrigenous sediment deposition on the offshore western continental margin associated with cycles of sea-level change driven by global climate perturbations. Changes on shorter timescales are also recorded offshore, for example Holocene sediments on the Namaqualand Mudbelt record changing sea levels and Orange River palaeofloods (Herbert and Compton, 2007).

Due to its generally low relief, rivers of the continental interior of southern Africa have complex evolutionary histories. This makes it difficult to identify climate change *per se* as a driver of runoff dynamics because tectonic movements have tilted and warped the land surface resulting in substantial shifts in catchment areas and, accordingly, changes in river discharge. The Zambezi, the largest river in southern Africa, provides an excellent example of this (Moore *et al*., 2007). During the Quaternary, drainage reorganisations occurred in the low relief area around the plains where the national borders of Namibia, Zambia, Botswana and Zambia converge that have resulted in a complex array of alluvial and lacustrine sediments and landforms (Moore and Larkin, 2001), prominent among them being Lake Palaeo-Makgadikgadi (Thomas and Shaw, 1991); doubtless such features retain an imprint of accompanying climate changes, but they are very cryptic to interpret.

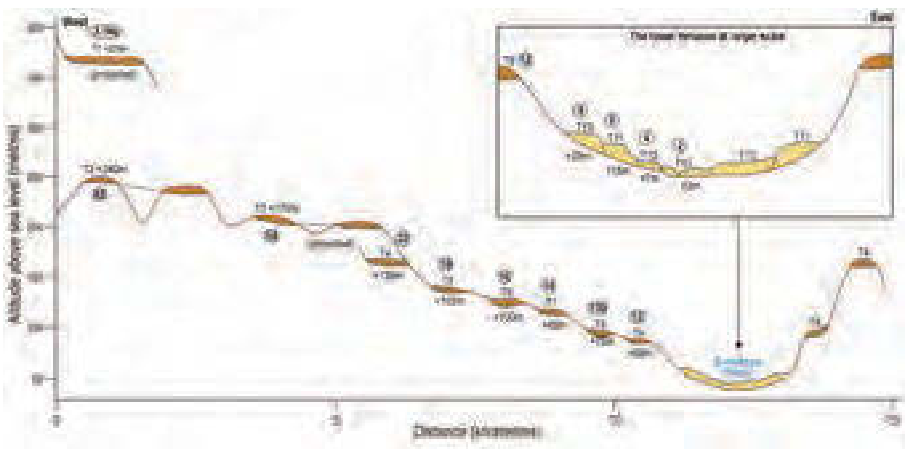

Figure 12.2. Terraces of the Sundays River, South Africa: schematic transverse profile across the terrace staircase at Kirkwood, Eastern Cape province, South Africa. Suggested MIS assignments are shown, based on incision/uplift modelling (after Bridgland and Westaway, 2008).

Smaller river systems have also been widely studied in the region, for example Tooth *et al.* (2009) demonstrate how alluvial sedimentation along the upper Klip River (a tributary of the Vaal) in eastern South Africa has responded to climate change and human impact over the last 30 000 years or so. The sensitivity of floodplain wetlands to various forms of human impact is perhaps their most significant (and worrying) finding, since such wetlands are widespread in the southern African interior and their hydrological, ecological and geomorphological importance is noteworthy.

Indeed, the balance between accumulation and erosion of sediments on slopes and in valley-bottom situations has been a focus of research in the region over many decades (see Figure 12.3 for an example of the complexity of some of the sequences). Precipitation shifts across the semi-arid/sub-humid boundary of approximately 500 mm per annum (and argued by Schumm, 1965, to be a significant sediment yield threshold) have occurred frequently during the southern African Quaternary in many regions. This led to attempts at palaeoenvironmental reconstructions of a series of cut and fill events in the dambos of central Malawi (Meadows, 1985). Subtle mechanisms of sediment movement on slopes have resulted in the widespread occurrence of colluvium (up to 20% of Africa south of the Zambezi is characterised by such material, see Watson *et al.*, 1984) which may exhibit stratigraphy and palaeosol development that may be indicative of cycles of landscape stasis and instability (see, for example, Botha and Federoff, 1995; Botha and Porat, 2007; Clarke *et al.*, 2003). These deposits seem to provide contrasting signals according to locality, since some appear to be highly responsive to environmental change (see, for example, Erikson *et al*., 2000, in Central Tanzania), while still others indicate periods of landscape stability (Holmes *et al.,* 2003). Fluvially associated tufas and travertines are widely interpreted as indicative of palaeoclimates in arid and semi-arid contexts and, in the Naukluft Mountains of Namibia, have been utilised to illustrate the changing balance of erosion and quiescence in the later Quaternary (Viles *et al*., 2007).

Figure 12.3. Colluvial and alluvial sediments exposed in a section in the northern Cederberg, Western Cape. Differences in characteristics of the sediments indicate phases of deposition and phases of stability.

The relationships between climate and soil formation are as complex as those between climate and geomorphology. Environmental change may leave characteristics of the soil profile in a state of disequilibrium, but some of the features may persist and act as an imprint of past conditions. Sequences of colluvial and alluvial deposits often contain palaeosols that can assist in the interpretation of the environmental factors responsible for their formation. In this regard, the pioneering work of Watson *et* *al*. (1984) is noteworthy and there are more recent examples from KwaZulu-Natal Province of South Africa (Temme *et al*., 2008) and from the Karoo (Holmes *et al*. 2008).

3.3 Lakes and pans: shorelines and sediments

As is the case elsewhere in Africa, there is a long tradition (see Grove, 1969 for an early example) in this region of utilising lacustrine sediments and associated fossil shoreline features in Quaternary palaeoclimate reconstructions (Burrough and Thomas, 2009). The basic principle is that higher lake levels correspond to changes in catchment hydrology that are often interpreted, too simplistically in many cases as it turns out, as a response to increased precipitation. Lake level fluctuations can result from changes in temperature and/or cloud cover, adjustments to groundwater flow and may also be effected by tectonic activity, such that their interpretation is rarely straightforward.

There is still powerful palaeoenvironmental reconstruction potential in these features, however, and there are numerous recent studies of lacustrine extensions during the late Quaternary. Examples of large late Quaternary lakes may be found at Etosha (Brook *et al*., 2007), Makgadikgadi (Burrough *et al*., 2009; see Figure 12.4), Tsodilo (Thomas *et al*., 2003), Mababe (Burrough and Thomas, 2008), Ngami (Burrough *et al*., 2007) and Chilwa (Thomas *et al*., 2009) and their palaeoenvironmental application may be improved through better chronological resolution emerging via the application of optically stimulated luminescence (OSL) dating.

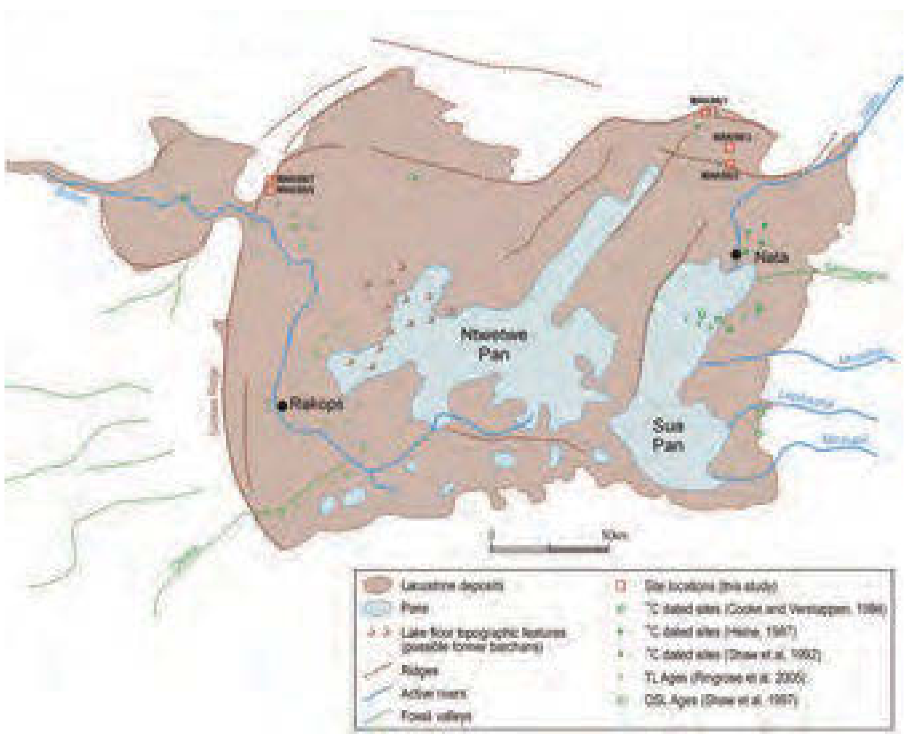

Figure 12.4. Palaeolake Makgadikgadi showing ancient shorelines (after Burrough *et al*., 2009).

There are innumerable smaller, closed depressions throughout southern Africa and doubtless some of these pans contain details of their palaeoenvironmental history recorded in their sediments and associated lunette dunes. Examples come from the summer-rainfall interior (Marker and Holmes, 1995; Thomas *et al*., 2002; Telfer *et al*., 2009), including the western Free State Province of South Africa where the largest concentration of such features are found (Holmes *et al*., 2008), the winter-rainfall coastal plain (Smith and Compton, 2004) and even the hyper-arid coast of Namibia (Compton, 2006).

3.4 Dunes: process and chronology

Much of arid and semi-arid southern Africa is underlain by unconsolidated sandy sediments, most spectacularly associated with the Kalahari and Namib. These sediments have been sculpted into various dune bedforms, many of which are today vegetated and form a major element of the arid landscapes discussed elsewhere in this volume (Thomas), although they may also be distributed more widely across the landscape without necessarily being obviously shaped into discrete bedforms, such as the so-called coversands of the southern Cape (Marker and Holmes, 2002; Holmes *et al*., 2007). The widespread occurrence of vegetation-covered dune bedforms beyond the modern desert margin may suggest that they are currently dormant and, therefore, relicts of past environments. However, there appear to have been numerous phases of activity and inactivity in these dunefields during the Quaternary and the narrative is far from clear; elucidating the complex record in these systems therefore remains an important research objective.

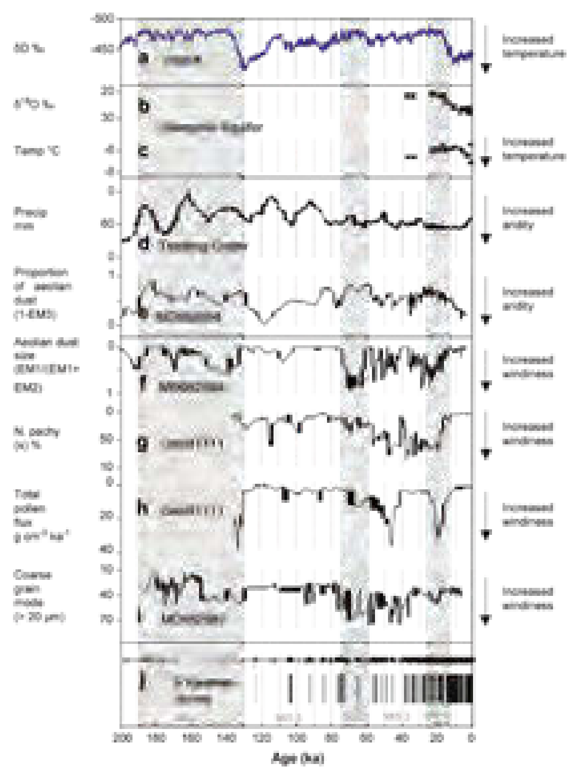

Figure 12.5. Comparison of luminescence ages from southern Kalahari Dunes with a range of proxy records for the southern hemisphere and southern African subcontinent (after Stone and Thomas, 2008).

Complex dunefield dynamics are illustrated in the ever-increasingly detailed chronology of activation and stasis founded on OSL dating of dune stratigraphy. Indeed, unravelling dune mobilisation records have proved to be a convoluted process. For example, at Tsodilo Hills in Botswana (Thomas *et al*., 2003), phases of late Quaternary dune activity (implying aridity) sometimes overlap with elevated lake levels (implying increased precipitation to evaporation ratios) and so it appears that models need to take account of variations in sediment supply and windiness as well as aridity in this context. An additional problem is that reduced partial pressures of atmospheric carbon dioxide around the Last Glacial may have reduced the vigour of C3 shrubs and trees and led to grassland expansion, thereby also influencing dune dynamics (Bond and Midgley, 2000).

Dune chronologies from the interior of southern Africa have, of course, been widely interpreted in palaeoenviromental terms. Recent examples span the subcontinent and include the Kalahari (Stone and Thomas, 2008, see Figure 12.5) and the Namib (Livingstone *et al*., 2010). The methodology is not without its complications and the original models that rather simplistically equated phases of dune activity with increased aridity are now being widely questioned (see, for example, Chase, 2009; Telfer *et al*., 2010).

3.5 Speleothems

Caves and associated speleothems have proved valuable archives of environmental change in southern Africa. Speleothems in the form of stalagmites and stalagtites provide a record of climate preserved in their isotopic composition and, coupled with the development of robust chronologies through U-series dating, can provide records of climate variation over extended periods (McDermott, 2004). Holmgren (2002) documented 17 southern African caves with speleothems and climate change records extending back several hundred thousand years and with decadal resolution in some cases. The most intensively studied records are from the large cavern system at Cango, in the southern Cape (Talma and Vogel, 1992) and Cold Air Cave at Makapansgat in the mountains to the south of the Limpopo (Holmgren *et al*., 1999).

3.6 Coastal geomorphology and sea-level changes

Geomorphological features at or near the coastline of South Africa have been widely studied to reveal the Quaternary history of climate, sea-level and human activity (see Carr and Botha, this volume). The most intensively researched region is the southern Cape, where the coastal platform south of the Cape Fold Belt Mountains has been an important sediment sink since the Cretaceous and provides excellent evidence of many aspects of environmental dynamics and landscape evolution over an extended time period (Marker and Holmes, 2010). The suite of dune cordons, especially in the vicinity of the Wilderness Lakes region, has been interpreted and reveals considerable detail regarding fluctuating sea levels and changing climates during at least the last two glacial and interglacial cycles (Illenberger, 1996; Bateman *et al*., 2004; 2011; Carr *et al*., 2006; Holmes *et al*., 2007. There are also detailed studies of geomorphological change revealed in dune characteristics and chronologies on the Agulhas Plain (Carr *et al*., 2007), the west coast (Chase and Thomas, 2007; Roberts *et al*., 2009) and the east coast (Botha and Porat, 2007; Porat and Botha, 2008). Dunes of the coastal platform are sensitive indicators of the nature and extent of late Quaternary sea level and climate change (Carr *et al*., 2010) and such information has been invaluable also in characterising early modern human behaviour and migration from localities that are proving to be of global significance from the archaeological perspective (Fisher *et al*., 2010; Compton, 2011).

4. Quaternary climate change and landscapes of southern Africa

Synthesising the range of records of environmental change remains a challenging task and would either be impossible or meaningless if based exclusively on geomorphological features. The brief synthesis that follows, then, is an attempt to collate a wide variety of proxies in order to provide a context for the foregoing discussion that highlights how landforms and associated processes are impacted by – and in turn provide evidence for – Quaternary climate changes in southern Africa.

Evidence that relates to the period prior to the Last Interglacial in southern Africa is extremely scarce, although the sequence of terrestrial sediments from the Tswaing Crater (see Partridge *et al*., 1999), in the core of the summer rainfall zone, has assisted in developing a synthesis of palaeoenvironments extending back through that period. Several lines of evidence from the Tswaing impact crater sediments indicate that, over a period spanning approximately the last 200 000 years, rainfall in the interior of the subcontinent has generally responded in phase with austral summer insolation. This is, in turn, a function of the 23 kyr BP orbital precession cycle (Partridge *et al*., 1997) and is accordingly out of phase with the North African palaeomonsoon signal. Despite uncertainties surrounding the age model, the original interpretation of the Tswaing record suggests that cooler (by around 5 °C mean annual temperature) and drier conditions prevailed during glacial phases, while warmer and more humid environments dominated the interglacials in response to southern hemisphere solar forcing.

The attempted reconciliation of evidence for late Quaternary climate change in the region by Scott *et al*. (2008) reveals that difficulties involving uncertain chronologies and interpretation of different proxies remain. Climate reconstructions for the summer rainfall region seem to defy any simplistic explanation as to forcing factors. The interplay of precessional solar forcing, movements of the ITCZ, sea surface temperature differentials between the Atlantic and Indian Oceans, together with possible increased winter rainfall influence at various times, all remain feasible mechanisms to explain the complex patterns of recorded late Pleistocene environmental change.

Longer records from the Winter Rainfall Zone are reviewed by Chase and Meadows (2007). Many of these indicate that there were periods during the late Pleistocene when austral winter cold frontal rains probably penetrated well beyond their current extent. The only really robustly dated palaeoenvironmental data for the penultimate glacial in this region comes from a series of marine cores offshore Namibia. Taken together, these suggest that MIS 6 was a period of increased humidity along the west coast of southern Africa, driven by an equatorward shift in the westerlies. Pollen evidence from core GeoB1711-4 is interpreted as indicating a northward expansion of elements of the Cape flora (Shi *et al*., 1998; 2000), while sedimentary data from core MD962094 shows a strong increase in the proportion of fluvial sediments in the core at this time (Stuut *et al*., 2002). In contrast, the transition to MIS 5 appears to have been relatively arid, with fluvial inputs to the southeast Atlantic dropping to almost zero during MIS 5e but increasing during MIS 5d and 5b (*ibid*). Chase and Meadows (2007) identify MIS 4 as a distinct transition period in the palaeoenvironmental record for southwest Africa. At around 70 kyr BP, sea surface temperatures, having gradually declined from around 22 °C at the Last Interglacial to approximately 19-20 °C, fell sharply to 15 °C and remained low throughout MIS 4 and 3 (Kirst *et al*., 1999). Fluvial inputs off the coast of Namibia also increase at the beginning of MIS 4 (Stuut *et al*., 2002). These changes correlate with terrestrial evidence for increased humidity during much of the Last Glacial period in the wider region. The various proxy indicators are not easy to interpret due to inevitable uncertainties in chronology and the fact that the sediments are preserved in a way that provides only cameos of the late Pleistocene. Nevertheless, MIS 4 appears to have been generally wetter, with charcoal data from Elands Bay Cave (Parkington *et al*., 2000) indicating a dominance of afromontane forest elements prior to 40 kyr BP. Records from a wide range of sites in the Namib Desert (reviewed by Lancaster, 2002) suggest that conditions were at their wettest immediately prior to the LGM (i.e. at ca. 35-30 and 28-24 kyr BP).

4.1 The Last Glacial Maximum

Although there are relatively few palaeotemperature records for southern Africa, those that do exist show a remarkable consistency for around the time of the LGM (Chase and Meadows, 2007). Speleothem analyses at Cango Cave (Talma and Vogel, 1992) and palaeogroundwater records from the Letlhakeng

(Kulongoski and Hilton, 2004), Stampriet (Stute and Talma, 1997) and Uitenhage (Heaton *et al.*, 1983) aquifers indicate that mean annual temperatures across the subcontinent were in the range 5.2-6 °C cooler around the time of maximum Northern Hemisphere glaciation.

In the Tswaing Crater sediments, there are indications in the ratio of organic to inorganic carbon which, coupled with mineral geochemistry, are consistent with both cooler and drier conditions (Kristin *et al.*, 2007). Precessional forcing, so prominent in the lower part of the sequence at Tswaing, becomes less pronounced in the upper sections and it is argued (Kristin *et al.*, 2007) that latitudinal shifts in the position of the ITCZ and/or changes in ocean circulation drive more pronounced incursions of winter rainfall (Chase and Meadows, 2007). Aeolian activity occurred in the linear dunes and coversands of western Zimbabwe from 32-19 kyr BP (Stokes *et al.*, 1998; Munyikwa *et al.*, 2000), and there is independent palaeoecological evidence that also indicates drier conditions between 21 and 17.5 kyr BP (Ning *et al.*, 2000). There is, however, evidence of at least one more humid phase in the Kalahari during the Last Glacial phase. Speleothem growth is reported at Gcwihaba Cave at the start of the LGM from 26-21 kyr BP (see Brook *et al.*, 1998). Highstands occurred in Palaeolake Makgadikgadi centred around 27 and 17 kyr BP (see Thomas and Shaw, 2002; Huntsman-Mapila *et al.*, 2006; Burrough *et al.*, 2007, 2009; Burrough and Thomas, 2008) and there was a seasonal lake to the southwest of the Tsodilo Hills from 27-22 and 19-12 kyr BP (Thomas *et al.*, 2003). Stromatolite growth is reported from Urwi Pan in the southwest Kalahari between 19.6-18.8 kyr BP (Lancaster, 1986), with high lake levels also documented at Alexandersfontein (Butzer, 1984). This evidence stands in contrast to the conclusions drawn from Tswaing crater sediments (Partridge *et al.*, 1999), and suggests that a simple model of cooler and drier climates for the summer rainfall region of southern Africa as a whole is no longer applicable. The lack of synchroneity in some of the records across the savannas of southern Africa may therefore result from local, rather than regional, influences on climate.

Terrestrial sediments are scarce for the winter rainfall zone during MIS 2 (Meadows and Baxter, 1999), although marine cores do supplement the meagre record. A pattern is beginning to emerge of a significantly expanded winter rainfall zone during at least some phases of the Last Glacial period, the most prominent of which occurs around the period up to and perhaps including the LGM, 32-17 kyr BP (Chase and Meadows, 2007), but poorly resolved chronologies often limit more precision. Pollen evidence from core GeoB1711-4 off western Namibia indicates that the start of MIS 2 at around 24 kyr BP was the wettest period on the west coast of southern Africa during the last 135 000 years (Shi *et al.*, 2000). At Eland's Bay Cave on the west coast of South Africa, for example, where environmental conditions are currently dominated by aridity with sporadic winter rainfall, there is persuasive pollen and charcoal evidence of a significant increase in available moisture shortly before the LGM (Parkington *et al.*, 2000). As Chase and Meadows (2007) note, however, while generally wetter than present, the LGM in southwest Africa appears to have been a climatically complex and perhaps more of a transitional and dynamic period marked by a progressive reduction in humidity prior to a drier late glacial.

