The Erosional and Cainozoic Depositional History of the Lower Orange River, southwestern Africa

Volume 1

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Abstract

The Orange River and its main tributary, the Vaal form the principal westward-directed drainage of southern Africa, and have done so for at least the last 100 Ma. Study of this long-lived drainage, in particular its distal reaches as in this case, provides insights into the evolution of the African South Atlantic passive margin from both onshore and offshore perspectives, as well as the development of gem-quality fluvial diamond placers along the Lower Orange River. Although many larger rivers drain passive margins into the Atlantic Ocean, the Orange River is unique in being a continental-scale river locked into a steep bedrock channel flowing through a high-relief landscape only 100 km from its mouth. This high relief, forming part of the "Great Escarpment", represents the western margin of the interior plateau of southern Africa, that mostly lies above 1 km elevation due to major epeirogenic uplift initially in latest Cretaceous to earliest Tertiary times. This sub-continental uplift superimposed a Cretaceous meandering (fine bedload, no diamonds) large river into the western plateau margin resulting in a steeper, bedrock-incised, gravel bearing (including diamond), smaller drainage that has persisted throughout the Cainozoic.

A series of terraces flanking the Lower Orange River in the study area were deposited after ca. 90% of the incision had occurred, thus only the late stage incision/depositional history of this margin is able to be addressed here. Two principal suites of river terraces are distinguished by their palaeo-courses, bedrock strath levels, overall geometry and clast assemblages: an older, higher lying Proto suite and a younger Meso suite. The Proto suite represents a long, post-Eocene, through the Oligocene into the Early Miocene, phase of incision, followed by a prolonged period of aggradation where up to 90 m of fluvial, diamondiferous deposits accumulated during the Early-Middle Miocene. The Meso suite of deposits represents shorter phases of incision and aggradation in the Plio-Pleistocene. The Proto and Meso deposits were built in response to both base level rise and increased supply of material from tributaries draining the Great Escarpment locally, with clast assemblage and downstream fining data indicating the latter to be the more important variable.

In the river terrace deposits, clast assemblages change temporally from older siliceous, agate-dominated (exotic) gravels derived from deep in the interior of southern Africa in Eocene times to material sourced locally out of the Great Escarpment in Miocene and younger times. The changes in proportions of exotic clast types reflect the stripping out of the post-Gondwana and Karoo cover in the southern African hinterland, as well as the demise of the Vaal River clastic input due to progressive aridification of the central and western parts of southern Africa through the Cainozoic. Analysis of one of the rarest clast types, viz., diamond, in the Lower Orange River deposits, is consistent with a waning - but coarsening - supply of diamonds through the drainage with the peak in diamond supply occurring in the Oligocene to Early Miocene terrace deposits. Diamond input is correlated with the Vaal River suite of clasts, pointing to that catchment as a more prominent source than the Upper Orange catchment, which has largely supplied the water for this system, particularly during the Late Cainozoic.

Composition and downstream fining data from the modern and ancient deposits indicate that clast abrasion is a more prominent process than selective sorting in the degradational deposits, whereas selective sorting is more prominent in the aggradational settings. Comparison of longitudinal bedrock profiles, rates of downstream fining, clast type proportions and roundness indicate that the modern deposits were laid down by a river considerably less competent than that in Proto and Meso times. Significantly, volumes of material entering the Lower Orange River via the tributaries draining the Great Escarpment is also less in the current drainage. Moreover, in Proto times, the Lower Orange was a more confined river than its younger counterparts and the coastal plain at that time exhibited more relief than today.

River incision into bedrock is a topic of great interest to fluvial geomorphologists, although most data are derived from active tectonic settings. The incision of a large river into a plateau surface is relatively rare, the best known example being the Colorado River in the young (6 Ma) Grand Canyon. The Orange River in the study area represents a long-lived example of this setting, with the present day dissected topography having evolved from more confined canyon-like walls following the early incision of the Orange River in the Early Tertiary. Although a long-lived incision, the modern channel is not graded in the study area, and is actively incising. The world-wide dataset of incision rates in modern rivers indicates that the Orange River could have completed its entire incision within less than a million years. The continued downcutting of this river so long after the initial incision event is indicative of the roles of intermittent, ongoing epeirogenesis and/or eustatic influences (both of which cannot be proven at this stage), tributary input from the plateau rim (Great Escarpment) or merely the long lag time involved in landscape adjustment following incision into a plateau surface.