4.2 The Last Glacial-Interglacial transition

Evidence from ODP site 1078 off the coast of southwest Angola suggests that vegetation in Angola changed from afroalpine scrub and open savannah during the LGM, through afromontane forest during Heinrich Event 1, to an early increase in lowland forest after 14.5 kyr BP (Dupont and Behling, 2006). This implies a gradual rise in temperature starting well before the Younger Dryas. Speleothem stable isotope evidence from Makapansgat (Holmgren *et al.*, 2003) and noble gas analyses from the Stampriet and Uitenhage aquifers (Stute and Talma, 1997) also indicate that post-glacial warming was underway across the southern African mainland by 17 kyr BP. However, warming was subsequently interrupted by an event that may, in some localities at least, be associated with the Antarctic Cold Reversal recorded in the Antarctic ice core record at Vostok (Petit *et al.*, 1999), although northern hemisphere drivers may also be important here (see Chase *et al.*, 2011). At Wonderkrater, warming seems to have commenced by around this time and is followed by a short cooling phase from 14 kyr BP (Scott, 1999).

In the Middle Kalahari of Botswana, the late glacial appears to have been characterised by generally wetter conditions. As noted above, a lake existed at Tsodilo Hills until approximately 12 kyr BP (Thomas *et al.*, 2003), and highstands are recorded in the Ngami (Burrough *et al.*, 2007), Mababe (Burrough and Thomas, 2008) and Makgadikgadi (Burrough *et al.*, 2009) depressions between approximately 18-12.5 kyr BP. Wetter conditions are also identified in the Xaudum and Okwa valleys during the late glacial (Shaw *et al.*, 1992), followed by declining groundwater levels (Thomas and Shaw, 2002) and the formation of valley calcretes (Nash and McLaren, 2003). This stands in contrast to the evidence for drier conditions at Etosha Pan in Namibia around 13 kyr BP (Brook *et al.*, 2007) and the widespread aeolian activity identified in many parts of the summer rainfall zone during the late glacial (e.g. Bateman *et al.*, 2003 Marker and Holmes 1993, Thomas *et al.,* 2002)). However, Chase (2009) argues that the enhanced aeolian activity may indicate increased wind strength rather than widespread aridity at the time.

Other widely reported global environmental changes are apparent in the record, albeit less clearly defined in this region. The Younger Dryas cooling phase is clearly evident only in very few southern African records but has now been very precisely identified in a hyrax midden pollen record from the Cederberg (Chase *et al.*, 2011).

4.3 The Holocene

While late glacial records are relatively rare in southern Africa, the Holocene is represented at a more substantial number of sites. Key events, recorded at a global scale during the Holocene, include the 8.2 kyr BP cold reversal, the mid-Holocene altithermal, the Medieval Warm Epoch and the Little Ice Age. There is evidence across southern Africa for at least some of these, although the temporal discontinuity of the record remains a significant constraint. Climate variability associated with the cooler period around 8.2 ka (Alley *et al.*, 1997) corresponds to a precessional minimum in solar radiation in southern Africa (Holmgren *et al.*, 2003). However, there appears to be no secure or well-dated evidence implicating this as a mechanism for change at any site in the region.

Cores from the floodplain of the Okavango River in northern Botswana indicate the enhanced accumulation of organic sediments at around 9.3 kyr BP (Nash *et al.*, 2006), which is thought to reflect higher rainfall levels over the Angolan catchment in the early Holocene. This is confirmed by marine records, which suggest enhanced run-off from the Angolan Highlands at this time (Gingele, 1996), by phases of shoreline building in the Makgadikgadi, Ngami and Mababe Basins (Shaw, 1985; Shaw *et al.*, 2003; Burrough *et al.*, 2007, 2009; Burrough and Thomas, 2008) and by high resolution $\delta^{13}C$ and $\delta^{15}N$ data from rock hyrax middens at Spitzkoppe, Namibia (Chase *et al.*, 2009). This latter record challenges the notion of precessional forcing of low latitude climates by identifying broad synchrony with northern and equatorial Africa records during the Holocene. It would appear that, rather than progressively wetter conditions occurring as the ITCZ migrated following the zone of maximum summer insolation, conditions at this site became progressively drier, pointing instead to high latitude northern hemisphere forcing (Chase *et al.*, 2009).

Partridge *et al.* (1999) review the evidence for the Holocene altithermal and show that maximum temperatures occurred between around 8 and 6 kyr BP. Individual sites suggest this warming may have been asynchronous in different parts of southern Africa, although the possibility that this is a consequence of inconsistent chronological control cannot be ruled out. For example, maximum temperatures in the Stampriet aquifer are recorded between around 9 and 6 kyr BP (Stute and Talma, 1997), while Heaton *et al.* (1983) place the warmer phase between 8 and 4.5 kyr BP based on speleothem isotopes at Uitenhage. The precipitation record for this period proves remarkably difficult to synthesise at a regional scale. Some summer rainfall localities, including much of the Kalahari Desert of Botswana, northern South Africa and eastern Namibia, indicate greater moisture availability at this time (see Nash *et al.*, 2006 for a review). The hyrax midden record from Spitzkoppe in Namibia, for example, indicates wetter conditions from 8.7-7.5, 6.9-6.7 and 5.6-4.9 kyr BP (Chase *et al*., 2009). However, other areas contain evidence indicative of increased aridity in the mid-Holocene, including parts of northern Namibia (Buch *et al*., 1992; Brook *et al*., 1998), the Okavango catchment in Angola (Nash *et al*., 2006), much of the (former) Transvaal and southern Zimbabwe (Partridge *et al*., 1999). High stands are recorded in the Mababe and Ngami Basins from 6.5-5 kyr BP (Burrough *et al*., 2007; Burrough and Thomas, 2008), so it is possible that conditions over Angola were wetter towards the end of the altithermal. The mid-Holocene is also notable for 20 000-year lows in levels of desert and semi-desert pollen in marine cores off Namibia, with afromontane pollen elements replaced by peaks of dry forest pollen between approximately 6.3-4.8 kyr BP (Shi *et al.,* 2000). In contrast to the summer rainfall zone, the altithermal within the winter rainfall zone appears to have been a period characterised by reduced moisture availability (Chase and Meadows, 2007) evident in pollen records and sediments from the Verlorenvlei (Meadows and Baxter, 2001) coincident with the widespread mobilisation of aeolian deposits along the western coast of South Africa as an indication of aridity, although elevated wind activity may also have been contributory.

Wetter conditions until around 3 kyr BP are indicated by pollen records from the Okavango Panhandle (Nash *et al*., 2006) and shoreline-building phases in the Ngami (Shaw, 1985, Shaw and Cooke, 1986; Shaw *et al*., 2003) and Makgadikgadi depressions (Helgren, 1984), possibly in response to wetter conditions over Angola. The rock hyrax record from Spitzkoppe in Namibia also indicates wetter conditions from 4.2-3.5 kyr BP (Chase *et al*., 2009). After this time, rainfall in the subcontinental interior appears to have increased, as indicated by speleothem development in Drotsky's and Bone caves (Burney *et al*., 1994) and lacustrine calcrete formation in the Etosha Basin (Rust *et al*., 1984). Wetter conditions are also reported from Lake Otjikoto in Namibia (Scott *et al*., 1991), the Gaap Escarpment (Butzer *et al*., 1978) and Wonderwerk Cave (Beaumont *et al*., 1984) in South Africa. However, Chase *et al*. (2009) report a marked shift towards more arid conditions at Spitzkoppe from 3.5-0.3 kyr BP, coinciding with abrupt decreases in solar activity (Solanki *et al*., 2004). In the winter rainfall zone, more humid conditions appear to have existed from 4-2 kyr BP at Klaarfontein (Meadows and Baxter, 2001).

In a review of oxygen isotope analyses of cave speleothems, shell midden mollusc remains and foraminifera in inshore marine deposits, coupled with pollen, micromammal and dendrochronological studies, Tyson and Lindesay (1992) identify a clear Little Ice Age signal on mainland southern Africa associated with widespread aridity in the summer rainfall region. Cooling is prominent between AD 1300 to 1850, with a warm episode evident between approximately 1500 and 1675. The Little Ice Age signal is particularly noticeable in the speleothem isotope geochemistry for Cold Air Cave (Holmgren *et al*., 1999). This high-resolution record for the Makapansgat Valley displays darker bands indicative of warmer and wetter conditions; the period from AD 1300 to 1800 is markedly cooler and drier, with lowest temperatures occurring around AD 1700. The rock hyrax record from Spitzkoppe (Chase *et al*., 2009) also includes a pronounced dry phase between AD 1450 and 1650. The opposite appears to be the case in the winter rainfall region. Meadows and Baxter (2001) note markedly increased regional precipitation indicated by the incursion of freshwater conditions immediately prior to colonial occupation around a coastal lake in the southwestern Cape that is consistent with wetter conditions. The proposed mechanism is consistent with the idea that the circumpolar westerlies strengthen and expand northwards during global cooler phases and, while the summer rainfall region becomes drier; the winter rainfall region becomes wetter (Tyson and Lindesay, 1998). The Medieval Warm Epoch from AD 850 to 1250 is also indicated at several localities, including Spitzkoppe (Chase *et al*., 2009).

Indications of human impacts on vegetation are widely recorded for the later Holocene, more especially for the colonial and post-colonial occupation periods (i.e. the mid-seventeenth century onwards). There has been much debate concerning the possibility of a significant pre-colonial human impact on vegetation and, given the fact that southern Africa has been the locus of anatomically modern human evolution, it is intuitively attractive to assume this to be the case. Nevertheless, unequivocal evidence for human perturbation of ecosystems prior to the arrival of European colonists has not been forthcoming, and the most prominent signs of people causing major environmental disturbance all date from the historical period. The most persuasive evidence is recorded in major vegetation shifts within the last 300 years at the Verlorenvlei coastal wetland site in South Africa (Baxter and Meadows, 1999).

5. Perspectives on the future

This chapter has illustrated that climate and climate change are, depending on the scale in question, potent forces in the shaping of southern Africa's landscapes. There are, however, increasing signs that humans are becoming major geomorphic drivers. The indications of significant human influence on southern African environments, certainly from the colonial period onwards, point to the extreme sensitivity of certain elements of the landscape (see, for example, Wiggs and Holmes, 2010). The impact of people is amplified not only by the absolute size of the population, but also through the means by which the landscape is utilised and there is evidently increasing pressure on environmental processes consequent upon, *inter alia*, globalisation and its associated effects on agriculture and industry, and global climate change.

Meadows and Thomas (2009) have highlighted how climate change may impact geomorphology in the face of projected temperature and precipitation trends. Clearly, major landscape scale features, for example inselbergs, are largely unaffected by short-term climate changes. But the widespread colluvial deposits that flank many southern African hillslopes may well be impacted, albeit that the direction and nature of the change is difficult to predict even if the future precipitation conditions are assumed. Accelerated soil erosion is certainly one potential outcome here (Meadows and Thomas, 2009).

The most obvious and pervasive geomorphic response to projected climate changes over the remainder of the twenty-first century is in the widespread reactivation of currently meta-stable longitudinal dune systems. This is illustrated graphically in the Kalahari in models developed by Thomas *et al*. (2005). According to these models, virtually any combination of temperature increase, precipitation reduction, soil moisture loss and enhanced windfield energy can be expected to result in loss of vegetation cover and re-mobilisation of sandy surface deposits across very substantial areas of southern Africa. As Meadows and Thomas (2009:271) note: "The resultant scale of impact on the agricultural and grazing economies of an enormous, vulnerable and already stressed rural poor population is incalculable, but certainly liable to be acute."

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13

Applied Geomorphology

Applied Geomorphology

Gerald G Garland and Peter J Holmes

1. Introduction

Beginning in western Europe and the USA, a significant change in focus occurred within geomorphology around the middle of the last century. The discipline changed from being primarily concerned with the description of landforms and landscapes to one which attempts to account for the origin and morphology of a landform or landscape by understanding the processes responsible for its formation, and the on-going processes which impact on it. Since it is process rather than form that is more likely to impact on human life and environment, this paved the way for geomorphology to come into its own as an applied earth science. An account of this transformation is provided by Cooke and Doornkamp (1990) who provided a number of reasons as to why geomorphology grew in importance as an applied discipline from the middle of the last century. Specifically, they accounted for the rise of geomorphology in environmental management in terms of:

- The growing importance in the study of process, as opposed to merely form (morphology);
- The growth in the number of more precise techniques for geomorphological mapping, monitoring and measurement which accompanied the increase in process studies;
- A resurgence of interest in environmental issues and the assessment, or evaluation of the Earth's surface in terms of resource availability and management. This was accompanied by a recognition of risk and hazard, which are often associated with exploitation of the biophysical environment; and
- An increase in the number of practicing geomorphologists, commencing in the early 1960s.

Today definitions of applied geomorphology are numerous. That by Hancock and Skinner (2000) is typical. They define applied geomorphology as follows:

> *Geomorphology has traditionally focused on the study of landforms and on the processes involved in their formation. Applied geomorphology is the practical application of this study to a range of environmental issues, both in terms of current problems and of future prediction. Applied geomorphology provides a strategic tool for informed decision-making in public policy development and in environmental resource management. Key areas of application include specific environmental settings, such as the coastal zone or dryland environments; the impacts of land use and management practice on earth surface processes; and areas susceptible to natural hazards.*

However, with increasing human desire to understand the Earth and the way it works, with a view to predicting its future, no area of geomorphology can logically be excluded from practical application.

The potential role of applied geomorphology in southern Africa was first described by Dixey (1962). Drawing on experience and examples from other parts of the world, and noting southern Africa's situation as a developing region with a soundly based academic geomorphology from which an applied science might develop, he reasoned that the years ahead would bring an increasing need for geomorphologists with the training and ability to contribute to the solution of engineering, environmental and development problems.

The period which followed did indeed see an increase in applied geomorphological work. The 1960s and 1970s saw significant economic growth in southern Africa, accompanied by the expansion of physical infrastructure, and a shortage of engineers. Expansion in, for example, the mining industry was at times accompanied by setbacks, and even disasters, as the physical environment was exploited. Subsidence in the dolomitic areas of the West Rand which accompanied groundwater extraction, and a series of collapse sinkholes were but one example of the type of occurrence which motivated a better understanding of geomorphic processes. However, in South Africa, by far the greatest amount of what might be described as applied geomorphological work was undertaken not by geomorphologists, but by other earth scientists, such as engineering geologists and soil scientists. Engineers also made a contribution. In the late 1970s for example Brink (1979) published the first volume of his seminal series on the *Engineering Geology of Southern Africa*.

Reynhardt (1982) ascribed the dearth of geomorphologists involved in applied work in southern Africa to the type of geomorphological training available. Typically, geomorphologists have been the product of academic geography departments often located in an arts or social science context, and it is probably true to say that many graduates lacked the skills to practice alongside their colleagues from other earth-science based disciplines. Indeed, an understanding of surface processes especially was regarded as the domain of physical geologists and engineers. Nevertheless, geomorphology's traditional niche as part of physical geography has subsequently proved to be a strength. This has become particularly apparent in recent decades as an understanding of the complexity of earth systems has grown. Geography provides a broad and holistic perspective of pertinent environmental issues, and geomorphologists have identified with this. The adage that geomorphologists confront an issue by looking outwards, while engineers and geologists focus inwards, is perhaps noteworthy. Changes in national legislation (the *South African National Environmental Management Act* (Act No. 107 of 1998) and its amendments from 1998 to 2009, and the new Environmental Impact Assessment (EIA) Regulations of 2010, now necessitate environmental impact assessments as early as the pre-planning stage for certain new developments. This has provided new opportunities for applied geomorphology. New legislation requiring practicing geomorphologists to register as natural scientists in terms of the *Natural Scientific Professions Act* (Act 27 of 2003) has also strengthened the credibility of applied geomorphology in southern Africa. It is particularly in terrain evaluation (the pre-planning phase of physical development) that geomorphologists can add value. Working alongside planners, geologists, engineers, surveyors, soil scientists and GIS practitioners, geomorphologists bring a unique perspective to bear on how the biophysical environment functions, and how it responds to human impacts. This is particularly well illustrated in a number of the case studies below.

Another driving force has emerged from the fact that the modern world is, *de facto*, increasingly run by economics, and the reductionist approach, that all worldly phenomena must in some way have a financial value, is now strongly embedded in almost all aspects of life. The last twenty years has seen this become convincingly established in broad environmental science, and especially in ecology, where *ecological economists* have developed precise approaches and techniques for assessing the monetary worth of ecological goods and services. Of note is that geomorphology has been swept up in this in a number of ways. As soon as geomorphological goods and services can be assigned hard cash values, society sees them as important, so they must be carefully and sustainably used. Ways of valuing some geomorphological materials – agricultural soil, sand and gravel and so on are well developed, but by 1996 Panizza and Fabri had conceptualised and extended the notion of geomorphological value to landforms and processes as well (Figure 13.1). They reasoned that certain landforms, if they have some intrinsic and valuable use for people, may also be assets to society, and can therefore have a financial value ascribed to them. For example, a beach is an obvious coastal landform, but if located at a coastal tourist resort it becomes a crucial asset for tourists. Income generated by tourists using the beach can be the basis for a calculation of the value of the beach.

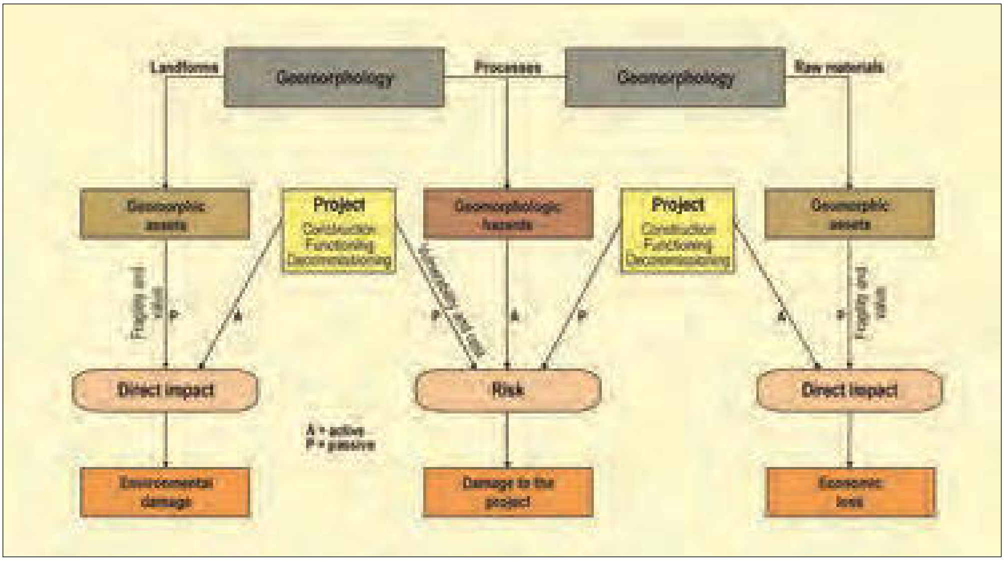

Figure 13.1. A conceptual basis for geomorphological impact assessments (after Panizza and Fabri, 1996).

Thus any use of the beach which destroys or degrades it, making it less desirable for tourist activity, could result in a loss of income from tourism. Beach geomorphologists must review and guide any such use which might impact on the natural integrity of the beach. The idea of financial geomorphological value is developed further in a later case study in this text.

Internationally, the rise in importance of applied geomorphology in the post Second World War period saw the publication of a number of texts devoted mainly, or exclusively, to the topic (Garland, 1988). These included Hails (1977), Craig and Craft (1982), Nir (1983), Verstappen (1983), Costa and Fleischer (1984), and Hart (1986). Subsequently, three important texts on applied geomorphology appeared in the space of just more than a decade (Cooke and Doornkamp, 1990; Slaymaker, 2000; Allison, 2002).

It should be noted that the boundaries between physical geology and geomorphology have, in some instances, become blurred. Whether there is any point or profit in attempting to distinguish discrete areas of expertise is questionable. Works on applied fluvial geomorphology (Thorne *et al.*, 1997) and environmental geology (Bennet and Doyle, 1997) could arguably belong to either discipline, or even to civil engineering. In addition, there have appeared fairly recently publications with an applied bent which, either intentionally or indirectly, blur the lines between traditional earth science disciplines (Gerrard, 1992; Murck *et al.*, 1997). Texts have also appeared which do not necessarily purport to represent applied geomorphology but which, effectively, make a useful contribution to a greater or lesser extent. Examples include works on soil erosion such as Morgan (1995) and, particularly with regard to southern Africa, Schmitz and Rooyani (1987) and Showers (2005).

There is no doubt that a growing demand for qualified practitioners has caused formal education in applied geomorphology to expand and develop. Many university geomorphology and geography programs offer modules in applied geomorphology, and some universities, for example the University of Sussex (UK) and the University of North Texas (USA) and others, offer complete taught Masters courses in applied geomorphology.

In southern Africa, Garland (1988) acknowledged that a review of applied geomorphology would prove difficult because much of this work is for clients, and is not published. Such work often appears in inaccessible or confidential reports. Nevertheless, referring to Verstappen (1983) he identified the main areas where applied geomorphology has played a role. These are: earth and vegetation sciences for natural resource development, environmental surveys and natural hazards, rural development and agriculture, urbanisation, transport, river and coastal engineering. Garland (1988) went on to provide a number of examples or case studies. For a review of the status of applied geomorphology in southern Africa and case studies up to the late 1980s, the reader is referred to Garland (1988). This revised text provides up to date southern African examples and case studies from the 1990s to the present. It also includes salutary lessons with regard to situations where applied geomorphological expertise could have, but was not necessarily, utilised. The case studies are geomorphological, as opposed to geotechnical or geohydrological in nature. This implies issues around environmental management and planning, rather than the detailed, site-specific cases which would typically be dealt with by geotechnical or civil engineers. The review is necessarily selective rather than comprehensive, but although some important sectors may not have been included, what follows is (with the exception of soil erosion, which enjoys a separate chapter in this book (see Chapter 11)) a fair cross section of applied geomorphological work from South Africa in particular since 1988.