Quotation

I implore you, Good Jesus,

that in your mercy you

have given me to drink in

with delight the words of your knowledge,

so of your loving kindness

you will also grant me

one day to come to you,

the fountain of all wisdom,

and to stand forever before your face.

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List of Abbreviations

Where possible, abbreviations are used throughout the text, however reference can be made to this section.

AFTT	Apatite fission track thermometry
BIF	banded iron formation
Carats	cts
cpht	carats per hundred tons
cm	centimetre
dm	Decimetre
DGPS	Differential global positioning system
DTM	Digital terrain model
Ga	Billion years
km	kilometre
Ma	Million years
Mamsl	Metres above mean sea level
Mbmsl	Metres below mean sea level
MCS	Maximum clast size
m	Metre
NMP	Namaqua metamorphic province
SEM	Scanning electron microscope
SG	Specific gravity
Stone	stn
USA	United States of America

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Declaration

The material presented in this thesis is the result of my own independent research, between January 2001 and January 2005. All previously published or unpublished work by other researchers quoted in this thesis is given full acknowledgement in the text.

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1 Introduction

1.1 Overview

The Orange River and its major tributary the Vaal River, together form the principal fluvial conduit that drains the majority of the central and western regions of southern Africa (Fig.1.1a). Although a less competent river today than the Orange, the geomorphic setting of the Vaal River indicates that this drainage is the main trunk stream of this large basin.

The study area is the lowermost ca. 300 km of the Orange River drainage system, between Noordoewer and the Atlantic Ocean (Fig.1.1b). In this area, the Orange River forms the border between Republics of Namibia and South Africa and breaches a portion of the 'Great Escarpment', which separates the relatively high elevation interior plateau and the low elevation coastal plain. This dissected area where the river traverses the escarpment is referred to as the Richtersveld on the South African side (Fig.1.1b). In the reach between Noordoewer and the river mouth, a series of terraces are well preserved, whereas upstream of Noordoewer, terraces flanking the modern river are scarce. On the Namibian (right) bank, Namdeb Diamond Corporation Pty (Ltd) mines the palaeo-river terrace deposits for diamonds and on the south (left) bank, diamond mining is undertaken by the Transhex Group.

1.2 Rationale

The Orange/Vaal River system is a large, long-lived continental-scale river draining into the South Atlantic passive margin and presents an ideal opportunity to analyse the post-breakup history of that margin, from both the offshore and onshore records. An understanding of passive margins is important in that they constitute about 50% of the overall length of present day continental margins (Gallagher & Brown 1997). The evolution of passive margins have traditionally been made from the offshore sedimentary basins using seismic and sequence stratigraphic methods to analyse the structural features and depositional record (*inter alia,* Gerrard & Smith 1984; Light *et al.* 1993; Brown *et al.* 1995; Pazzaglia & Brandon 1996; Clemson *et al.* 1997; Bagguley & Prosser 1999; Menzies *et al.* 2002) and onshore, using traditional geomorphic methods such as the recognition and correlation of erosion surfaces and analysis of weathering deposits and duricrusts (*inter alia,* King 1962; Ambrose 1964; Partridge & Maud 1987; Widdowson 1997; Hill 1999; Twidale 2003). These onshore approaches have been augmented by modern quantitative thermochronological methods such as apatite fission track thermochronology (AFTT) (*inter alia,* Gleadow 1981; Green *et al.* 1989; Brown *et al.* 1990; Brown 1991; Gallagher & Brown 1997; Brown & Gallagher 1999; Gallagher & Brown 1999a; Brown *et al.* 2002) and cosmogenic isotope analysis (*inter alia,* Fleming *et al.* 1999; Cockburn *et al.* 2000; Bierman & Caffee 2001; Van der Wateren & Dunai 2001) to analyse the onshore denudational record. In addition, a variety of surface process numerical models have been developed that attempt to link lithospheric processes to surface processes in the evolution of passive margins (*inter alia,* Gilchrist & Summerfield 1991; Gilchrist & Summerfield 1994; Beaumont *et al.* 2000; van der Beek *et al.* 2002). Relatively few studies attempt to combine the onshore and offshore data (Martin 1973; Ollier & Pain 1997).

All of the above-mentioned approaches to passive margin evolution suffer shortcomings. The offshore studies furnish useful information about volumes, rates and lithological nature of sediment supply over time (Rust & Summerfield 1990; Pazzaglia & Brandon 1996), however, they do not provide direct evidence of the sediment source regions (Gallagher & Brown 1999b). In the onshore studies, there are notable disparities between the traditional geomorphic and quantitative (AFTT, cosmogenic, modelling) approaches. A problem with the former methods is their lack of dating control and therefore strong reliance on the correlation of surfaces based on morphological or sedimentological differences (van der Beek *et al.* 2002), whereas AFTT studies tend to ignore lithological variations and are reliant on the assumption that the palaeogeothermal gradient is similar to the present, which itself is variable in different areas (Hill 1999; Gunnell 2000). Cosmogenic isotope studies are limited to relatively short timescales of c.5-6 Ma (Gallagher & Brown 1999b), and numerical modelling studies are difficult to evaluate as they are often based on numbers obtained from AFTT studies (van der Beek *et al.* 2002) and are not tied back to field-based evidence.

The key feature in onshore studies is denudation: How much material has come off the land surface and when? Rivers are the primary control on the rate at which a landsurface is lowered (Snyder *et al.* 2000), and it follows that large continental-scale rivers on passive margins are important in that they are potentially the most effective agents in denuding the landscape, create large-scale geomorphological features and are also the principal contributors to the offshore sedimentary record. Although there are many large rivers associated with Atlantic passive margins (eg. Amazon, Plate-Parana, Mississippi/Missouri/Red, St Lawrence, Congo and Niger Rivers), the Orange River system is unique in that it remains locked into an incised, gravel bearing-bedrock channel surrounded by a high relief landscape within 100 km of its mouth, unlike the rivers mentioned above, which debouch onto wide, low elevation coastal plains. This is due to the comparatively high elevation of southern Africa compared to the other passive margins (Nyblade & Robinson 1994; Partridge 1998), with the majority of its ca.1 million km^2 drainage basin being elevated at more than 1 km. In addition the Orange River experiences a nett decrease in discharge downstream, as the river flows across the semi- to hyper-arid western areas of the sub continent (Everson 1999), as well as the fact that in its lowermost reaches, it cuts through the high-relief Great Escarpment area.

This study addresses both the erosional and geomorphological features of the landscape, as well as the depositional history present in the Lower Orange River terraces, thus providing field-based constraints on, and interpretations of these mutually exclusive offshore and onshore approaches.

River incision is a topic of great interest to fluvial geomorphologists and the literature abounds with numerous examples from all scales of rivers (*inter alia,* Schumm 1999; Wohl 1999; Whipple 2004). Most examples are from relatively small rivers draining tectonically active areas such as in California (Merritts & Vincent 1989), or large rivers crossing mountain orogens such as the Indus River through the Himalayas (Harbor *et al.* 1994; Burbank *et al.* 1996). The incision of an uplifted, plateau landscape, however, is a small subset of these studies, the best known example being the Grand Canyon in the Western USA (*inter alia,* Hamblin 1994; Lucchitta *et al.* 2000; Pederson *et al.* 2002) which is a relatively young incision and well inland. The Orange River presents an opportunity to study a long-lived incision into a plateau landscape near its terminus (the Atlantic Ocean) in an arid setting. This effect of a large incising river on the landscape, as well as the late stage depositional history of this fluvial system with the complications of river terracing, terrace types, and correlations of terrace levels, are also addressed in this study. Intensive diamond mining operations on the terraces have exposed large tracts of palaeoriver bed, unavailable to other studies of this nature.