2. Slope management for urban and transport development

Slopes are integral components of landforms. Indeed it could be argued that in the simplest sense landforms are nothing more than assemblages of slopes of different lengths, shapes, steepness and materials. For this reason the understanding of slopes has always been a crucial part of geomorphology. Initially the focus was on slope form and origin, but from the 1970s onward interest started to encompass slope behaviour, using analytical techniques from soil and rock mechanics developed by engineers, and rock mass strength (Moon, 1982; Selby, 1982). Infrastructural development, especially the surge of road and railway construction during the nineteenth and early twentieth centuries, together with the need for more and ever larger buildings, expansion of mining, particularly opencast mining, meant that slope failure and collapse could be catastrophic. It was necessary to all elements of slopes – form, history, origin, behaviour and the influence of extrinsic variables, especially rainfall, lithology, gradient and so on, to understand exactly how to manage them safely. This has created a well-defined role for applied geomorphology.

In South Africa, from the earliest attempts by pioneers such as Andrew Geddes Bains to construct a road linking the original colonial settlements of the Western Cape Province of South Africa to the interior in 1853, slopes have presented their fair share of challenges. Indeed, the narrow (by international standards) 3 foot 6 inch railway gauge still utilised in southern Africa was chosen partly to facilitate the construction of tighter curves than would otherwise have been possible in mountainous areas. The problematic nature of many slopes, from both a lithological and a structural perspective, thus lend themselves to applied geomorphological expertise.

By 1985 a landslide susceptibility map of southern Africa (Paige-Green, 1985) had divided the region into “landslide susceptibility zones” (Figure 13.2). It showed that, largely due to topography, lithology and climate, the east and south of South Africa and the Drakensberg Escarpment was where most slope failure could be expected.

Figure 13.2. A landslide susceptibility map for southern Africa (after Paige-Green, 1985).

Later Garland (1988) pointed out that much urban applied geomorphological work was concerned with slope behaviour. Examples included the collapse of a wall in a hornfels quarry used for oil tank housing in Cape Town in the Western Cape, and failures in Berea red sands in Durban in KwaZulu-Natal Province of South Africa (Kantey and Brink, 1985). In fact Durban, as the only major metropolitan region falling within Paige-Greens *highly susceptible* landslide zone, features frequently. This is largely due to ever-increasing construction and development on shales of the Ecca Group within the municipal boundary. To reduce risk of failure during construction, shale areas were divided into three categories:

i. Areas where no development is permitted;

ii. Zones where development may be undertaken subject to special construction techniques being observed; and

iii. Areas where normal development is permitted. Criteria for the classification include the dip of the shale, especially in relation to topographic slope; presence of clay seams; proximity of dolerite intrusions; seepage; clay overlying shale; and the nature of the proposed developments.

2.1 Case studies

2.1.1 Chapman's Peak Drive, Western Cape

Since its inception, Chapman's Peak Drive has been controversial. The road links the village of Hout Bay to the southern Cape Peninsula. It was conceived by the first administrator of the Cape Province, Sir Frederic De Waal, and built between 1915 and 1922. The road largely follows the gently undulating unconformable contact (Figure 13.3) between granite of the Cape Granite Suite, and the sandstones, siltstones and mudstones of the overlying Graafwater and Peninsula Formations of the Table Mountain Group (Theron, 1984; Geological Society of South Africa, undated).

Figure 13.3. Refurbishment of Chapman's Peak Drive in 2003. The original road follows the unconformable contact between granite (below the road) and inherently unstable Graafwater Formation mudstones and sandstones (above the road).

Intermittent, but frequent rock falls occur along this road, primarily due to this inherently unstable geology. The cliffs above the road comprise resistant sandstone beds, interdigited by thin, less resistant siltstone and mudstone beds. Differential weathering removes the less resistant strata, leaving sandstone overhangs, which fail without warning, resulting in rockfalls onto the Drive below. This problem was exacerbated in the early part of the previous decade by increasing traffic volumes, and by fires which destroyed the natural vegetation above the drive. The Drive was closed in 2000. A geomorphological study into the behaviour of the strata above Chapman's Peak drive was undertaken by Scott (2002). However, it seems the findings were not taken into account during the subsequent rehabilitation of the Drive.

From a geomorphic perspective, the inherent instability of the strata above Chapman's Peak Drive represents an insoluble problem. Attempts to rock-bar (the removal of lose material using crowbars) during 2001 proved unsuccessful, and a major hard-engineering approach was decided upon to stabilise the Drive. This included the building of half-tunnels, the emplacement of catch nets to trap falling debris, and the use of cement compounds (gunite) in an attempt to stabilise cliff faces. The road was reopened as a toll road in 2003. The stabilisation measures, while reducing the risk of rock falls, have not necessarily proved effective, as evidenced by periodic closures of the Drive, as recently as 2010, particularly when winter rains render it even more vulnerable to falling rock debris. The controversial Chapman's Peak Drive upgrade project serves as an example of a case where the contribution of available expertise towards understanding an extremely dynamic geomorphic system was possibly trivialised or overlooked.

2.1.2 Links between landslides and rainfall in Durban, KwaZulu-Natal

It was noted above that of all South African cities, Durban is the most susceptible to slope failure, being located in Paige-Green's (1985) *highly susceptible* zone. The city's subtropical climate, with warm wet summers, cool dry winters and a mean annual rainfall of between 1 000 and 1 100 mm combined with rugged topography and some highly unstable lithologies has made Durban susceptible to landslides, especially as demand for housing and other infrastructure increases. Until the 1990s most landslide analysis had focused on lithological and engineering aspects, although a link between rainfall and occurrence had been recognised by King (1982). He proposed that many landslides result from two distinct rainfall events often months apart. Water from the first event needs time to saturate the overburden sufficiently, before the second event increases pore water pressure to a critical point where failure is triggered.

King's (1982) premise motivated Garland and Olivier (1993) to undertake a thorough assessment of the relationship between rainfall and landslides in Durban. Landslide data from 1970 to 1991 were obtained from the City Engineer's Department and local press, enabling the locations of 120 landslides to be plotted with dates. These were grouped into annual and monthly landslide frequency, total landslide events, number of landslides per event, and time elapsed from September 1 (the start of the wet season).

As well as standard rainfall parameters such as maximum three day rainfall in a landslide month, maximum monthly rainfall and wet season total, a number of rainfall indices specifically designed for landslide analysis by Crozier (1986) and Guidicini and Iwasa (1977) were also used. The latter included the minimum probability threshold (PTn: the annual rainfall value below which landslides do not occur) the maximum probability threshold (PTx: the annual value above which landslides will certainly be recorded, the cycle coefficient (cumulative rainfall preceding a landslide event, divided by mean annual precipitation) and the event coefficient (rainfall of the landslide event divided by the mean annual precipitation). The last two may be summed to provide the total coefficient.

Numbers of landslides in a year correlated quite strongly with maximum monthly rainfall ($r = 0.73$) and wet season landslide totals were related to wet season rainfall ($r = 0.71$), implying a clear cause and effect link between landslides and rainfall.

The PTn and PTx probability thresholds for Durban were 860 mm and 1 225 mm respectively, and 27.5% of all landslides occurred in the probability margin (between 860 and 1 225 mm).

Perhaps more revealing were the plots of number of landslides per event against both the cycle and event coefficients. These showed that in any wet season, a four-landslide event is unlikely until the cycle coefficient exceeds 0.61 (when 61% of the average wet season rainfall has fallen) followed by an event coefficient of 0.24 (24% of the wet season average falls in a single event). This is unlikely to happen before late February.

The total coefficient plot (Figure 13.4) indicated a clear two stage equilibrium condition for the rainfall mass movement relationship, controlled by average wet season rainfall. Once average wet season rainfall in any season is exceeded multiple landslide events become much more likely. This means that once average wet season rainfall is exceeded, the number of sites which become susceptible to failure increases rapidly indicating that critical regolith moisture content has been exceeded.

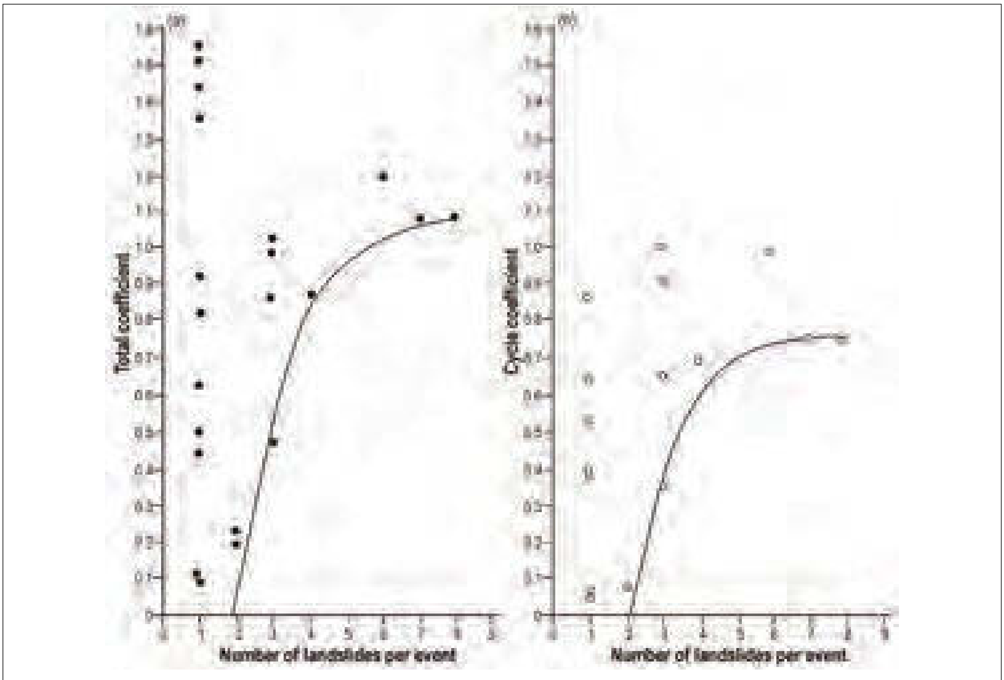

Figure 13.4. The total coefficient (a) and the cycle coefficient (b) and the number of landslides in an event (after Garland and Olivier, 1993).

3. Coastal and near-coastal zone management

With a coastline some 3 000 km in length, supporting several large ports and harbours and some of the most attractive tourist areas, infrastructural development along South Africa's coastline has burgeoned in the last two decades. Examples of this are manifold. Amongst the largest is the construction of a new port, Coega, near Port Elizabeth in the Eastern Cape Province of South Africa, but expansion of most major ports has either been recently completed or is planned. Because of their enormous cost and value, such projects cannot be contemplated without detailed understanding of coastal geomorphological processes and history.

It is now almost certain that sea level is rising around the coast of southern Africa. In an analysis of tidal gauge data, Mather *et al*. (2009) were able to confirm that in the last 40 years or so sea level off the South African coast has increased at annual rates of 3.55 mm off the east coast, 1.55 mm off the south coast, and 0.42 mm off the west coast. The potential geomorphological impacts of this can be significant along sandy, high-energy coasts. One probable result is enhanced coastal erosion potential in the medium to long term (Figure 13.5), and some specialists believe that under certain conditions 10 mm of sea level rise can result in 3-4 m of shoreline recession. Any such erosion is unlikely to take place as a slow steady annual increment, but rather in distinct surges of several metres at once following favourable combinations of storm and tidal conditions. Such an occurrence may well be followed by a period of dune and beach re-generation, but the net medium to long term response will be erosive, and it is imperative that this is taken into account in coastal construction and development planning.

Figure 13.5. Damage to the beach, property and infrastructure following a spring-high-tide storm surge, Ballito, KwaZulu-Natal, 2007.

Developments in coastal management policy have also provided opportunities for geomorphologists. South Africa's coastal policy was presented in a Government White Paper (Chief Directorate: Marine and Coastal Management; Department of Environmental Affairs and Tourism, 2000). Compliance with it requires significant geomorphological expertise as indicated in some of its goals and objectives. For example, Goal C5 (2000:74) is:

> *To plan and manage coastal development so as to avoid increasing the incidence and severity of natural hazards and to avoid exposure of people, property and economic activities to significant risk from dynamic coastal processes.*

The two stated objectives (2000:74) comprising this goal are:

- ***Objective C5.1*** *Coastal development shall be planned and managed to minimise disruption of dynamic coastal processes and to avoid exposure to significant risk from natural hazards.*
- ***Objective C5.2*** *The potential consequences of medium- and long-term climate change and associated sea-level rise shall be taken into account in all coastal planning and management***.**

The result of all these influences has led to growth in applied geomorphological studies in the past two decades, and some examples are described below.

3.1 Case studies

3.1.1 Hout Bay – Sandy Bay Dune Field, Western Cape

Headland bypass dunefields are a feature of a number of South African coastal environments. The Hout Bay Dunefield was previously the most prominent such dunefield in the Cape Peninsula. From the middle of the last century, the impacts of development in the village of Hout Bay, as well as the encroachment of alien *acacia spp*., have effectively destroyed the functioning of this dunefield. Holmes and Luger (1996) undertook an applied geomorphic study of the dunefield and calculated sediment transfer rates between Hout Bay and Sandy Bay. They were able to demonstrate a net sediment gain for Sandy Bay, given an unimpeded aeolian sediment transfer corridor. Stabilisation of the dunefield as a direct result of urban expansion, and alien vegetation encroachment, has resulted in its destruction as a functioning geomorphic system. Holmes and Luger (1996) predicted that, starved of a wind-blown sand supply to nourish its beach, Sandy Bay could revert to a boulder beach. Further applied work serves to support this prediction (MacHutchon, 2012).

3.1.2 Knysna Coversands, Western Cape

Although the presence of Neogene-Quaternary aeolian sands in the Knysna Basin has previously been recognised (see, for example, Partridge and Maud, 1987), mapping of these sands was only undertaken in the 1980s by the Saasveld (Department of Forestry) Research Group. These maps were not published. Detailed mapping was undertaken a decade ago (Marker and Holmes, 2002), who proposed the nomenclature *coversands* (Figure 13.6). These authors also described the peculiar geomorphic properties of the coversands, and their propensity to erode, particularly where vegetation is removed and the sands, which typically overlie a less permeable substrate, become saturated. The engineering properties (Atterberg limits) of the Knysna Coversands were subsequently described (Marker and Holmes, 2005). The Knysna area is undergoing rapid development, and a failure to recognise the potential erodability of the coversands resulted in severe erosion of slopes mantled by coversands upstream of the Knysna Estuary. This occurred particularly on the original Simola Golf Estate development, which has subsequently been rehabilitated. More cautious development at what is today the Pezula Estate (originally Sparrebosch) on the Eastern Head paid dividends, in that this golf course did not suffer extensive erosion, in spite of the occurrence of coversands on the Knysna Heads.

Figure 13.6. Deep exposure of Knysna coversand unconformably overlying a thin band of Knysna Formation lignite (arrowed) which in turn unconformably overlies Table Mountain Group quartzite.

3.1.3 Traditional fish trapping in Kosi Bay Estuary, KwaZulu-Natal

Green *et al.* (2006) report on an unusual situation in Kosi Bay Estuary in Greater St Lucia Wetland Park, in KwaZulu-Natal, just south of the Mozambique border. The estuary comprises a series of four interconnected tidal lakes supporting small mangrove communities (Figure 13.7). The total surface area of the wetlands and surrounds is ~240 000 ha. Tidal level within the lakes varies considerably with distance from the mouth, but in all cases is dominant over fresh water inflow, due to the very small catchment area of some 55 000 ha.

Figure 13.7. The location of mangrove stands and fish traps in Kosi Bay (after Green *et al.*, 2006).

Although recently powerboats have resulted in bank erosion and shallowing of the interlinking channels, a more contentious issue is the existence and expansion of an indigenous subsistence fishery. Fish from Kosi Bay have been the most significant source of protein for the local community for some centuries, and the shallowness of the lakes has allowed a unique fish trapping system to evolve. Traps comprise fences constructed of brushwood and wooden stakes embedded into the lake floor, laid out in such a way as to allow fish to pass between them into the lake on the rising tide, but trap them on the ebb tide as they return to the sea. The number of fish traps was relatively constant at between 60 and 100 from 1942 till the late 90's. At that point socio-economic conditions resulted in a surge in numbers so that by 2000 they had reached in excess of 180.

The effects of traps on sedimentation in the estuary were brought into focus as long ago as the mid-twentieth century (Broekhuisen and Taylor, 1956), and 46 years later Wright (2002) postulated that the traps were responsible for modification of the tidal prism resulting in shallowing of the system, with a gradual increase in mangrove area and expansion of the supra-tidal area within the lakes. If correct this would alter the tidal nature of Kosi Bay and in all probability destroy the indigenous trap fishery.

Green *et al.*'s (2006) geomorphological analysis used five sequences of air photos from 1942 to 2000 to map and measure mangroves, supratidal areas and fish traps. A weak correlation was found between fish trap numbers and mangrove expansion, suggesting that if fish trapping expansion rates continued unchanged, the system would be completely choked with mangrove by 2500. This however is unlikely to happen. Catastrophic flooding of the system occurs in a twenty year cycle, and when floods occur significant mangrove die off takes place, and sediments are eroded. In addition, rising sea level, now a firmly established fact off this coast (Mather *et al.*, 2009), would reduce supratidal exposure and mangrove development. The study concluded that at least from a sedimentation perspective, the fishery and its expansion is sustainable.

4. Planning and land management

At both the macro- and the mesoscale, morphological mapping can prove valuable in the process of terrain evaluation. In its simplest terms, this entails assessing a terrain (landscape) in terms of its inherent characteristics. This can be indirectly construed as a catalogue of potential opportunities and constraints which that terrain might offer for development or conservation. From a geomorphic perspective, for South Africa as a whole, Kruger (1983) produced a small-scale map of terrain morphological units covering the whole country (see Chapter 14) while Partridge *et al*. (2010) produced a detailed physiographic subdivision for earth and environmental scientists classifying the geomorphic provinces of South Africa, Lesotho and Swaziland (see Chapter 14).

4.1 Case Study

4.1.1 Table Mountain, Western Cape

The Cape Peninsula Protected Natural Environment (CPPNE) was a forerunner of the Table Mountain National Park. As part of a broader strategy to properly manage the CPPNE, Van Wierengen and Holmes (1994) undertook preliminary terrain mapping and evaluation of the physical environment. Eleven different land facets were identified within the CPPNE. This classification was undertaken on the basis of inherent geological, geomorphological (primarily slope) and surficial debris or soil properties, as well as climatic and geohydrological characteristics. The sensitivity of each facet to various human activities was then assessed. This provided a base for future considerations with regard to potential land use within the CPPNE. The above serves as an example of applied geomorphology in the context of potential land use management.

5. River and catchment management

Fluvial geomorphology is one of the best developed sub-disciplines in geomorphology, and has long been a crucial tool in managing rivers. Water resources, flood control and river navigability all depend on an understanding of the geomorphological behaviour of rivers.

5.1 Case Studies

5.1.1 Kleinmond Golf Course, Western Cape

A decision to divert the Middle River to flow through a recently reconstructed golf course in the seaside town of Kleinmond proved problematic when severe rain during the winter of 1990 (more than the entire June monthly average in one event) destroyed a number of small dams on the river diversion (Wesseman, 1991). In spite of commissioning a specialist geomorphic and botanical report (unpubl.), which recommended the return of the river to its original course so that the area could be rehabilitated, this was not done. Severe rains the following winter subsequently resulted in further damage (Figure 13.8), washing away bridge and road infrastructure. Subsequently, gabions were built to stabilise the damaged channel. Nevertheless, had the proffered recommendation been followed in the summer of 1991, the situation could, in part, have been avoided.

Figure 13.8. Kleinmond Golf Course (1991) damage to a wetland and infrastructure.

5.1.2 Geomorphological classification for the management of river ecosystems

Amongst other things, the *South African National Water Act* of 1998, recognised the importance of the ecological health of rivers, and the need to define an "Ecological Reserve". Rowntree *et al.* (2000), note that this relates to the quality and quantity of water necessary to protect the sustainable functioning of aquatic ecosystems of a particular water resource. They pointed out that the *Act* stopped short of defining the ecological reserve, and based on the concept of longitudinal zonation developed in the 1960s and 1970s by river ecologists derived and tested a river classification system appropriate for the ecological reserve based on geomorphological principles.

After significant research they noted that characteristics of river reach bore a strong relationship to river gradient as measured on the 1:50 000 topographic map series. This is particularly useful as it reduces significantly the need for field work.

Using this approach they identified seven zones, each with a characteristic gradient and channel features, with an additional three zones specific to rejuvenated rivers (Table 13.1). These were tested against field evaluations for the Sabie, Olifants and Buffalo Rivers, which showed that the zone descriptions corresponded well to the field evidence. However, the authors note that the system will develop further as rivers in different areas are included. The zonation is currently in use by the Department of Water Affairs and Forestry under the National River Health Programme.

Table 13.1. Geomorphological zonation of river channels (after Rowntree *et al.*, 2000).

ZONE	ZONE CLASS	GRADIENT CLASS	CHARACTERISTIC CHANNEL FEATURES
A. Zonation associated with a "normal" profile			
Source zone	S	not specified	Low gradient, upland plateau or upland basin able to store water. Spongy or peaty hydromorphic soils.
Mountain-head water stream	A	>0.1	A very steep gradient stream dominated by vertical flow over bedrock with waterfalls and plunge pools. Normally first or second order. Reach types include bedrock fall and cascades.
Mountain stream	B	0.04 – 0.099	Steep gradient stream dominated by bedrock and boulders, locally cobble or coarse gravels in pools. Reach types include cascasdes, bedrock fall, step-pool, Approximate equal distribution of "vertical" and "horizontal" flow components.
Transitional	C	0.02 – 0.039	Moderately steep stream dominated by bedrock or boulder. Reach types include plane bed, pool-rapid or pool-riffle. Confined or semi-confined valley floor with limited floodplain development.
Upper foothills	D	0.005 – 0.019	Moderately steep, cobble-bed or mixed bedrock-cobble bed channel, with plane bed, pool-riffle or pool-rapid reach types. Length of pools and riffles/rapids similar. Narrow floodplain of sand, gravel or cobble often present.
Lower foothills	E	0.001 – 0.005	Lower gradient mixed-bed alluvial channel with sand and gravel dominating the bed, locally may be bedrock controlled. Reach types typically include pool-riffle or pool-rapid, sand bars common in pools. Pools of significantly greater extent than rapids or riffles. Floodplain often present.
Lowland river	F	0.0001 – 0.0009	Low gradient alluvial fine bed channel, typically regime reach type. May be confined, but fully developed meandering pattern within a distinct floodplain develops in unconfined reaches where there is an increased silt content in bed or banks.
B. Additional zones associated with a rejuvenated profile			
Rejuvenated bedrock fall / cascades	Ar, Br or Cr	>0.02	Moderate to steep gradient, confined channel (gorge) resulting from uplift in the middle to lower reaches of the long profile, limited lateral development of alluvial features, reach types include bedrock fall, cascades and pool-rapid.
Rejuvenated foothills	Dr or Er	0.001 - 0.019	Steepened section within middle reaches of the river caused by uplift, often within or downstream of gorge; characteristics similar to foothills (gravel/cobble bed rivers with pool-riffle/ pool-rapid morphology) but of a higher order. A compound channel is often present with an active channel contained within a macro channel activated only during infrequent flood events. A limited floodplain may be present between the active and macro-channel.
Upland floodplain	Fr	<0.005	An upland low gradient channel, often associated with uplifted plateau areas as occur beneath the eastern escarpment.

5.1.3 Economics, landforms and the Disa River catchment, Western Province

The 3.55 km^2 catchment of the Disa River in Orangekloof Nature Reserve, Hout Bay, plays a crucial role in flood protection and attenuation for the built up area downstream. Were it impacted, for example by development and catchment hardening, an engineered flood protection scheme would have to be constructed to buffer the downstream impacts of storm water. By conceptualising the catchment as an assemblage of landforms, Midgely and Garland (1997) used the SCS hydrological model to compute runoff rates for the catchment from the twenty year return period storm of 113 mm hr^{-1} with antecedent soil moisture contents from 20-80% for three different cover conditions: the current vegetated state, and two different residential development layouts, each with a cover factor of 38%. The cost of an engineered storm protection structure capable of dealing with the maximum expected runoff rate under these conditions of 27 m^3 sec^{-1} was calculated at R27 million at 1995 prices. This represents the investment required to protect downstream areas from flooding, and is the substitution value of the landforms in the catchment.

Figure 13.9. The lower catchment of the Disa River, Hout Bay, showing encroaching development on the floodplain (image courtesy of Google Earth).

6. Conclusion

Since Garland's (1988) review, applied geomorphology in South Africa, though not necessarily southern Africa, has continued to grow in application and breadth. This review has presented case studies based largely on the authors' own work and experience, and is by no means exhaustive. The examples merely serve to illustrate the sort of work that geomorphologists are undertaking within the applied realm in southern Africa. Although many projects are still linked closely with engineering, construction and development, economics and changes in national policy have broadened the areas of application. This is most evident in the wider environmental sphere, where development and strengthening not only of the *South African National Environmental Management Act of 1988*, but also the National Coastal Policy, and the *South African National Water Act of 1988* make the point that certain elements of the environment cannot be successfully and sustainably managed unless landforms and processes are taken into account. In addition, new trends in assessing the financial value of landforms and processes to society have also started to make an impact. These have all created enhanced, clear and significant roles for geomorphologists within environmental impact assessment, sustainable resource management and risk and hazard assessment, and will continue to do so in the foreseeable future.

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14

Landscape Inventories and Remote Sensing

Landscape Inventories and Remote Sensing

Frank D Eckardt, Charles H Barker and Michael E Meadows

1. Introduction

The concept of a landscape inventory is used increasingly to refer to digital information about landscapes at various scales, from local to global. At its very essence, however, a landscape inventory is a product of data collection and classification and its most frequently illustrated output is a map. The mode of data acquisition has certainly evolved in recent years with the development of remote sensing technologies, but the fundamental questions that geomorphologists ask about the features of the land surface have remained the same, *viz*. how do we inventorise, classify, monitor, analyse and model the morphological characteristics of landscapes and the processes that form them? While today we might use a term such as Geographic Information Technology to describe our attempts to answer such questions, in the past this may have been simply described as "geomorphological mapping". In this chapter, we explore the history of efforts to map the southern African landscape from a geomorphological perspective and illustrate how recent technological developments have facilitated a more resolved picture of landscape dynamics at various geographic scales. In so doing we highlight the availability of remote sensing products in particular that can facilitate further investigations into the geomorphology of southern Africa by researchers, professionals and students alike. The development and application of evolving Geographic Information Technologies are explored in the creation of landscape inventories in southern Africa and consider the application of these tools to the understanding of geomorphic dynamics into the future. Many easily accessible and generally free products are now available for download that enable users to develop their own interpretations of the southern African landscape and there is an attempt to highlight some of the more useful of these. We begin with a consideration of geomorphological mapping before exploring the remote sensing tools that have revolutionised the production of such maps and facilitated an ever-increasing array of applications.

2. Representing geomorphology on maps: a brief history

Geomorphological mapping provides geomorphologists with the ability to identify and analyse the various surficial forms of the landscape (Pavlopoulos *et al.*, 2009, Smith *et al.*, 2011). As with all other maps, a geomorphological map is the result of a complex analytical process whereby the compiler graphically indicates how phenomena, processes and relationships are distributed in space (Gellert, 1972). The regional approach to landform mapping is promoted by Baker (1986) as he states that it provides explanations for the variety in landforms related to structure, material, processes and age.

Gellert (1972:15) defines geomorphological maps as "... concerned with the relief of the continents and the bottom of the oceans as the boundary between the solid body of the Earth (lithosphere) and its fluid and gaseous envelopes (hydrosphere and atmosphere)." A geomorphological map may contain information about morphometry, morphography, hydrography, lithology, structure, age and process (Gustavsson *et al.,* 2006). From this, it becomes evident that a geomorphological map may have a number of purposes ranging from the development of theoretical geomorphology to the application of scientific geomorphological knowledge in practical applications (Gellert, 1972). Scale is the determining factor in any type of mapping and no less so in relation to geomorphic features. It dictates what can be mapped, the area to be mapped and the amount of fieldwork that will be necessary (Cooke and Doornkamp, 1990) to ground-truth the data. Gustavsson (2005) states that since the Babylonian mapmakers started to depict the surface of the Earth, several methods were used to indicate the location of prominent landforms. The "mound method" together with symbols for vegetation and hydrography were used on maps of southern Africa until the middle of the nineteenth century when hachured maps gradually appeared (see Norwich, 1993; Duminy, 2011). The first contour lines captured with plane table surveys were drawn by the British Royal Engineers during the Anglo-Boer war of 1899 to 1902 (Liebenberg, 1973). Topographic maps using photogrammetric methods from terrestrial photos were published in 1934 at a scale of 1:25 000, but plane table surveys were still used for the first 1:50 000 map published in 1935 (Liebenberg, 1973). Only after 1948, with the acquisition of stereo plotters by the government Department of Surveys and Mapping, were topographic maps published on a regular basis (Republic of South Africa (RSA), 2010).

Of course, an indication of elevation alone does not provide geomorphologists with all the necessary information needed to map landscapes. The first comprehensive system for geomorphological mapping was published under the auspices of the International Geographical Union in 1972 by Jaromir Demek as editor. A comprehensive legend for geomorphological maps is included in the publication (Demek, 1972). Since 1970, "new" developments such as computer technology, satellite and sensor technology including radar instruments, together with improved aerial photography provided geomorphologists with advanced tools in data analysis and automated mapping (Pavlopoulos *et al.,* 2009). From the 1990s onwards, Geographic Information System (GIS), Global Positioning System (GPS) and digital photography have all been added to the list, all under the collective term Geographic Information Technology.

3. Landscape inventories in southern Africa

The earliest southern African map depicting physical environmental features appears to be that constructed by Passarge in 1908 (Figure 14.1) and divided the region into three major zones, *viz.* coastal plains (*Küstenvorland*), terraced slopes (*Stufenländer*) and an interior of plateaus and basins (*Hochflächen und Beckenlanschaften*); these three units were the further subdivided into to seven different *natural geographic regions*. This map retained its currency for many decades and suggests acceptance of the importance of landscape and its relationship to underlying structure (for example Taljaard's 1948 monograph which, in outlining the geology of the country, makes reference to geomorphological features as being directly related to the different rock types and structure). Du Toit's (1939) map of the physiographic regions of South Africa indicates nine distinctive regions, namely: Kalahari, Karoo, the Basotho Highlands, the Cape Fold Belt, an Upland region (in turn subdivided on either sides of the main watershed (i.e. the Great Escarpment), a Tertiary peneplain, the Lebombo belt, a Coastal region and the Urema *sunklands* of Mozambique (Figure 14.2).

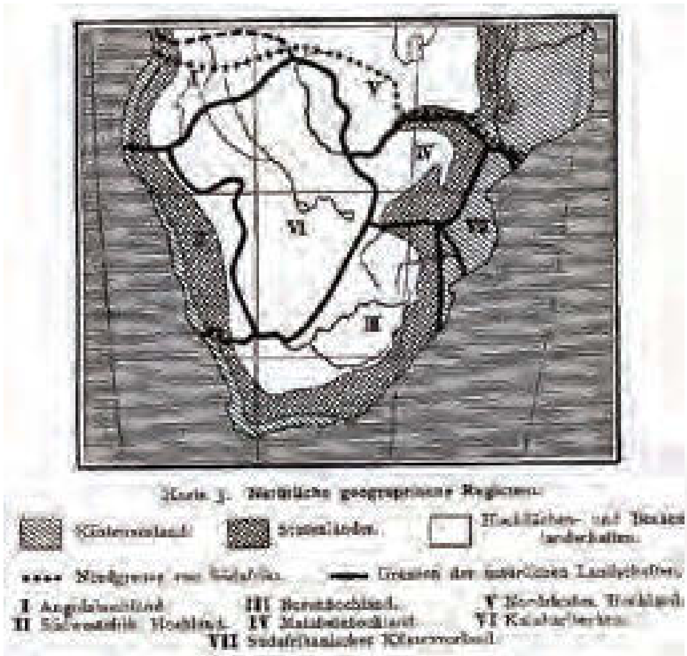

Figure 14.1. Natural geographic regions of southern Africa (after Passarge, 1908).

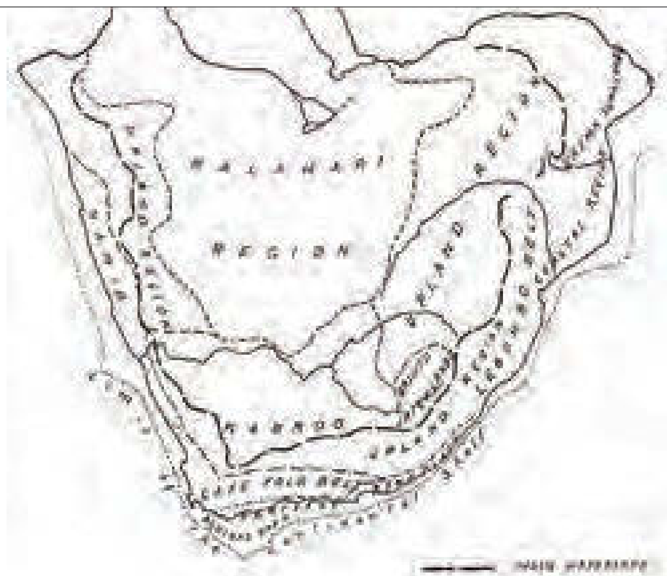

Figure 14.2. Du Toit's physiographical regions of southern Africa (after Du Toit, 1939).

In developing these ideas, Wellington (1955) further delineated the physiography of the sub-continent (Figure 14.3) and King (1963) went on to recognise no less than 18 geomorphological provinces (Figure 14.4), albeit loosely based on Wellington's physiographic regions. Kruger (1983) mapped 12 regions (termed *terrain patterns*) and provided a detailed legend for the map that identified six broad morphological classes (Figure 14.5). In turn these were divided into a further 30 subdivisions on the basis of slope form, relief, drainage density, stream frequency and percentage area with slopes <5%. The original map on a 1:2 500 000 scale was subsequently published in digital format by the then Department of Environmental Affairs and Tourism as part of the Environmental Potential Atlas of South Africa (ENPAT) (DEAT, 2001). Similar attempts include those by Van Zyl (1985) in which there are 30 major landform regions. Drescher and De Frey's (2009) geomorphological map of the country was based on Hammond's (1954) relief, slope and profile, Dikau's (1991) landform classification and Morgan and Lesh's (2005) methodology and identifies 16 terrain types. More recent national geomorphological maps are based on digital technology associated with new remote sensing products (see below). Partridge *et al.,* (2010) utilised Digital Elevation Models (DTM) to create longitudinal profiles and valley cross sections of selected main rivers. These were then used to determine zones of homogeneity called *macro-reaches* which in turn resulted in revision of the boundaries of King's (1963) classification resulting in 34 *provinces* and 12 *sub-provinces* (Figure 14.6).

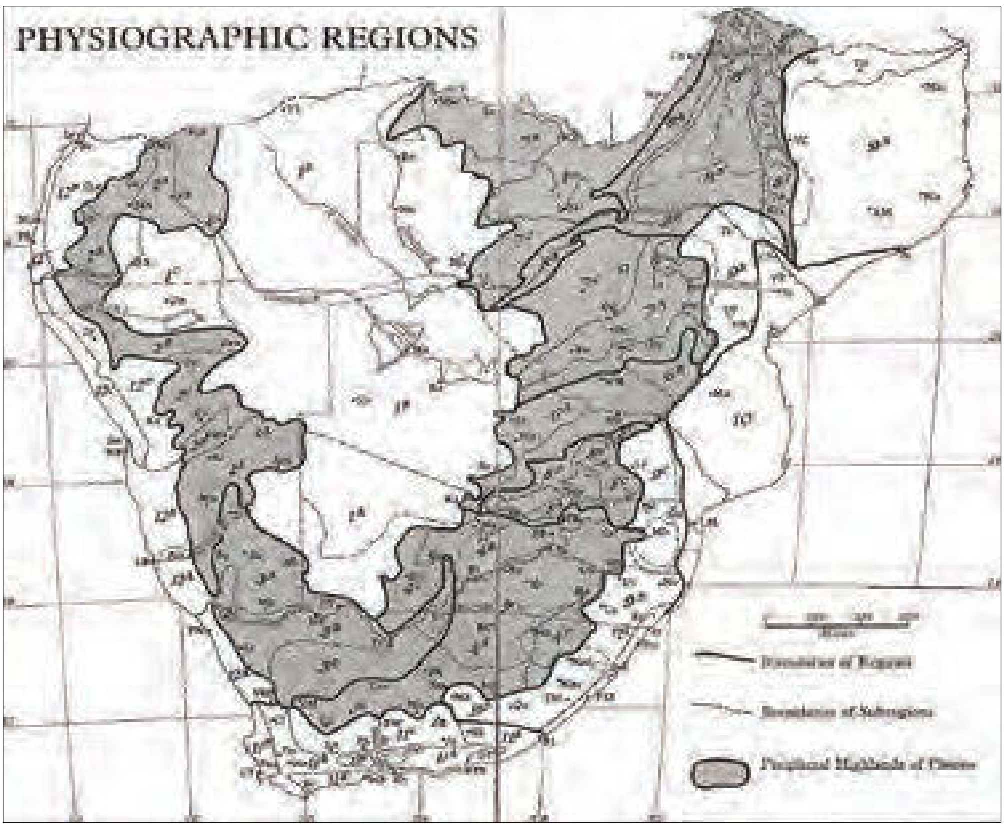

Figure 14.3. Wellington's physiographic regions (after Wellington, 1955).

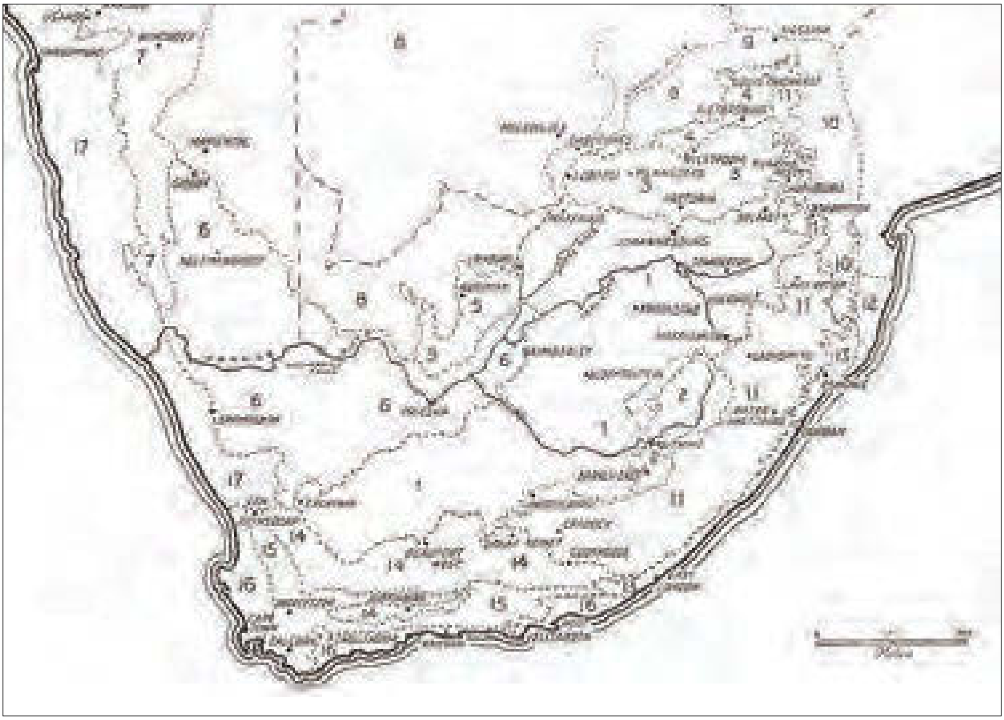

Figure 14.4. King's geomorphological provinces (after King, 1963).

Figure 14.5. Kruger's broad morphological patterns (adapted from DEAT, 2001).

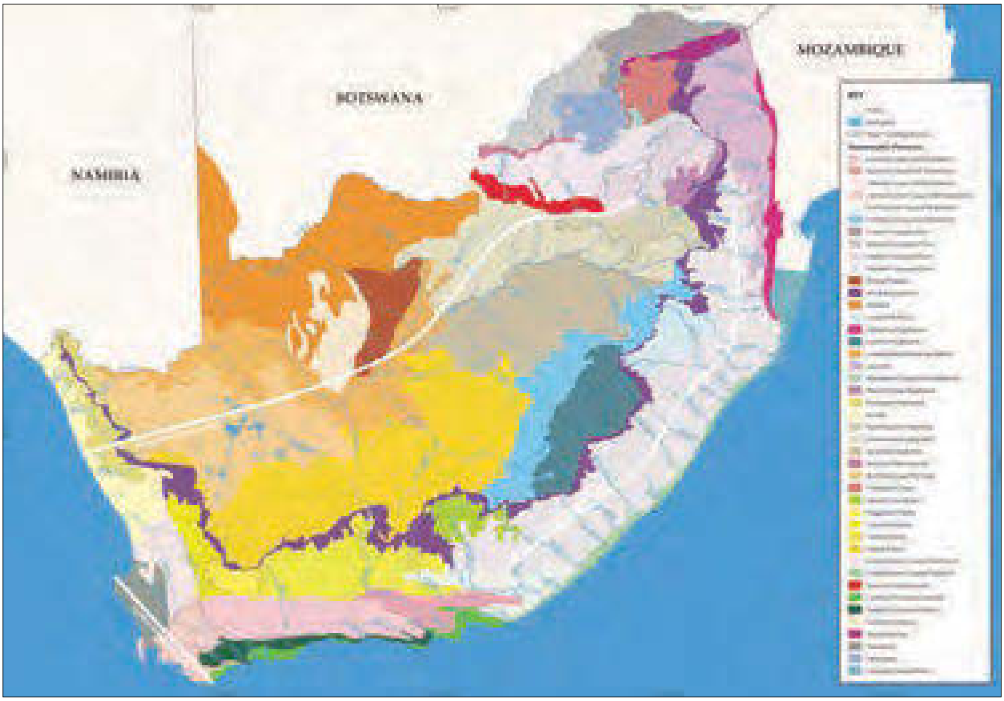

Figure 14.6. Geomorphic provinces of South Africa, Lesotho and Swaziland (after Partridge *et al.,* 2010).

4. Geoinformation technology

4.1 Introduction

The largely descriptive and classificatory methods that informed most of the maps referred to above have been significantly augmented in the recent past through the development of data capture and manipulation technologies involving GIS. De Mers (2009) defines GIS as an information system with the capability to collect, store, manipulate and analyse data and present information regarding georeferenced objects on or close to the surface of the Earth. This has obviously been facilitated by the emergence of widely available remote sensing (RS) products that involve the acquisition of data about an object without being in physical contact with the object being studied (Jensen, 2007). Mesev and Walrath (2007) integrate GIS and RS as Geoinformation Technology (GIT) with many obvious applications in geomorphology.