Much work has been done on the downstream fining of gravel clasts and the relative roles of sorting and abrasion (Sneed & Folk 1957; Gasparini *et al.* 1999; Hoey & Bluck 1999; Rice 1999). However, most studies have been conducted on rivers in aggradational regimes (Hoey & Ferguson 1997; Gomez *et al.* 2001; Rice & Church 2001) whilst relatively few have addressed vigorous bedrock rivers (Kodama 1994a) and even fewer studies have concerned themselves with downstream fining and composition of gravel through time (Toyoshima 1987). The Lower Orange River Cainozoic deposits allow the relative roles of sorting and abrasion in a degradational environment, through time, to be assessed. In addition, few studies have used detailed clast assemblages per size fraction (Bluck 1969a;1980; Haughton & Bluck 1988) to infer regional geological and landscape evolution, both of which are attempted here. Due to the paucity of fossils in the Orange River terrace deposits, clast assemblages are used as an aid to correlate terraces. The temporal and spacial variation in concentration and stone size of one of the rarest clast types, diamond, found in the Lower Orange River terraces is also addressed, providing further insight into the Tertiary evolution of this passive margin.

1.3 Previous work on the Orange/Vaal system

The main features of the Orange River were described by Wellington (1955; 1958)who divided the river into the Upper Orange above its confluence with the Vaal, the Middle Orange between this confluence and the Augrabies Falls (Fig.1.1) and the Lower Orange from the Falls to the sea. Numerous workers have studied the terraces flanking the Vaal River (du Toit 1910; Partridge & Brink 1967; Butzer *et al.* 1973; Helgren 1979b) and the Middle Orange River reaches (Helgren 1979a; McCarthy 1983; McCarthy & Tooth 2004). An investigation of the modern channel between Upington and Augrabies Falls was conducted by Tooth & McCarthy (2003). Along the Lower Orange River, comparatively less work has been undertaken. The earliest reports referring to the terraces are those of Cornell (1920) and Reuning (1931), both of whom did several reconnaissance trips into the area, whilst de Villiers & Sohnge (1959) made reference to the terraces whilst describing the bedrock geology of the area. Fowler (1976; 1982) described the terrace deposits occurring on the north (Namibian) bank on the coastal plain between Sendlingsdrif and the mouth; he also studied the occurrence of heavy minerals in these deposits. On the south bank, van Wyk & Pienaar (1986) described the deposits and the distribution of diamonds within the deposits, and Mouton (1999) investigated the sedimentology of the Koeskop deposit. The palaeontology of the Lower Orange River, famous for its Miocene fossils, has been described by many workers (Corvinus & Hendey 1978; Siesser & Salmon 1979; Pickford *et al.* 1996b;a; Pickford & Senut 2002) and reference has also been made to the Lower Orange terrace deposits in papers dealing with regional aspects of the Cainozoic geology of southern Africa (Partridge & Maud 1987; de Wit *et al.* 2000; Pether *et al.* 2000). In addition, two field guides on economic aspects of the Orange/Vaal system have been produced for conference field trips (de Wit *et al.* 1997; Ward *et al.* 2002).

Various geologists have also studied the bedrock geology within the study area (de Villiers & Sohnge 1959; McMillan 1968; Germs 1972; Ritter 1980; von Weh 1993; Jacob 2001).

2 Regional Context

2.1 Introduction

The Orange River drainage basin, the largest in southern Africa, has a total catchment area of ca. 1,000,000 km^2 of which almost 60% is inside the Republic of South Africa with the remainder falling in Lesotho, Botswana and Namibia (Fig.1.1; Table 2.1). The effective catchment is difficult to determine since it includes many pan areas and also several large tributaries which rarely contribute to flows in the main river channel (Bremner *et al.* 1990; Everson 1999).

The headwaters of the Orange River rise in the highlands of Lesotho as the Senqu River, on the eastern edge of the continent at an altitude of ca. 3100 m above mean sea level (mamsl), and flows to the west into the Atlantic Ocean. Although quoted as being 2300 km long (Everson 1999) and 2173 km (Bremner *et al.* 1990), detailed GIS-based measurement in this study measures it as ca. 2600 km from source to the Atlantic Ocean, intensive meandering in the upper reaches probably accounting for the difference. Approximately half way along its length (ca. 1300 km), the Orange is joined by its major tributary, the Vaal River that enters the right bank at an altitude of 971 mamsl. The Vaal is a longer, lower gradient river than the Orange and rises ca. 1500 km from the confluence at 1780 mamsl.