In a GIS the printed map is replaced by a computer, a database and the software to provide a user with access to the information. A GIS does not eliminate the shortcomings of maps, but it can reduce certain limiting factors of conventional cartography through an effective, well-developed database structure. The ease of integrating and representing different types of information is a great improvement on conventional maps (DeMers, 2009). Vast amounts of spatial and attribute data can be saved and easily updated. This "raw" data is not subjectively generalised and can thus be used in more than one application. Without doubt, aside from the evolving technical ability of computers that resulted in smaller and smaller processors being able to deal with larger and larger amounts of data, arguably the key developments for geomorphological mapping have been in the realms of remote sensing products. Smith and Pain (2009) have recently summarised data and potential applications available to geomorphologists.

GIT can of course be applied to model various processes related to landforms. Since the publication of an edited volume on the monitoring, modelling and analysis of landforms by Lane *et al.,* (1998), GIT has been central in various efforts to apply the technology in geomorphic research. Fieldwork and data collection, together with data analysis, are a restricting factor (in terms of time and cost) in geomorphological research. With the advent of powerful desktop computers and accompanying software, this has changed. Geomorphologists can use the technology to their advantage in collecting, modelling and simulating processes and landforms and the advent of remote sensing products may justifiably be claimed to have revolutionised the science and art of constructing landscape inventories in South Africa and globally.

4.2 Application of remote sensing

Remote sensing has come a long way since the first astronauts acquired a few opportunistic photos from low Earth orbit. There has been a tremendous improvement in spatial detail, repeat acquisition, as well as accessibility of data, including freely available raw imagery and the online creation of virtual viewers such as Google Earth. Here we provide a brief historic summary of major space borne Earth observation sensors and data, which provide global coverage and are freely available to southern African geomorphologists. In addition we provide examples of southern African remote sensing applications of geomorphological relevance.

Table 14.1. Useful remote sensing data websites.

TYPE OF DATA	DESCRIPTION	WEBLINK
Image Data	Global Land Cover Facility	http://www.glcf.umd.edu/
Image Data	Landsat Data	http://landsat.usgs.gov/
Image Data	MODIS Imagery	http://rapidfire.sci.gsfc.nasa.gov/
Elevation Data	ETOPO (5 ,2 and 1 arc min)	http://www.ngdc.noaa.gov/mgg/global/global.html
Elevation Data	GTOPO	http://eros.usgs.gov/products/elevation/gtopo30/gtopo30.html
Elevation Data	SRTM (30, 3 arc second)	http://srtm.csi.cgiar.org/
Elevation Data	ASTER GDEM (3 arc second)	http://www.gdem.aster.ersdac.or.jp/
Map of Human Impact	Human Footprint	http://www.wcs.org/humanfootprint/
Photography	NASA Astronaut Photography	http://earth.jsc.nasa.gov/
Vegetation Data	NDVI Products	http://earlywarning.usgs.gov/fews/
Southern African Airborne Data	SAFARI 2000 Campaign	http://daac.ornl.gov/S2K/safari.shtml
South African Data	National Geo-spatial information	http://www.ngi.gov.za
South African Satellite Data	CSIR data catalogue	http://www.csir.co.za/SAC/catalogue.html
Online Geomorphology Book	NASA Geomorphology from Space	http://disc.gsfc.nasa.gov/geomorphology/
Online Atlas of Global Change	UNEP Atlas of Global Change:	http://na.unep.net/atlas/onePlanetManyPeople/book.php
Online Atlas of African Change	UNEP Atlas of African Change:	http://www.unep.org/dewa/africa/africaAtlas/
News and Current Observations	NASA Earth Observatory	http://earthobservatory.nasa.gov/
Global Landforms	Landform Classification	http://eusoils.jrc.ec.europa.eu/website/SRTMTerrain/viewer.htm

Over the last few decades global terrestrial remote sensing has played a pivotal role in monitoring the dynamic cryosphere, kept tabs on tropical deforestation and in general been witness to an ever increasing human footprint (Sanderson *et al.,* 2002). In geomorphological terms, southern Africa can be considered relatively stable. A 1985 publication on Global Geomorphology (Short and Blair, 1985) stressed the cratonic nature of southern Africa with emphasis on the Kaapvaal Craton, the Damara Orogen and the Great Dyke of Zimbabwe. This is in contrast to tectonically active areas where rapid image acquisition and swift repeat coverage is required as, for example, in assessing the impact of the 2004 Indian Ocean Tsunami, and landslide induced lakes and overspill flooding in Pakistan (see http://earthobservatory.nasa.gov/NaturalHazards/view.php?id=44144). Such speedy assessment of high-resolution imagery is made possible by numerous, competitive, commercial operators, which routinely produce data at a spatial resolution of less than 1 m. Such data may be used to assess, among other features of geomorphological significance, the impact of recent flooding in Thailand (see http://earthobservatory.nasa.gov/NaturalHazards/view.php?id=76932). Table 14.1 lists some useful websites that facilitate generally free access a wide range of currently available products.

4.3 Remote sensing platforms

Very high spatial resolution modern data are a far cry from the first routine Earth observation missions such as Landsat 1, with its Multispectral Scanner (MSS) launched in July 1972. The MSS (http://landsat.gsfc.nasa.gov/about/history.html) which produced data in four visible and near infrared channels, had a spatial resolution of only 80 m. Early observations included extensive southern African coverage (Short 1976). Landsat 1-3 utilised MSS, which was later replaced with the Thematic Mapper (TM) in 1982 (Landsat 4). The TM is an instrument with seven channels at a spatial resolution of 30 m and, compared to MSS, includes additional mid- and thermal infrared channels. As a result of the growing awareness of global change, deteriorating magnetic MSS data tapes were restored thus allowing for decadal global surface change observations. Retrieval, storage and distribution of more than 30 000 geo-corrected Landsat images by the Global Land Cover Facility (GLGF; http://glcf.umiacs.umd.edu/), for example, contributed to the United Nations Environmental Programme's Atlases of Global Change: (http://na.unep.net/atlas/onePlanetManyPeople/book.php), and of African Change: (http://www.unep.org/dewa/africa/africaAtlas/). Landsat 6 failed to reach orbit and Landsat 7 was soon plagued by technical issues, although its Enhanced TM (ETM+) sensor, with an additional 15 m panchromatic channel, was able to complete global coverage early in the new millennium, essential for the purpose of producing a systematic global image database for the 70's, 80's and late 90's to 2000 period. In March 2009, the TM on Landsat 5 celebrated its 25th year of operation. It is probably fair to say that much of the modern terrestrial remote sensing community was trained using the instrument from this platform. While numerous other sensors have undoubtedly produced a variety data, none of these are as accessible, pre-processed and appropriate in geomorphological work as the Landsat coverage (http://earthexplorer.usgs.gov/).

4.4 Remote sensing and the environment

Population growth and development captured in the decadal time series using MSS, TM and ETM+ manifests itself in urban and rural expansion, mining, forestry, irrigation, agriculture, ranching, impoundment of rivers and conservation and management. All of these are clearly depicted at the global scale as well as, in the specific South African context, in relation to risk and vulnerability to global change (DEAT, 2010). These observed changes have many indirect geomorphic manifestations and, collectively, they influence processes of landscape denudation, for example accelerated soil erosion. Direct endogenic and exogenic changes are especially evident in, *inter alia*, the shrinking icecaps on Kilimanjaro (Tanzania) and Rwenzori (Uganda) and the lava flows on Mt Nyiragongo (DRC) that catastrophically impacted the town of Goma in 2002 (Favalli *et al.,* 2006) – all are clearly captured in satellite imagery. The recent landslide at Mt Elgon in Uganda in 2010 was imaged from the (Earth Observer 1) EO-1 (http://earthobservatory.nasa.gov/IOTD/view.php?id=43130).

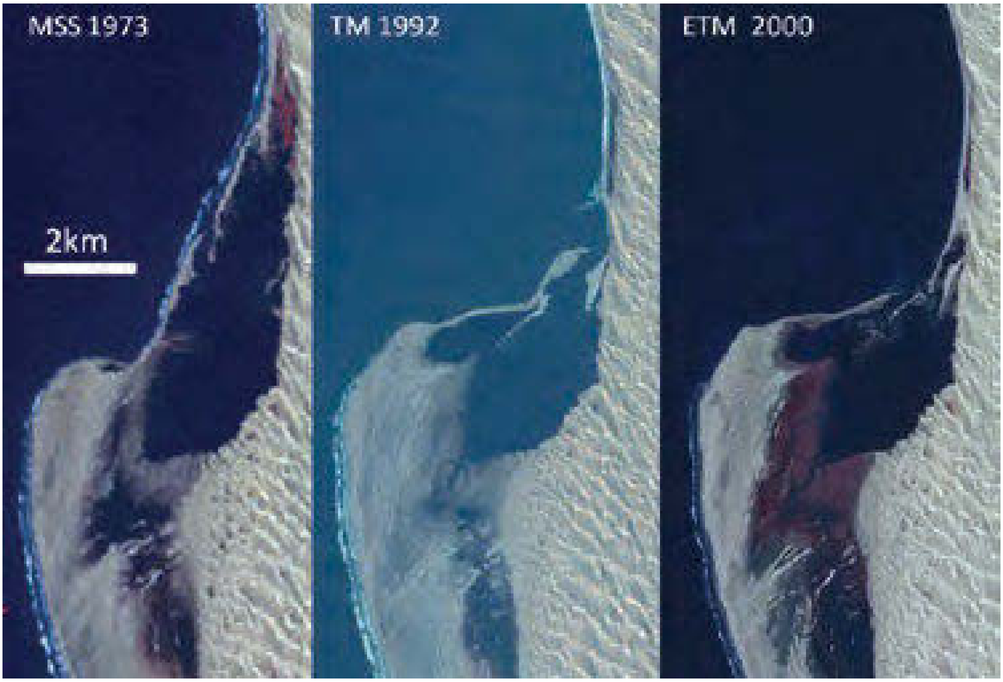

Figure 14.7. Example of data from the Global Land Cover Facility depicting coastal change at Sandwich Harbour Namibia using Landsat MSS (1973) 80 m resolution, TM (1992) 30 m resolution and ETM (2000) data 30 m resolution. False Colour Depiction of Bands 432 as RGB.

A time series generated from GLCF data and of direct geomorphologic interest is illustrated in Figure 14.7, which shows the effect of long shore drift at Sandwich Harbour, Namibia. This lagoon was frequented by North American whalers in the 1800's and has gradually been filled up and closed as observed by Wilkinson *et al.,* (1989) using handheld shuttle photography. In addition, MSS imagery also captured the Boteti Wetland during its last period of peak flooding, which resulted in Lake Xau and the artificially created Mopipi Dam (Figure 14.8a). It has taken more than 20 years for the Boteti to return to Mopipi (Figure 14.8b), which was reached again in September 2010. This example underlines the importance of long term monitoring and archival of data, if we are to capture and understand the changing land surface at the global and regional scale.

There is a great deal of interest in very high-resolution (<1 m) data, but for appropriate change detection there is a requirement for older data. While MSS is often heralded as the start of space-borne remote sensing, military data was captured several decades prior to that. Data from the space-borne reconnaissance programme, under the name of CORONA (Ruffner, 2005), covering the period of July 1959 to July 1973, was declassified in 1995. The so-called *Keyhole* satellite series produced photographic data at a resolution of 7.5 to 1.8 m which is not only available to the public, but scanned and geocoded negatives can produce panchromatic high resolution data comparable to the latest sensors in orbit. The data are surprisingly global in extent and, for example, facilitated the production of a detailed image mosaic of the Okavango Delta and its channels (see Figure 14.9, Hamandawana *et al.,* 2007) and high resolution monitoring of land use change and impacts in the region (Hamandawana and Chanda, 2010). For smaller, more site-specific change studies, aerial photography can be equally useful and extends considerably further back in time. Such photography, while variable in quality and scale, has demonstrated the decline in bare sand cover for various dune systems in the Western Cape (Holmes and Luger, 1996) following stabilisation of sand through processes of urbanisation and alien tree infestation (Figure 14.10).

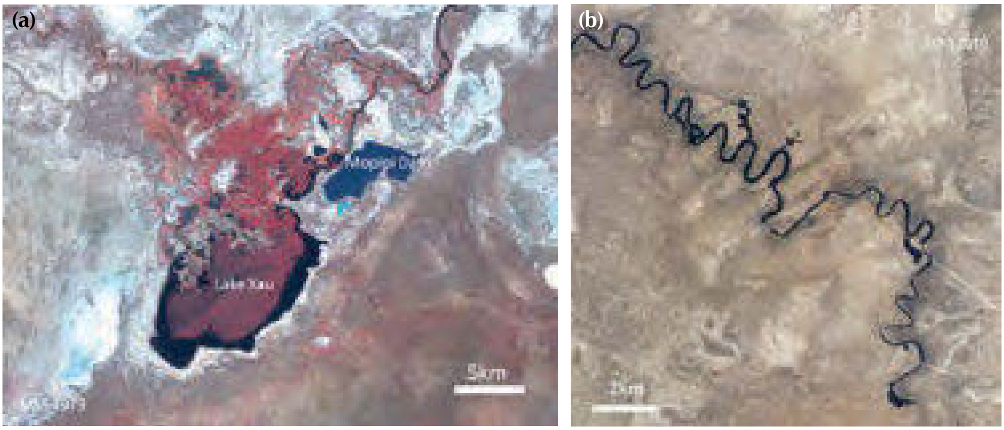

Figure 14.8. a) The Lower Boteti and Lake Xau in 1979 (MSS) 80 m resolution (RGB 421) and return of flood waters 20 years since last flow in 2010 as seen from EO-1. b) In true colour. (Source: GLCF and Nasa Earth Observatory.)

Figure 14.9. Image subset of declassified reconnaissance image from the Corona series depicting parts of the Okavango Delta in 1967 in a panchromatic black and white photo (Hamandawana *et al.*, 2007).

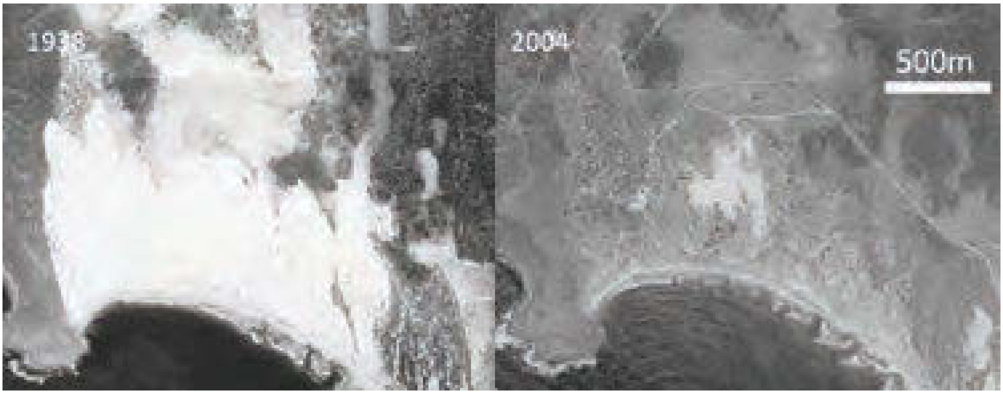

Figure 14.10. Historic aerial photography depicting stabilisation of headland bypass dunes by encroaching vegetation cover. Tabak Bay 1938 and 2004 (Surveys and Mapping, Mowbray).

For the purpose of mapping geological and geomorphological features, Landsat data is available free of charge (http://landsat.usgs.gov/) and remains extremely useful. Channels 5 (1.55-1.75 um) and 7 (2.08-2.35 um) in particular, on both TM and ETM+, respond to subtle mineralogical surface contrast. This allows for better distinction of underlying lithology and soil types. For example, hydrated evaporites halite becomes readily identifiable; the data have enabled the establishment of the distribution and geochemical evolution of playas and provides convincing evidence of ongoing gypsum deflation in these systems (Eckardt *et al.,* 2001a). The available imagery has also revealed sub-dune Damara lithology on Kalahari Dunes as a bi-product of biological overturning by termites (McFarlane *et al.,* 2005).

Modern, high-resolution sensors, including (*Satellite Pour l'Observation de la Terre)* SPOT 5 and Pleiades 1 do not carry equivalent mid- and far-infrared channels as such detectors require dedicated cooling, which makes such satellite platforms both bulky and costly. Most contemporary land surface sensors simply carry visible and near infrared channels. While such images yield a high spatial resolution, and render the surface in true colour, they also tend to cover much smaller areas at a much larger cost and do not feature all wavebands needed. The data vacuum created by the demise of Landsat 7 has, as yet, not been filled. Data continuity including appropriate mid infrared coverage is to be achieved with future Landsat 8 (http://ldcm.nasa.gov) and 9 (http://www.landimaging.gov).

4.5 Digital elevation data and models

While it is possible to infer land surface morphology and topography from an inherently two-dimensional satellite image, mapping and examining terrain in elevation data represents a very logical extension and improvement in the study of landforms. During the last decade, global topographic elevation or Digital Elevation Model (DEM or Digital Terrain Model, DTM) data have created new user communities and products.

Initially available global-scale relief digital elevation products including, for example, ETOPO was derived from ground-based (and ship-based) data, which acted as precursors to a host of space derived products. ETOPO 5 has a 5-km resolution, although it includes seafloor bathymetry. GTOPO improved to a 1-km resolution, but showed significant data variability due to unequal mapping quality. SRTM (Shuttle Radar Topography Mission of NASA), available at 90-m resolution globally, suffers from various artefacts and data voids due to topographic radar shadowing and adverse surface properties. Relative accuracy is around 2 m, but absolute accuracy for southern Africa has been reported closer to 5 m (Rodriguez *et al.*, 2006). Nevertheless it is possible to map features in such subtle landscapes as the Kalahari in great detail. Earlier national mapping efforts were entirely based on Landsat MSS hardcopies. The morphology of dunes and shorelines is captured equally in both products, although actual height of features and contours could only be derived from SRTM data (Figure 14.11). Makgadikgadi shorelines at the higher 1 000 m contour have been mapped using this technique as well as the tectonic modification of Kalahari dune ridges along the Gumare fault (McFarlane and Eckardt, 2007). SRTM data can be sub sampled to 50-m resolution (Groham and Steiner, 2008) and 25-m data is available for selected areas through the German Aerospace Center (DLR) portal (Rabus *et al.*, 2008). There are several SRTM datasets making use of improved void filling techniques as well as a better definition of the coastlines, but fundamentally still draw on the original SRTM data acquired in 2000 and released in 2003 (http://srtm.csi.cgiar.org/).

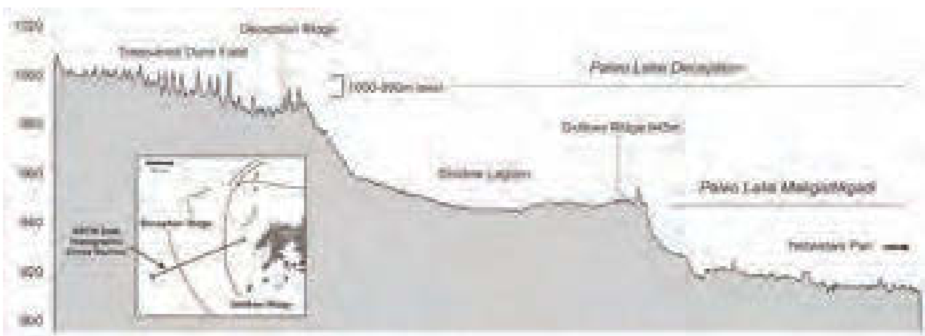

Figure 14.11. 120 km long topographic profile across the western rim of the Makgadikgadi margin captured in SRTM 90 m resolution data. It depicts two shorelines at 945 m (Gidikwe Ridge) and 1,000-990 m at (Deception Ridge) formed in response to past lake levels. Current dry lake level is at 900 m (McFarlane and Eckardt, 2008).

The technicalities in such applications are not trivial. A critical step in the extraction of parameters for geomorphological modelling is the issue of accuracy (Barker, 2011). Desmet (1997:5) distinguishes between DEM that represent any phenomenon with a unique *z* value as a surface area, and digital terrain models that represent ground height above a datum. Burrough and McDonnell (1998) define both terms as a quantitative model of a part of the Earth's surface area in digital format. As noted by Barker (2002) a digital terrain model may be defined as a raster representation of a land area in three dimensions, with *x* and *y* co-ordinates that determine location and the *z* co-ordinate representing elevation of the land area above average sea level. Employing a DEM generally entails the following tasks (Weibel and Heller, 1991:270), *viz*. construction (data collection and model construction), manipulation (modify and refine), interpretation, visualisation and applications or modelling. There are clear advantages to this over traditional methods including speed of access, greater degree of objectivity and repeatability. Incorporating GIS means that the information can also readily be adjusted and analysed.

The selection of data sources is also critical for the quality of a DEM (Weibel and Heller, 1991). The first step in the DEM construction process is collecting altitude data and additional information on phenomena that affect the shape of the terrain surface. The Chief Director: National Geo-spatial Information currently has available a DEM of the whole of South Africa at a resolution of 25 m (Republic of South Africa (RSA), 2011). The data were captured using photogrammetric methods and from vectorised contour lines.

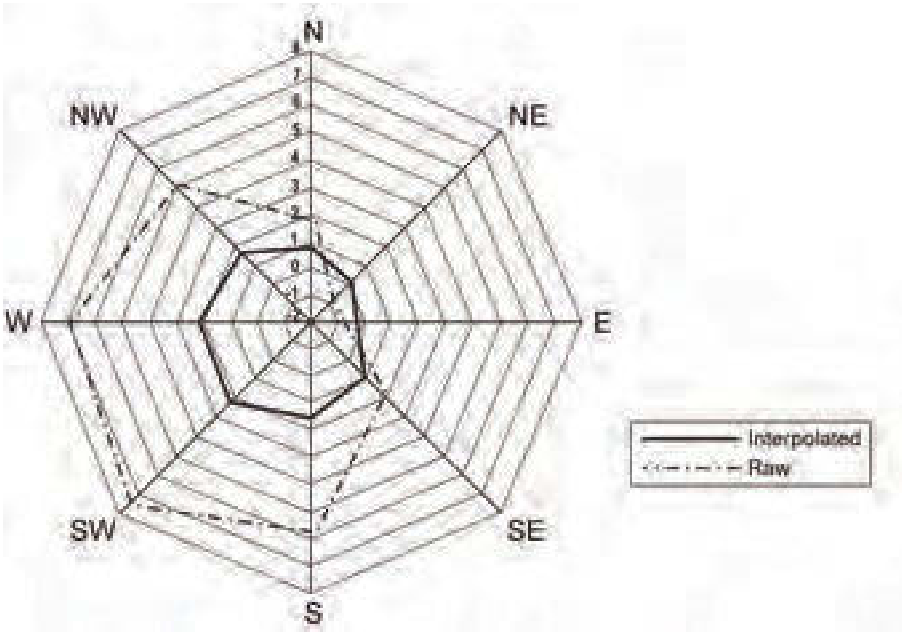

Figure 14.12. The relationship between raw SRTM and re-interpolated altitude values and aspect (after Barker, 2009a).

In an investigation of the suitability of SRTM data for topographical modelling, Barker (2011) suggests a combination of the SRTM data and the published relief data available for South Africa. The process entails the conversion of both contour lines and the SRTM data to points and the re-interpolation of the dataset. One of the advantages of this approach is improved accuracy, since the SRTM datasets not only have areas with *no data* (usually in water bodies), but also show prominent differences in altitude on slopes with a western to southwestern aspect in the central Free State. Figure 14.12 presents the deviation from the mapped altitude (in metres) of raw SRTM data and a re-interpolated surface relative to the aspect of slopes.

Global data coverage has since seen the addition of ASTER GDEM, (http://www.gdem.aster.ersdac.or.jp/) with a 30-m resolution. It has been reported to yield greater topographic accuracy; unfortunately the data set is also very noisy. Although there are no major voids, there are numerous artefacts, introduced by automated photogrammetric methods, applied to over one million *Advanced Spaceborne Thermal Emission and Reflection* (ASTER) image stereo pairs. The ASTER *Global Digital Elevation Map (*GDEM) does manage to fill such obvious data holes as the Namib Sand Sea (Figure 14.13), which was a clear omission in SRTM data, but it is still widely considered a research product in need of significant improvement. Both ASTER and SRTM lend themselves to the production of two-dimensional topographic long profiles such as that for the Molopo River talweg (Figure 14.14). Elevation data are also ideal for the detection of some of Africa's very subtle topography, including large fan shaped areas, termed mega fans, which give some indication as to how inland basins form and change over time (Wilkinson *et al*., 2009; Miller *et al*., 2010). TerraSAR-X and TanDEM-X currently produce digital terrain data of the highest quality with a 12-m spatial resolution, but data is currently only available through commercial channels.

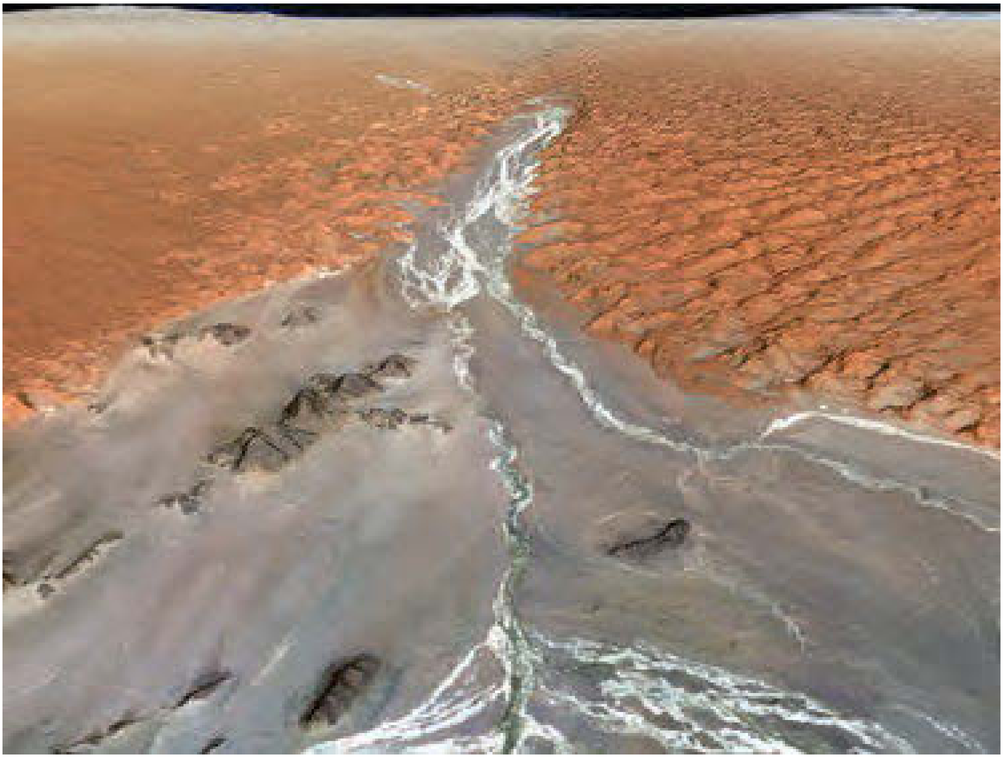

Figure 14.13. Example of ASTER GDEM data for Sossus Vlei, Namibia with Landsat 7 image drape. Rendered view towards Atlantic coast, depicts Sossus River, Inselbergs and 300 m high dunefield of the Namib Sand Sea. This area was not covered by SRTM data. (Source: ASTER GDEM and GLCF.)

Figure 14.14. Molopo River long profile and associated knickpoints. Steep knickpoint above Orange River confluence indicative of active stream profile rejuvenation.

The greatest elevation accuracy from space is currently obtained from the Ice, Cloud, and land Elevation Satellite (ICEsat) laser altimeter. This instrument produces point height data for approximately 40 m ellipses at 250 m post-spacing. While the primary objective is to obtain accurate dynamic volume estimates for the Greenland and Antarctic Icecaps, it is also possible to obtain accurate height data for the rest of the terrestrial world; including tree canopy height and associated global forest cover biomass. Laser altimetry only covers the area directly beneath the satellites pass and, as such, is not available in a rasterised format. Validation experiments on salt lakes in South America produced an absolute vertical accuracy of greater than 2 cm (Fricker *et al.,* 2005). Comparing ICEsat against SRTM and ASTER GDEM for the area in and around the Makgadikgadi Basin (Figure 14.15) reveals that SRTM overestimates height by approximately 5 m while ASTER underestimates by less than 2 m, when compared to 65 000 laser-derived point heights. However, the analysis also indicates greater data noise for ASTER, as noted above.

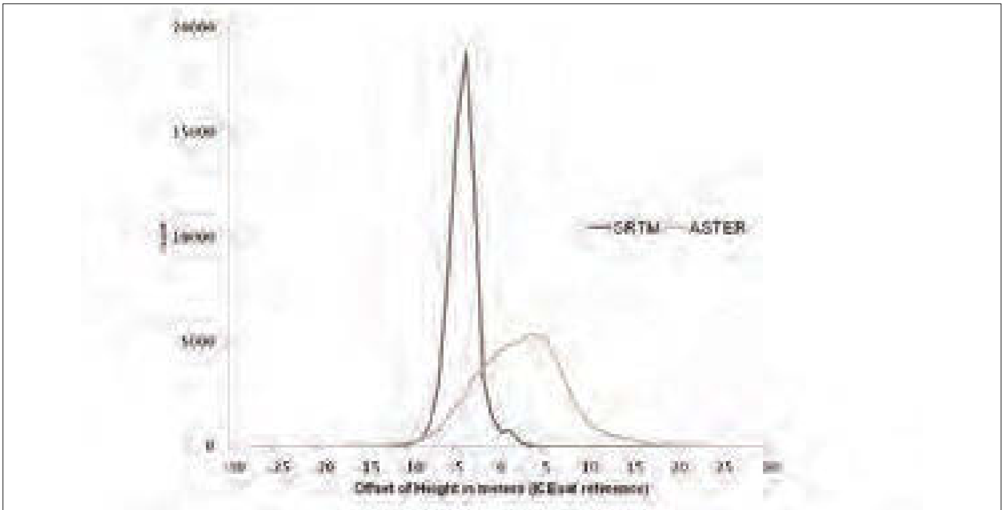

Figure 14.15. Comparing ASTER and SRTM elevation data against 65,000 ICEs at laser point heights in the Makgadikgadi Pan surface and basin (altitudinal range is less than 90 m). Note that SRTM data underestimates height by approximately 5 m, while ASTER data overestimates by about 2 m. Also note wider data spread for ASTER data which is indicative of noise and artefacts.

In addition to identifying features, elevation data can also be used to classify the landscape using a variety of land surface parameters (Hengel *et al.,* 2008). These methods are partly automated and have produced global geomorphology maps. When examining these maps alongside the geological maps of Japan and the continental United States for example it appears that surface morphology as extracted from elevation data tallies well with geological features (Iwahashi and Pike, 2007), although a detailed analysis for southern Africa is yet to be realised.

4.6 Analysis and interpretation of changing environments

The cratonic landscapes of southern Africa are largely the result of ancient denudation meaning that much of the contemporary landscape is relatively *quiet*. Drainage incision is responding to sea level change, the escarpment is retreating slowly, and there is some incipient rifting as well as intra-continental plate movement. Mantle plume processes may be one of the few drivers for epeirogenic topographic adjustment. Gravity maps from the Gravity Recovery and Climate Experiment (GRACE) mission represent an attempt to place some of the mega-geomorphology into its deeper context (Tapley *et al.,* 2004). A follow up mission Gravity field and steady-state Ocean Circulation Explorer (GOCE) is currently attempting to add spatial detail to gravity retrieval and also enhance our global handle on dynamic gravity, capturing hydrological and cryogenic dynamics with subtle fluctuations in mass. The mass of water from regional catchments, such as the Okavango River catchment, could be recorded before they are manifested in Okavango Delta flooding and thus facilitate flood management.

It is clear that the geomorphology of the southern African lithosphere is remarkably stable and, as a result, landscape changes are slow and largely (fortunately in the human context) not catastrophic. However, surface processes involving interactions between the lithosphere, atmosphere, hydrosphere and, of course, biosphere are still highly variable due to the subtropical and semi-arid conditions. The Meteosat satellite series has recorded Africa's weather, producing several images per hour since 1977, and in the process regularly depicts convective systems as part of the Zaire Air Boundary, Tropical Temperate Troughs and the Angolan low as well as rain associated with westerly cold fronts and cut-off lows. Atmospheric moisture availability is driven by a complex range of parameters, including the El Niño Southern Oscillation (ENSO) so that monitoring African rain would be impossible without a range of instruments covering the world's oceans, in particular the Pacific and Indian Oceans. In turn, vegetation responds to rainfall variability, which is expressed using Normalized Difference Vegetation Index (NDVI), a standardized ratio that makes use of the opposing vegetation response in the red and infrared waveband regions and is used routinely as a drought indicator while also assisting in producing crop estimates at the micro- to mesoscale using both SPOT and Landsat imagery and at the regional and global scale using lower-resolution data from *Advanced Very High Resolution Radiometer* (AVHRR) (1 km) and Moderate Resolution Imaging Spectroradiometer (MODIS) (250 m).

The daily imagery at thermal wavelengths from the AVHRR (since 1981) and MODIS (since 2000) allows for the detection of global fires, which are a seasonal response to the drying of biomass coupled with anthropogenic activity. Southern African fires and their atmospheric emissions were subject to a detailed investigation during the SAFARI 2000 campaign (Swap *et al.,* 2003), which produced a host of airborne imagery in particular NASA ER-2 data from hyper spectral Airborne MODIS simulator sensor (AMS) and the RC-10 camera (http://daac.ornl.gov/S2K/safari.shtml) that has yet to be fully exploited.

In addition to drought cycles over the last 10 years, MODIS has kept track of Mozambican flooding in the lower Zambezi and Limpopo during the February 2000 landfall of Hurricane Eline (Reason and Keibel, 2004) and also captured flooding in the upper Zambezi during the 2009 period. Coarse spatial resolution sensors such as AVHRR and MODIS also kept track of the hydrodynamics of Etosha (Bryant, 2003) and the Okavango Delta (McCarthy *et al.,* 2003), both globally significant wetlands. The expression of extreme inter-annual and inter-seasonal rainfall variability requires detailed monitoring of such wetland systems as; in the dry season (winter) these may be major sources of atmospheric dust. Global mineral aerosols maps such as Total Ozone Mapping Satellite (TOMS); (Figure 14.16) have pinpointed some of the world's major dust sources in areas of recent or ongoing flooding (Engelstaedter and Washington, 2007). Dryland pans not only act as short-lived lacustrine systems, of importance to semi-arid flora and fauna, but their decay also serves mineral aerosol production. Closer examination of plumes using MODIS true colour and *Meteosat Second Generation* (MSG) pink dust detection, sensitive to the thermal properties of suspended dust, suggests that source areas are associated with both ground and surface water dynamics in selected regions of the pans. This would indicate that many of the southern African playas, especially the large pans, are supply-limited systems, which only deflate during periods of sediment availability derived by flooding, or evaporation (Vickery, 2010) and possibly crust production. This is in contrast to the Bodele depression in Chad (Washington *et al.,* 2009), which consists of a perpetually exposed, ancient lake floor that is ground up and mobilised by saltating sand grains. On any occasion, when there is sufficient wind, plumes hundreds of km in length become evident. The Bodele is therefore a transport-limited dust source, which ranks highest among the dustiest places on Earth. Aside from the major pans (Bryant *et al.* 2007), there are other areas in southern Africa prone to deflation, including west coast river mouths and coastal pans between Cape Town and as far north as the Cunene. These sources are most evident in Sea-viewing Wide Field-of-view Sensor (SeaWIFS) and MODIS imagery (Eckardt *et al.,* 2001b, Vickery, 2010) as well as in the earlier shuttle photography (Eckardt *et al.*, 2000).

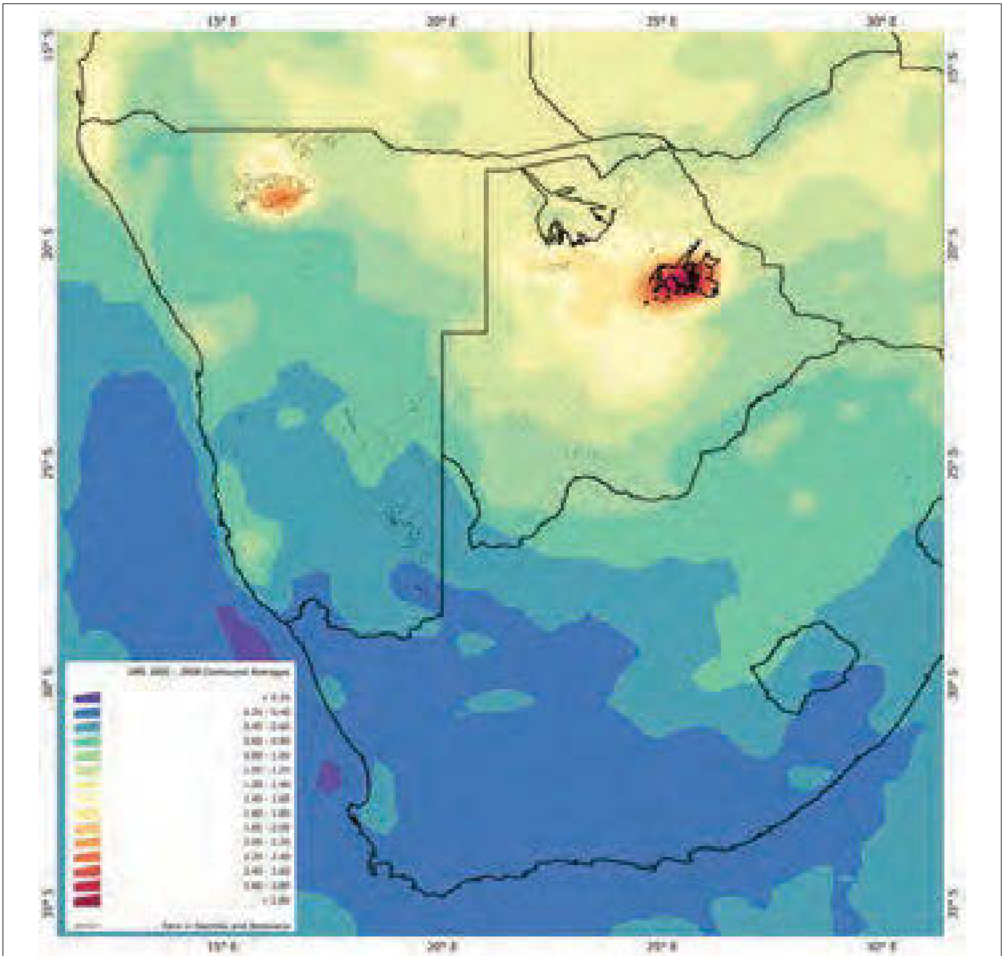

Figure 14.16. TOMS Total Ozone Mapping Spectrometer (TOMS) Ozone Monitoring Instrument (OMI) view of southern African dust sources such as the large Etosha and Makgadikgadi Pans stand out (after Vickery, 2010).

Radio detection and ranging (RADAR) imagery as well as light detection and ranging (LiDAR) data is important in establishing surface roughness as well as height and deformation (Lillesand *et al.,* 2004) but is often acquired with a specific user and application in mind and nowhere near as accessible and global in extend as passive sensors presented up to this point. Radar imagery has been used successfully in the southern African. While not strictly geomorphological in nature, *European remote sensing satellite* (ERS) -based interferometry examined crustal deformation of Lesotho highlands with the emplacement of Katse Dam (Doyle *et al.*, 2001a). The same technique was used to examine the surface deformation at Welkom in response to a mining induced earthquake in April 1999 (Doyle *et al.*, 2001b). Radar data has been used in an attempt to identify the shallow burial of drainage channels in the Namib (Lancaster *et al.,* 2000) and also to examine the buried portions of the Roter Kamm impact structure (Grant *et al.,* 1997).

4.7 Remote sensing prospects

Fifty years after the first human space flight a total of more than one million photos have been placed online and are freely available to those with appropriate technology and bandwidth. Earlier Skylab photography (NASA, 1977) was used to map global sand seas, but today's astronauts with the latest digital cameras operating behind distortion free windows, are able to take photographs under a wide range of lighting conditions and viewing angles and capture short-lived phenomena often undetected by existing orbiting instrumentation. Systematic mapping, however, requires the establishment of detailed databases, which draw on a range of geo-coded products. A good example is the recent establishment of the Namib Sand Sea Atlas (Livingstone *et al.,* 2010), which is a collation of all available geospatial data.

This review may give the impression that US sensors are the only systems providing data, but this is far from the truth. South Africa, for example, has SumbandilaSat in operation, which is starting to generate data for the national user community (Mostert *et al.,* 2008). It is significant that South Africa has a vibrant mining and exploration sector and, while it appears that hyper-spectral and LiDAR instruments are frequently flown, much of the data are handled as confidential and the wider academic community is excluded from access. Recent airborne campaigns in the KNP saw the dedicated deployment of hyper-spectral and LiDAR data for the purpose of mapping vegetation communities including bush encroachment and the geomorphological applications of this application are obvious if, as yet, unexplored (Wessels *et al.,* 2010).

Although many other nations, such as Russia, Japan, India and Brazil and consortia (for example the European Space Agency, ESA) produce data, their respective ground segment priorities place unequal emphasis on making the processed data available. Data acquisition channels are in place but are often not user-friendly or are overly expensive. South Africa, however, has the added advantage that it is home to the Satellite Application Centre (SAC) run by the Council for Scientific and Industrial Research (CSIR) (see http://www.csir.co.za/SAC/). Not only does this actively receive a wide range of satellite data, made accessible to the national remote sensing community, but in order to circumnavigate internet speeds and data volume costs the SAC recently made raw image data directly available to approximately 20 academic institutions in the country. As part of the data democracy initiative, the CSIR has distributed the so-called Fundisa disk, a 700 GB (gigabyte) external hard disc with up-to-date national geospatial raster and vector datasets including 600 SPOT5 scenes acquired for South Africa in 2008. It is an extraordinary facility and one that should continue to reap benefits for South African researchers in the years to come.

5. Spatial analyses and modelling in geomorphology

So, the data potential is enormous, but what about applications? Apart from fundamental mapping as discussed previously, geomorphologists must attempt to model geomorphic processes and landforms using the GIT toolkits available in southern Africa. The volumes by Doornkamp and King (1971) and Chorley (1972) were landmarks in demonstrating the use, not only of statistics in geomorphological research, but also in showing that there are relationships between physical objects and also between objects and their environment that require special methods of analysis. Following the introduction of accessible GIS on personal computers, there has been a veritable explosion of work that indicates the potential of spatial analysis in geomorphology (see, for example, Bonham-Carter, 1994; Lane *et al.,* 1998; Burrough & McDonnell, 1998; Dikau and Saurer, 1999; De Barry, 2004; and De Smith *et al.,* 2009.

In scientific terms, a model is described as a simplification of reality but, regardless of its complexity, is always incorrect in some way (Thorn, 1988). Pelletier (2008) elucidates the complexity of geomorphological systems and the possible role that analytical and numerical modelling can play in an understanding of the possible evolution of landscapes. It is beyond the scope of this chapter to explore geomorphic modelling in any detail, but helpful reviews may be found in Darby and Van de Wiel (2003) who provide a succinct discussion on models in fluvial geomorphology, while Odoni and Lane's (2011) approach is somewhat broader. The use of GIS to analyse and simulate dynamic processes can also be included under the label of modelling (Longley *et al.*, 2011).

5.1 Case study: the Modder River

How is it possible to understand the evolution of geomorphological features through the application of technology and modelling? In this brief description of work done on the Modder River it is possible to see the potential of combining remote sensing methodologies with modelling and statistical analysis to explore the evolution of a drainage network. The Modder River in the central Free State Province of South Africa has an interesting drainage structure in that the western part of the catchment is largely endorheic (Figure 14.17), while the eastern part shows a typical dendritic drainage pattern (Barker, 2002). There have been various hypotheses regarding the development of this drainage pattern in this area (see Helgren, 1979; Partridge and Maud, 1987; Marshall, 1988; Myburgh, 1997), but no consensus as to how such an unusual pattern could evolve. In a detailed investigation of factors influencing landscape development in a fluvial system, Barker (2002) identified 24 parameters classified into five types (Table 14.2). Basin parameters were classified using factor analysis, the basins grouped together via cluster analyses and the relationships among the various parameters investigated through correlation and regression analyses (Figure 14.18).

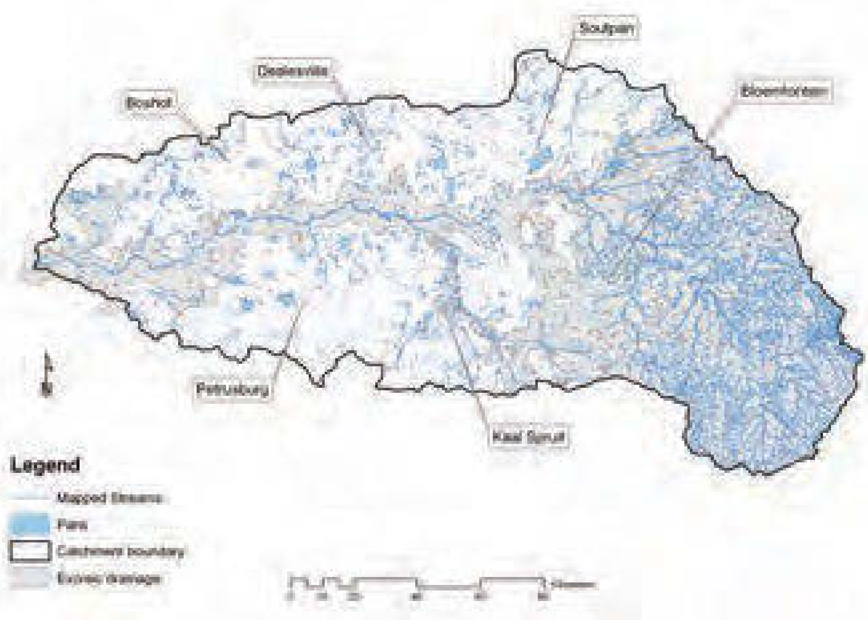

Figure 14.17. Exoreic drainage in the Modder River catchment related to mapped streams and pans (after Barker, 2002).

Table 14.2. Parameters used in the Modder River study (Barker, 2002).

TYPE	PARAMETER	SYMBOL
Stream	Number of streams	COUNT
	Drainage density	DD
	Stream length	S_LENGTH
	Flow length	F_LENGTH
Basin	Length	B_LENGTH
	Perimeter	PERIMETER
	Area	HECTARES
	Gradient (°)	SLDEG
	Altitude asl	ALT
	Relief	REL
Form	Circularity	CIRCLE
	Elongation ratio	ELONG
	Shape ratio	BV_SHAPE
Erosion	Cover	C_LU
	Erodibility	K
	Erosion	USLE_LU
	Erosivity	R
	Slope: length ratio	LS
Process	Wetness index	WET
	Plan curvature	PLAN
	Profile curvature	PROF
	Sediment transport index	TPT
	Stream power index	POW
	Flow accumulation	FACC

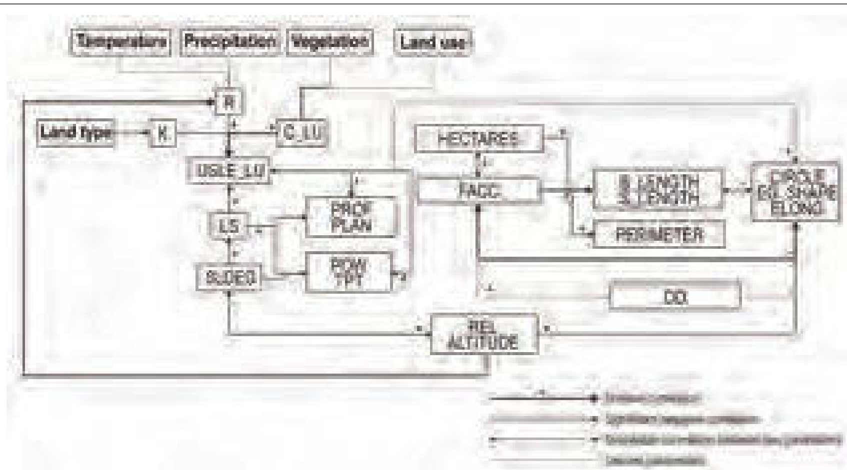

Figure 14.18. A conceptual model of processes and landforms in the Modder River catchment (after Barker, 2002).

The Modder River catchment streams were then modelled Barker (2002 and 2011), with and without the influence of pans. Using standard hydrological functions in a GIS, flow direction and flow accumulation were used to identify flow lines thereby forming modelled streamlines. These lines crossed existing pans (natural depressions in the landscape), which were eliminated by a fill function in the original DEM. Barker (2011) then proposed a form of pan etching to model streams closer to a natural situation. Cluster analysis was performed on second order basins (Barker, 2002) and twelve main distribution clusters in the Modder River catchment were identified based on the parameters identified in Table 14.2. An interesting result, however, was the location of five westerly sub-clusters of one of these clusters' Standard Deviation Ellipses (SDEs, Figure 14.19) and highlights the possible relationship between modelled streams and large pans, while Figure 14.20 illustrates distribution of pan density and relative size.

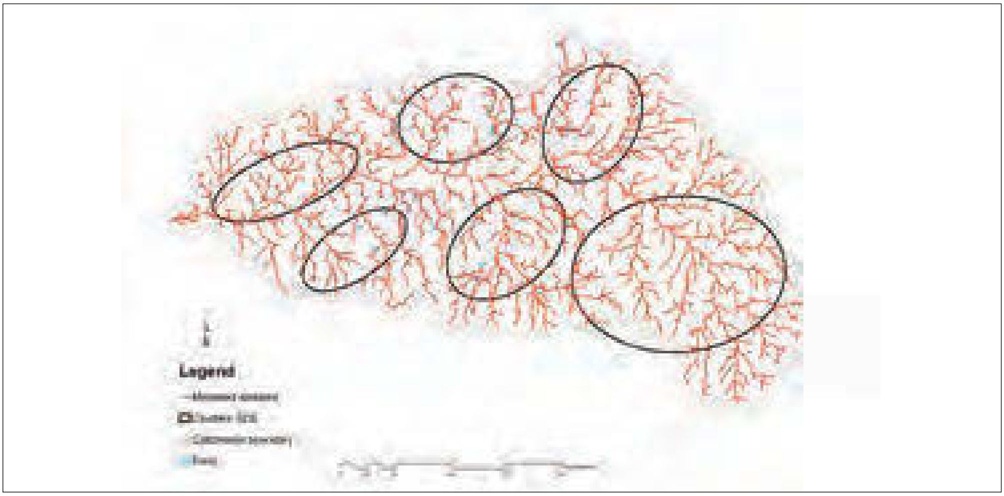

Figure 14.19. Standard deviation ellipses of sub-clusters in the western Modder River catchment shows the relationship between modelled streams and pans (after Barker 2002).

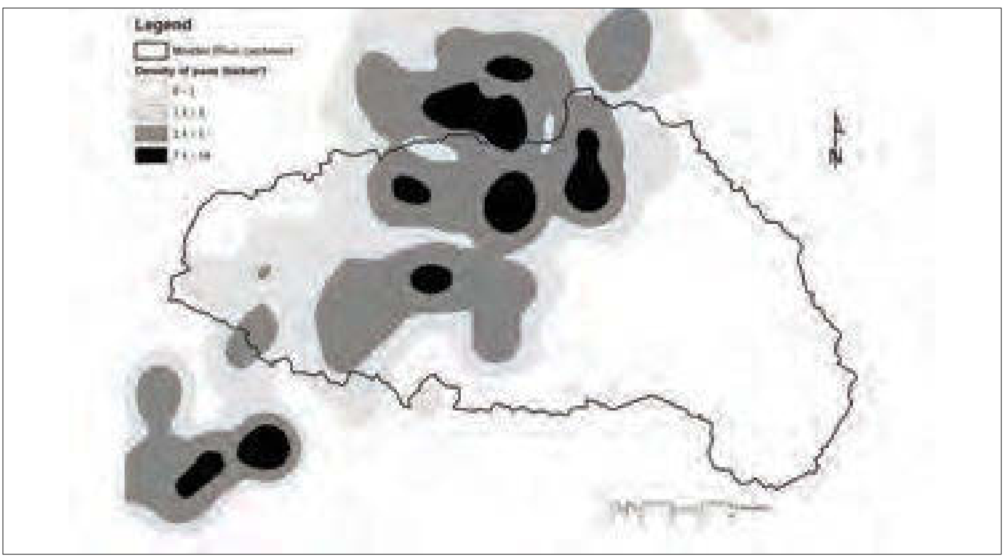

Figure 14.20. The density and size of pans in the Modder River catchment (after Holmes *et al.,* 2008).

6. Conclusion

The perspective of the southern African landscape has changed significantly through time. The study of its landscape has historically made global contributions to the understanding of landscapes in general, but in turn we would argue that more recently southern African geographers and geomorophologists abroad and at home have not fully capitalised on the freely available global datasets including variety of imagery and elevation data as well as national datasets and archives. This chapter has revealed that the potential for understanding the southern African environment, let alone its geomorphology, has been enormously enhanced through the increasing availability and resolution of remote sensing products that make the construction of landscape inventories – for many diverse applications. Access to a networked personal computer, even with relatively modest specifications, opens up detailed examination of southern Africa and the world. This data opens up new avenues for exploration, highlights environmental trends and also provides the broader picture into which many of our existing observations can be placed.

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Southern African Geomorphology: the Past and the Future

Southern African Geomorphology: the Past and the Future

Andrew S Goudie

1. Introduction: the past

In his 1947 Presidential Address to the Geological Society of South Africa, Lester King (1947) took the liberty of reminding his audience that the African continent was full of "queer departures from normal in its topography," but stressed that it is "superlatively rich in certain types of landform." He continued with hyperbole:

> *It yielded to the early masters of geomorphology, W.M. Davis and S. Passarge, information and instances of primary value. From South Africa inselbergs were first described; its interior plateau early excited wonder as an example of planation by the elements; its dongas were among the first such features recorded; and all have remained wonders ever since … it has revealed an astonishing erosional history without known parallel in any other continent.*

As Moon and Dardis (1988:1) wrote: "The southern African landscape constitutes an ideal natural geomorphological laboratory in view if its great diversity and, compared to many of the world's landscapes, great antiquity." On the other hand, as Moon and Partridge (1993:383) pointed out, "southern Africa differs fundamentally from continents of the Northern Hemisphere in the absence from its landscapes of the effects of Pleistocene glaciations and Cenozoic mountain building." They added,

> *... as a result ancient landforms are extensively preserved and the influence of large-scale epeirogenic movements is readily discernible. These characteristics, coupled with a fairly regular east-west climatic gradient across the sub-continent, have favoured a broad perspective, and a substantial literature attests to southern Africa's preeminence in the field of macrogeomorphological research.*

Southern African geomorphologists have (often for geographical reasons) not for the most part made major contributions to the study of the geomorphology of the humid tropics, of current volcanism, or of glacial processes, and they have perhaps been less concerned with the study of geomorphological processes at the small catchment scale than many of their British counterparts (though see Chapter 5 for a discussion of what has been achieved in the study of fluvial geomorphology). Also coastal geomorphology has not developed as much as one might have anticipated, given the high diversity and splendour of the coastline, though as Chapter 10 shows major progress is now taking place. Furthermore, as King (1947) and Beckedahl *et al.*, (2002) lamented, South African geomorphologists have also displayed a limited interest in theory, as well as in applied research and human-environment interactions. On the other hand, building upon the towering contributions of such figures as Alex du Toit, John Wellington, Frank Dixey, Lester King and Tim Partridge, southern African geomorphologists have maintained a sustained interest in large-scale, long-term evolution at a time when short-term reductionist process studies were the vogue in many other countries (see Chapter 1). They have recognised the importance of plate tectonics, mantle plumes, southern Africa's position on a passive margin, isostatic responses to erosional unloading, and long-term drainage system evolution (see, for example, Partridge and Maud, 1987; Moore *et al.,* 2009). They have used such techniques as fission track analysis and cosmogenic nuclides for estimating ages of ancient landforms and rates of long-term landscape evolution. In addition, geomorphologists working in southern Africa have in recent years punched above their weight in terms of developing knowledge of such matters as weathering cycles, calcrete, silcrete and gypcrete formation (see Chapter 8), rock mass strength classifications, and aeolian forms and processes. There has also been a spasm of work, sometimes controversial, on high altitude periglacial phenomena, and also on Marion Island (see, for example, Chapter 4).

Although as been hinted above, southern African geomorphological studies have been characterised by some areas being the subject of sparse attention, while others have been pursued with great vigour and distinction, the discipline has benefitted from great international cross-fertilisation. Large numbers of overseas geomorphologists have come to undertake fieldwork in southern Africa, and especially in the arid lands of Namibia (see Hüser *et al.,* 2001), where for 50 years the Gobabeb Research Station has been an invaluable home and inspiration (see Chapters 6 and 7), while others have taken up long-term appointments, especially in the universities of South Africa itself. Recent studies of erosion in Swaziland have been undertaken by a group with scientists from Russia, Germany and Italy (Sidorchuk *et al.,* 2003), and by various British groups (e.g. Morgan *et al.,* 1997). In Lesotho the study of erosion was carried out expertly by Swedish groups (see, for example, Chakela, 1980), while Dutch scientists have contributed to the study of soil erosion dynamics in Kwazulu-Natal (Sonneveld *et al.,* 2005) and UK scientists to the study of erosion in the Karoo (Boardman *et al.,* 2003) (see also, Chapter 11).

2. The future

2.1 Hazard geomorphology

The rest of this chapter examines where geomorphology may develop in the years ahead and in particular:

- to see how it may develop in a region where there are great socio-economic needs
- to see how it may develop in a world where financial resources for pure research may be limited
- to see how it may develop in relation to some current developing themes of world geomorphology and how it may build upon some of the competitive advantages that southern Africa has for geomorphological research.

Undoubtedly, geomorphology is important. As King (1947:xlviii) wrote:

> *The National value of the subject is plain. Through soil and water conservation it is part of the 'life-blood' of the nation; through erosional disclosing of all the mineral resources it is part of the 'wealth' of the nation; through scenery appreciation and preservation it is part of the 'Soul' of the nation.*

Geomorphological hazards are prevalent in southern Africa and an indication of their range is given in Table 15.1, as are examples of recent studies that have been undertaken. The study of such hazards is likely to be a major future focus for research in the region, and examples of such work are given in Chapter 13.

Table 15.1. Selected geomorphological hazards in southern Africa.

HAZARD	EXAMPLE
Badlands, Great Karoo	Boardman *et al.*, 2003
Channel Avulsion	Grenfell *et al.*, 2009
Channel change	De Villiers and Schmitz, 1992
Channel change and alien plants	Holmes *et al.*, 2005
Channel changes below dams	Ronco *et al.*, 2010
Channel instability, Eastern Cape Province of RSA	Rowntree and Dollar, 1999
Coastal erosion, Eastern Cape Province of RSA	Lubke, 1985
Coastal erosion, Mozambique	Massuanganhe and Arnberg, 2008
Dune migration	Barnes, 2001
Dust storms, Namibia	Eckardt and Kuring, 2005
Erosion associated with mining	Pereira, 2009
Estuarine flooding, Mgeni	Cooper *et al.*, 1990
Flooding and sedimentation, Orange River	Bremner *et al.*, 1990
Gully erosion in Swaziland	Marker and Sidorchuk, 2003; Sidorchuk *et al.*, 2003
Landsides, Durban area	Bell and Maud, 2000; Garland and Olivier, 1993
Mining pit instability	Bye and Bell, 2001
Mining subsidence Piping	Stacey and Bell, 1995 Beckedahl, 1998
Reservoir siltation	Russow and Garland, 2000; Chakela, 1980
River flooding, Sabie River	Heritage *et al.*, 2001; Smithers *et al.*, 2001
River flooding, Swaziland	Goudie and Price Williams, 1984
Road surface erosion	Beckedahl *et al.*, 1999
Rockfall, Cape Peninsula	Volkwein *et al.*, 2005
Salt weathering, Namibia	Goudie *et al.*, 1997
Sinkholes in Gauteng Province of RSA	Buttnick and Van Schalkwyk, 1998; De Bruyn and Bell, 2001
Slope instability in Maputo	Vincente *et al.*, 2006
Soil erosion, KwaZulu-Natal Province of RSA	Sonneveld *et al.*, 2005
Soil erosion, Mpumalanga Province of RSA	Wentzel, 2002
Soil erosion, Eastern Cape Province of RSA	Rowntree *et al.*, 2004; Kakembo and Rowntree, 2003
Soil erosion, Free State Province of RSA Soil erosion, Lesotho	Wiggs and Holmes, 2011 Showers, 1989
Soil erosion, Swaziland	Morgan *et al.*, 1997; Manyatsi and Ntshangase, 2008

HAZARD	EXAMPLE
Soil erosion, Western Cape Province of RSA	Meadows, 2003
Spit instability, Namibia	Wilkinson *et al.*, 1989
Tailing dams failure and liquifaction	Fourie *et al.*, 2001
Weathering of rock art	Hall *et al.*, 2007; Hoerlé *et al.*, 2006

As Rosenfeld (2004:423) wrote:

> *A significant practical contribution of geomorphology is the identification of stable landforms and sites with a low probability of catastrophic or progressive involvement with natural or man-induced processes adverse to human occupance or use. Hazards exist when landscape developing processes conflict with human activity, often with catastrophic results.*

Geomorphic events can kill people and damage property. Although high magnitude, low frequency, catastrophic events, such as hurricanes or tsunamis, gain attention because of the immediacy of large numbers of casualties and great financial losses, there are many more pervasive geomorphological changes which are also of great significance for human welfare. These may have a slower speed of onset, a longer duration, a wider spatial extent and a greater frequency of occurrence. Examples include weathering phenomena and soil erosion. The latter is of huge significance in South Africa and has been massively increased by human activities during the Holocene (Compton *et al.,* 2010). Indeed, there is a great diversity of geomorphological hazards. One major category is mass movements, such as rockfalls, debris flows, and landslides (Diop *et al.,* 2009). There are also various fluvial hazards, such as floods and river channel changes (e.g. avulsion). In coastal environments one has inundation and erosion caused by storm surges, rapid coastal erosion and siltation, sand and dune encroachment, shoreline retreat and sea-level rise. There is also a wide range of subsidence hazards caused by solution of limestone, dolomites and evaporites (e.g. gypsum or halite). In the drier parts of the sub-continent, hazards are posed by wind erosion and deflation of susceptible surfaces, dust storm generation, and by dune migration. More generally, water erosion causes soil loss and gully (donga) or badland formation (see Chapter 11), while weathering can be a threat to a wide range of engineering structures. The incidence of such hazards can be increased or triggered by human activities, and in particular by land use and land cover changes, as in the former homelands of South Africa (Wessels *et al*., 2004).

There is also an increasing concern that the incidence of hazards will be changed in a warmer world. However, another important consideration is the extent to which human societies are placing themselves at an increased risk as population levels increase and new areas are exploited or settled. Potentially hazardous areas, such as floodplains or steep, deeply weathered slopes may become occupied, placing human groups at risk. Large urban populations, such as are developing in southern Africa, may be especially at risk. There is evidence that for these sorts of reasons, damage to property and loss of life caused by geomorphological hazards is increasing globally (Alcántara-Ayala and Goudie, 2010).

The roles of the geomorphologist in hazard research are many. Of great importance are: the mapping of hazard prone areas (Huchschild *et al.,* 2003) and inventorisation (Diop *et al.,* 2009); constructing the history of occurrence of past hazardous events; establishing their frequency and magnitude; predicting the occurrence and location of future events; identifying sources of sediment in catchments (e.g. Compton and Maeke, 2007); monitoring geomorphological change; and using knowledge of the dynamics of geomorphological processes to advise on appropriate mitigation strategies (Alcántara-Ayala and Goudie, 2010). In recent years the capabilities of geomorphologists in these roles has increased and the application of geomorphology to the solution of environmental problems has developed. Techniques such as remote sensing and Geographic Information Systems (GIS), dating by means, for example, of lichenometry, luminescence dating, and dendrochronology, instrumentation of slopes and other phenomena with data-loggers, and computer modelling, have all made major contributions. Particularly since the 1960s geomorphologists have become far more knowledgeable about processes and the mechanisms of geomorphological change. They have also become more aware, by adopting a systems framework, of the inter-relationships between different phenomena and of feedback loops, some positive. Concepts of geomorphological resilience and sensitivity have also been explored (Brunsden, 2001), but so far have been little developed in southern Africa.

2.2 Applied geomorphology

Related to the study of hazards is the more general contribution of applied geomorphology. As Moon and Partridge (1993:392) pointed out, geomorphology has been an applied discipline in southern Africa since the 1960s, but much applied/environmental geomorphology has been carried out by engineering geologists, hydrologists, soil scientists and geomorphological engineers rather than by geomorphologists *per se*. There is great scope for developing this area of research, not least because of the benefits it can generate for rapidly evolving societies.

A very basic, but highly important role (Table 15.2) is to map geomorphological phenomena as a basis for terrain evaluation (see Chapter 14 for the new techniques that are available). Landforms, especially depositional ones (including duricrusts) may be important sources of useful materials for construction, while maps of slope categories may help in the planning of land use and maps of hazardous ground may facilitate the optimal location of engineering structures (Diop *et al*., 2009; Singh *et al*., 2008).

Table 15.2. The roles of the applied geomorphologist.

1.	Mapping of landforms, resources and hazards.
2.	Use of maps of landforms as surrogates for other phenomena (e.g. soils).
3.	Establishment of rates of geomorphological change by direct monitoring, use of sequential maps, archives, etc.
4.	Establishing causes of change.
5.	Assessment of management options.
6.	Post-construction assessment of engineering schemes.
7.	Post-event evaluations (e.g. palaeodischarges).
8.	Prediction of future events and changes.
9.	Conservation of sites and explanation to visitors.

A second role of the applied geomorphologist is to use landforms as the basis for mapping other aspects of the environment, the distribution of which is related to their position on different landforms. This is important because landforms are relatively easily recognised on air photographs and other types of remote sensing imagery. An important example of the use of landforms as surrogates for other phenomena is the use of landform mapping to provide the basis of a soil map. Cartographic skills have been revolutionised in recent years through the use of new technologies including, Remote Sensing by satellites, Global Positioning Systems, Geographic Information Systems, Digital Elevation Models, and LIDAR. Maps are especially important for land use planning and zoning. An important recent study of the application of mapping techniques to hazard prediction is that of Le Roux *et al*. (2008). Some major advances in mapping landforms can now be achieved at no financial cost though the use of such increasingly high resolution sources as Google Earth (see, for example, Goudie, 2010).

The third role of the applied geomorphologist is to recognise and measure the speed at which geomorphological change is taking place. Such changes may be hazardous to humans. By using sequential maps and remote sensing images, archival information or by monitoring processes with appropriate instrumentation, areas at potential risk can be identified, and predictions can be made as to the amount and direction of change.

Fouthly, having decided on the speed, location and causes of change, appropriate management solutions are required. Although the solution to a particular geomorphological problem may involve the building of an engineering structure (e.g. a sand fence, a sea wall, a check dam, or a shelterbelt), these structures may themselves create problems and their relative effectiveness needs to be assessed (Theron and Schoonees, 2007). The applied geomorphologist may make certain recommendations as to the likely consequences of building, for example, groynes to reduce coastal erosion. Examples of engineering solutions having unforeseen environmental consequences, sometimes to the extent that the original problem is heightened and intensified rather than reduced, are all too common, especially in coastal situations (La Cock and Burkinshaw, 1996). Management issues involve a consideration of ecological issues, as when one decides on the most appropriate form for a river channelisation scheme, and are likely to become increasingly important, as decisions have to be made about how to manage the landscape in the face of global climate change. More and more alternatives are being sought to ecologically injurious "hard engineering" solutions. Geomorphologists need to be involved in the restoration or redesign of disrupted and degraded landscapes.

Fifthly, and related to environmental management and the use of engineering solutions, is the field of assessment of the success of particular schemes (e.g. Schoonees *et al*., 2006). An audit of performance is required as the basis for formulating best practice.

A sixth role of the applied geomorphologist is to undertake *after-the-event* surveys. These put on record the magnitude and consequences of extreme events as a basis for improving engineering designs and land zoning policies (Goudie and Price Williams, 1984). For instance, establishing the Holocene flood histories of rivers by surveying and dating slack water deposits laid down by earlier floods give an important tool for predicting possible future flood peaks, especially in ungauged catchments. This is an area where there has already been a substantial amount of work in South Africa (Zawada, 1994; 1997; Smith and Zawada, 1990).

Related to this, another role of the applied geomorphologist is to look forward and to predict. How long will it take for a railway to be blocked by a wandering barchan, when is this slope likely to fail, how quickly will this reservoir be rendered useless through sedimentation, will the surface of a delta be built up by fluvial sediment inputs more quickly than sea-level rises? These are examples of where geomorphologists can help to answer questions about the future. Their answers can be based on studies of the past rate of operation of geomorphological processes or by developing their modelling capability. Even more importantly, geomorphologists can contribute to understanding the impacts of potential future global change associated with the enhanced greenhouse effect. There are other facets of global change besides the changes associated with global warming. These include such geomorphologically vital issues as accelerated soil erosion caused by vegetation clearance and other mechanisms, not least in communal areas (Hoffman and Todd, 2000). The role of urbanisation, agricultural intensification, deforestation, lumbering and mining in causing such modification of landscape and the rate of operation of processes is an area of significant public concern.

Finally, geomorphologists are involved with designating and advising on the conservation of geomorphological sites that are of especial value as World Heritage Sites, Geoparks and Geosites (see

Goudie and Seely, 2011), for a discussion of desert sites, and Grab *et al.,* (2011) for a discussion of one particular site, Golden Gate Highlands National Park, South Africa).

2.3 Geomorphological implications of global change

Global climates will probably change substantially during this century and will affect the operation of geomorphological hazards and the nature of landscapes in southern Africa as a result of changes in temperatures, precipitation amounts, rainfall intensities, and soil moisture conditions (Schulze, 1997; Meadows, 2006; see also Chapter 12). It should be remembered that climate change will work in tandem with various anthropogenic activities including changes in land cover and land use (Slaymaker *et al.,* 2009) which could be of equal or greater importance (Table 15.3). Some environments will change more than others – "geomorphological hotspots" (Goudie, 2006) – especially when crucial thresholds are crossed. For instance, some landforms and land-forming processes change across crucial climatic thresholds. Aeolian activity in drylands like the Kalahari is dependent on wind energy, sand supply and the nature and extent of vegetation cover. If the last falls below a certain level due to a reduction in moisture availability, wind action is sharply intensified. Another example of threshold dependence is the coral reef. These features are, highly sensitive to any changes in cyclone activity, but also to coral bleaching caused by elevated sea-surface temperatures (as has happened during recent El Niño years in the Indian Ocean).

Table 15.3. Climate change and other human pressures on selected environments from developing countries.

ENVIRONMENT	CLIMATE CHANGE PRESSURES	OTHER PRESSURES
Mangrove swamp	Sea-level rise and spread of salinity up estuaries	Deforestation, pollution, fish farming, infilling
Coral reef	Sea-level rise, excess temperatures leading to bleaching, greater hurricane attack	Pollution (including siltation), trawling, dynamiting, introduction of invasive predators, mining, dredging, etc.
Rainforest	Rainfall change leading to drought, change in albedo, more fire, etc.	Deforestation, burning
Deltas	Sea-level rise, landward spread of salinity, increased erosion, hurricane attack	Sediment starvation and less freshwater flow (because of dams), subsidence caused by over-pumping of water, hydrocarbons, etc.
Desert margin	Reduced soil moisture, dune reactivation, reduction in river flow, contraction of drainage net	Land disturbance, over- cultivation, deforestation, inter-basin water transfers
Sandy beaches	Sea-level rise leading to increased erosion, more storm surges	Sediment starvation, vegetation removal, updrift engineering structures, mining

Secondly, there are many examples of landform change being promoted by a combination of climate changes and other human pressures – the compound effect. Indeed, desertification is often most intense when climatic (drought) and human pressures (e.g. over-grazing, deforestation, or over-cultivation) coincide. In coastal regions, beaches, marshes and deltas starved of sediment by the damming of rivers and the construction of coastal defences will be especially prone to erosion and inundation as sea levels rise (Blum and Roberts, 2009).

Thirdly, some landforms are robust, while others are less so. For example, dongas (erosional scars) in colluvial aprons in southern Africa are gully systems that have a history of cut and fill in response to climate and land cover changes. This is because they are developed in materials that have high exchangeable sodium percentages and thus are prone to severe erosion if the vegetation cover is changed (Rienks *et al.*, 2000). Likewise, once their protective vegetation cover is removed, sand dunes in semi-arid areas are easily reactivated.

Fourthly, the severity of climate change, and thus its likely impact, will vary spatially. Reductions in soil moisture and stream flows may be especially great in some areas that are currently relatively dry, so that African stream networks will shrink (De Wit and Stanciewicz, 2006). Features such as the dambos of Zimbabwe, Zambia and Malawi are among those fluvial systems that may be considerably altered.

Sea-level rise of around 50 cm is likely to occur by 2100 (Miller and Douglas, 2004). However, should Greenland's ice melt at a faster rate than is currently predicted, the amount of rise will be greater, plausibly leading to 0.8 m of sea-level rise by 2100 (Pfeffer *et al.,* 2008). Coral reefs, such as occur in Maputaland and Mozambique, because of their low- lying nature, may be subject to over-washing, inundation and salt water incursion as sea-level rises (Woodroffe, 2008), particularly if, because of pollution or siltation, they are unable to grow upwards at an adequate rate to keep up with sea-level rise. They may also suffer from increasing ocean acidification (Doney, 2006). This will harm those organisms like corals that depend on the presence of carbonate ions to build their hard parts (Orr *et al.,* 2005). Another threat to coral reefs is coral bleaching caused by excessive seawater temperatures (Baker *et al.*, 2008). The frequency of thermal stress events is likely to increase, leading to declines in coral cover, shifts in the composition of corals and other reef-dwelling organisms, and stress on the human populations who rely upon these ecosystems for food, income and shoreline protection (Donner, 2009).

Mangrove swamps are developed along the eastern coastline of southern Africa (McNee, 1963). They are ecologically important and also offer some protection against coastal erosion. Many mangroves communities will be affected by sea-level rise. Species zonation may change, with trees to seaward being inhibited by extended submergence times, while those on the landward margins might be able to extend, provided that suitable habitat was available and assuming that their migration was not impeded by sea walls and other engineering structures. Mangroves are relatively slow growing and this may make them susceptible to fast rates of sea-level change. On the other hand there may be some expansion in their range as sea surface temperatures rise (Hogarth, 1999). They are also subject to other anthropogenic pressures (Rajkaran *et al.,* 2009).

More generally, sea-level rise may accelerate the rate of erosion of soft coastlines. These are already suffering from erosion because of sediment starvation caused by a reduction in the amounts of material being delivered to beaches by rivers that have been dammed. However, following the Bruun Rule, and bearing in mind that storms and surges may become more intense, it is likely that this ongoing erosional tendency will be increased. Barrier islands and spits which, for example, occur widely along the Mozambican coastline, are dynamic and often densely settled landforms that will tend, as sea-level rises, to be exposed to higher storm surges and greater flooding. Moreover it is not just sea-level rise that will be caused by global warming. There will also be changes in storm characteristics, wave climate, river inputs of water and sediment, and the like (Day *et al.*, 2008).

Future changes in rainfall may have a range of geomorphological and hydrological consequences (Schulze, 1997). Rainfall intensity is a major factor in controlling phenomena such as flooding, soil erosion, and mass movements. Under increased greenhouse gas concentrations some General Circulation Models (GCMs) exhibit enhanced global precipitation intensity and shortened return periods of extreme events (New *et al.*, 2001). Some models suggest that intense rainfall events may become more prevalent in southern Africa (Shongwe *et al.*, 2009), and some scientists believe that the recent increase in rainfall intensity in southern Africa could be a response to recent warming and be a harbinger of things to come (Mason *et al.*, 1999). A big issue of particular concern with regard to precipitation trends relates to the El Niño – Southern Oscillation (ENSO), which is important with respect, *inter alia*, to the occurrence of floods in southern Africa (Reason and Jagadheesna, 2005; Kane, 2009). Some models suggest a weakening of ENSO activity under a global warming scenario, while others do not.

It is possible, though by no means certain, that as the oceans warm up, so the geographical spread, intensity and frequency of hurricanes will increase (Emanuel, 1987; Knutson and Tuleya, 1999). The relationship between sea-surface temperature increases, and increasing global hurricane activity has been confirmed by Hoyos *et al.*, (2006) and Saunders and Lea (2008). Hurricanes are a major hazard for human communities and also have a whole series of hazardous consequences, including accentuated river flooding, greater wave heights (Komar and Allen, 2008), coastal surges, the triggering of landslides and accelerated land erosion and lagoonal siltation.

In some parts of southern Africa precipitation amounts many fall on an annual or seasonal basis, providing longer and more severe droughts in the drier parts of the region (Shongwe *et al.*, 2009). Stream flows in such areas may be reduced and a combination of lower precipitation and higher moisture losses through evaporation may cause a runoff reduction of 40% in the Zambezi, 30% in the Limpopo and 5% in the Orange Basins (Arnell, 1999). Various studies have suggested that the nature and extent of the Okavango Swamps could be transformed (e.g. Andersson *et al.*, 2006; Milzow *et al.,* 2010). Precipitation and runoff changes may also have an influence on land degradation (Meadows and Hoffman, 2003).

Changes in climate could affect wind erosion either through their impact on erosivity or through their effect on erodibility. The former is controlled by a range of wind variables including velocity, frequency, duration, magnitude, shear and turbulence. Unfortunately, GCMs as yet give little indication of how these characteristics might be modified in a warmer world, so that prediction of future changes in wind erosivity is problematic. Vegetation cover and surface type, both of which can be influenced markedly by climate, largely control erodibility. In general, vegetation cover, which protects the ground surface and modifies the wind regime, decreases as conditions become more arid. Likewise climate affects surface materials and their erodibility by controlling their moisture content, the nature and amount of clay mineral content (cohesiveness) and organic levels. However, modelling the response of wind erosion to climatic variables on farmland is vastly complex, not least because of the variability of soil characteristics, topographic variations, the state of plant growth and residue decomposition, and the existence of windbreaks. To this needs to be added the temporal variability of aeolian processes and moisture conditions and the effects of different land management practices (Leys, 1999), which may themselves change with climate change.

Sand dunes, because of the crucial relationships between vegetation cover and sand movement, are highly susceptible to changes of climate. Some areas, such as the South West Kalahari (Stokes *et al.,* 1997) may have been especially prone to changes in precipitation and/or wind velocity because of their location in climatic zones that are close to a climatic threshold between dune stability and activity.

Detailed scenarios for dune remobilisation by global warming have been developed for the mega-Kalahari in southern Africa (Thomas *et al.,* 2005). Much of this vast region is currently vegetated and stable, but GCMs suggest that by 2099 all dunefields, from South Africa and Botswana in the south to Zambia and Angola in the north will be reactivated. This could disrupt pastoral and agricultural systems and might lead to loss of agricultural land, the overwhelming of buildings, roads, canals, runways and the like, abrasion of structures and equipment, damage to crops, and the impoverishment of soil structure. However, the methods used to estimate future dunefield mobility are still problematic and much more research is needed before we can have confidence in them (Knight *et al.,* 2004). Ashkenazy *et al.,* (2011), for example, in contrast to Thomas *et al.,* (2005) predict rather limited change in the Kalahari Dunefields in future decades.

Climate change may also have various effects on vegetation communities, including the contraction of the *fynbos* (Midgley *et al.,* 2003) and various impacts on alien invasive species (Parker-Allie *et al.,* 2009). These in turn could have geomorphological impact by affecting rates of erosion and stream channel stability (Holmes *et al.,* 2005).

In coming decades geomorphologists will need to identify potential “geomorphological hot spots,” to monitor rates of geomorphological change, to model future landscapes responses to climate change, to consider appropriate mitigation strategies and to evaluate any consequences that engineering solutions may create.

2.4 Landscape education and conservation

Southern Africa has many spectacular landscapes and landforms (Table 15.4), and such features need to be explained to the public and also to be conserved as World Heritage Sites, geosites, geoparks, or places for the burgeoning activity of geotourism. On a global basis an illustration of the explanation of spectacular landforms is provided in Migoń (2010), while Norman and Whitfeld (2006) provide a guide to some of the landforms of South Africa, and Hüser *et al.*, (2001) to those of Namibia.

Table 15.4. Some notable geomorphological sites in southern Africa (see Figure 15.1).

SITE	PHENOMENA
Botswana	
1. Okavango Swamps	Mega-fan
2. Makgadikgadi Pan	Pan
Lesotho	
1. Dongas	Erosion scars
Namibia	
1. Brandberg	Unroofed intrusion
2. Brukarros	Volcanic crater
3. Bulls Party, Ameib	Granite weathering, including pits
4. Etosha Pan	Pan
5. Fish River Canyon	Deep drainage incision
6. Horuasib Clay Castles	Aeolian blocked drainage
7. Koes Pan and Lunette	Deflation
8. Messum Crater	Eroded annular intrusion
9. Naukluft tufas	Semi-arid tufas
10. Sossusvlei	Star dunes, dammed drainage
11. Spitzkoppe	Inselbergs
12. Twyfelfontein	Weathering rinds
13. Waterberg Plateau	Sandstone weathering
14. Weissrand Dayas	Semi-arid limestone solution
South Africa	
1. Alexandria	Coastal dunes
2. Aughrabies Falls	Waterfall
3. Blyde River Canyon	Potholes
4. Boesmangat	Sinkhole
5. Cederberg	Sandstone weathering
6. Chapman's Peak Drive	Deep weathering of granite
7. Drakensberg	Periglacial phenomena
8. Golden Gate	Sandstone weathering
9. Grahamstown	Silcrete + deep weathering
10. Treasure Beach, Durban	Aeolianites and costal algal rims etc.
11. Tswaing Crater	Impact structure
12. Valley of 1000 Hills	Granite domes
13. Vredefort Dome	Impact structure
Swaziland	
1. Mashlila	Colluvial dongas
2. Sibebe	Granite dome

Figure 15.1. Some notable geomorphological sites in southern Africa. Cross-reference to Table 15.4 for names.

2.5 Some other future themes

2.5.1 Submarine geomorphology

Geomorphology does not stop at the edge of the ocean. Submarine geomorphology is an important field, not least for assessing the development of off-shore hydrocarbon reserves, as has occurred, for example on the Nile Delta, the Angolan Fan and the Congo Fan (Savoye *et al.,* 2008). Landslides, turbidity currents, salt dissolution features and the like pose problems for pipelines and rigs. The continental shelf off southern Africa is well known for the size of its landslides (see, for example, Dingle and Robson, 1985 on the coast off East London; and Green and Uken, 2008 on the KwaZulu-Natal coast), its submarine valley systems (Green, 2009) and canyons (Green *et al.*, 2007).

2.5.2 Biogeomorphology, ecosystems and landscape co-evolution

One area where southern Africa has a very strong competitive advantage is with respect to the high level of sustained research that has been undertaken on the wildlife and ecology of the area. This means that there is the potential for collaborative world class research on the role of organisms as geomorphological agents, but also on the importance of landforms and land forming processes for understanding and managing specific terrains that have ecological significance (Dollar *et al.,* 2007). With respect to the latter, the work undertaken in the USA on "geomorphological integrity" (Graf, 2001) could act as a template, while with respect to the former there is great scope to pursue studies of the role of organisms in affecting rates of process operation, as has been done for lichens (e.g. Wessels, 1988; Cooks and Otto, 1990) and termites at one end of the scale, rats at an intermediate scale, and elephants and large herbivores at the other (e.g. Boelhouwers and Scheepers, 2004; Harrison *et al.,* 2007).

2.5.3 Rock control

Southern African geomorphologists have for long had an interest in the effects of rock on relief (see Chapter 3 for a discussion of the study of granitic terrains). This is illustrated by the work that has, for example, been undertaken on karst (e.g. Marker, 1970), the work that has been undertaken by Moon and his co-workers on the relationship between rock-mass strength and slope form (Moon and Selby, 1983), and the work that has been undertaken to test petrological controls on inselberg distribution and morphology (Pye *et al.,* 1984). New techniques are always being developed for characterising rock properties (e.g. Viles *et al.,* 2011) and these need further development. In addition, because of the mining and construction industries there has been a great interest in the geotechnical properties of rocks and earth surface materials, so that there is scope for collaboration between geotechnical engineers and geomorphologists. Rock control has for too long been a Cinderella of world geomorphology, with only Japanese geomorphologists seeing it as a major priority, so there is great scope for making fundamental contributions.

2.5.4 Planetary analogues

A major area of international research is into the geomorphology of other bodies in our solar system, such as Mars, Titan and Venus. There are numerous features in southern Africa that have the potential to be used as analogues for Mars, including the dunes of the Namib (Bourke and Goudie, 2009), various sandstone weathering forms, and groundwater sapping features in the Kalahari (Nash *et al.,* 1994), while Etosha Pan might be used as an analogue for lakes on Titan (Cornet *et al.,* 2012). In short, geomorphologists working in southern Africa could pursue similar investigations to those taken in, for example, Egypt and Australia.

2.5.5 Past environments

Geomorphologists have carried out the study of past environments over large tracts of southern Africa (see Chapter 12). Given the general interest that the scientific community as a whole has in this issue, it is likely to continue as a flourishing area of research. There is particular scope for collaboration with archaeologists to explore the links between environmental change and human evolution and cultural change. Southern Africa is richly endowed with many geomorphological phenomena with palaeoenvironmental significance including caves, tufa deposits, dunes, and pluvial lakes. It has also proved to be fertile ground for the application of new dating techniques, such as optically stimulated luminescence (e.g. Burrough *et al.,* 2009).

2.5.6 Long-term evolution

Finally, and getting back to one of the most distinctive and distinguished facets of southern African geomorphology, there is likely to be a further expansion of studies of long-term landform evolution based on such techniques as cosmogenic nuclide analysis (Cockburn *et al.,* 1999; Bierman and Caffee, 2001) and apatite fission analysis (Raab *et al.,* 2005). These can give us data on the ages of ancient landforms and on the rates at which they have been denuded over millions of years.

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Southern African Geomorphology provides a unique account of the varied physical landscapes of the subcontinent and the landforms which comprise them. It describes these landscapes systematically, and explains their evolution and development against a backdrop of recent trends and new developments within geomorphology. ***It aims to fill a niche with regard to understanding Earth surface processes and*** their products in a southern African context. It is an academic text, yet one which will satisfy the lay reader who wishes to learn more about the southern African landscape, and the processes responsible for it.

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Peter Holmes is Professor in the Department of Geography at the University of the Free State, Bloemfontein, South Africa.

Michael Meadows is Professor and Head of the Department of Environmental and Geographical Science, University of Cape Town, South Africa.

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