2.2 Climate and Hydrology

The Orange River catchment varies radically in climate from east to west. At the source of the Orange River in the Lesotho Highlands, the precipitation, some of which occurs as snow, exceeds 2000 mm/year, which, together with the relatively shallow soil cover and low evaporation (1200 mm/year) results in significant run-off. As the river progresses towards the west, rainfall gradually diminishes to an average of less than 50 mm/year, falling mostly in the winter, and an annual potential evaporation exceeding 3000 mm/year in the extreme west (Fig 2.1). The mean annual precipitation for the Orange River catchment is ca. 400 mm, and is considered semi-arid (Everson 1999). The average daily temperature ranges from ca. 12°C in the Lesotho Highlands to more than 22°C in the Richtersveld region near the river mouth. Extreme temperatures in excess of 50°C are often experienced in the lower Orange River valley while in the Lesotho Highlands temperatures below -10°C are common (Everson 1999).

The Orange River is classified as one of the worlds major rivers with a mean annual runoff exceeding 11 km^3/year (Milliman & Meade 1983; Bremner *et al.* 1990), although it ranks well behind the largest rivers such as the Amazon (6300 km^3/year) and Congo Rivers (1250 km^3/year). In terms of sediment discharge, the Orange is small (17x10^6 tons/year) in comparison to the Ganges/Brahmaputra (1670x10^6 tons/year) (Bremner *et al.* 1990). However, the turbidity ratio of the Orange River (obtained by dividing the mean annual sediment discharge by the mean annual runoff) is high, ranking as the fourth highest in the world (of the major rivers) and the highest in Africa (Bremner *et al.* 1990).

Presently less than half of the natural runoff reaches the Atlantic Ocean, due to numerous dams, water transfer schemes and irrigation. In addition, the area downstream of the Orange/Vaal confluence (i.e. item 5 in Table 2.1) generally experiences a nett loss of water due to high rates of evaporation (Everson 1999). Before flow regulation by dams, the river was highly erratic with flows ranging from no flow at all to large floods every 10-15 years (Wilcox 1986; Bremner *et al.* 1990; Swart *et al.* 1990) and at the mouth, an annual water level range of 5-10 m was common (Stocken, Pers. Comm., 2004). The average annual flow reaching the river mouth for the 12 year period between 1976 and 1987 is estimated to be 5700 x10^6 m^3. The flow reaching the river mouth during the wettest year which includes the major flood of February 1988 was estimated to be 26000 x10^6 m^3. The average annual flow for the two driest years on record was estimated to be only 1100 x10^6 m^3 (Everson 1999).

Thus, from a hydrological perspective, it is an unusual river in that discharge decreases downstream, as the arid western regions are traversed. In these arid western regions, the unusual situation of a large river, deriving most of its water from outside of the immediate catchment results, the result being that the trunk stream has a far greater cutting power than the tributaries connecting it to the landscape.

2.3 Regional Geology

In their journey to the Atlantic Ocean, the Orange and Vaal Rivers traverse a wide range of geological units, ranging from Archaean to Cainozoic age (Fig.2.2). The basement upon which the geological formations of South Africa have subsequently developed is the Kaapvaal Province (Craton), which occupies the central and northeastern part of the country (Fig.2.3). This crustal block is made up largely of Archaean tonalitic and trondhjemitic gneisses and granitoids, along with lesser volumes of metamorphosed volcano-sedimentary rocks (greenstone belts). The gneisses and greenstones have both been folded, faulted and intruded by younger igneous bodies(Tankard *et al.* 1982; Wilson 1998).

Within the confines of the Kaapvaal craton, a series of Late Archaean and Early Proterozoic basins developed. Those that the Orange and Vaal Rivers encounter are the Witwatersrand, Venterdorp and Transvaal Supergroups, all of which accumulated in response to the progressive northward migration of depocentres (Tankard *et al.* 1982). All of these sequences display a cyclic pattern of basin development, commencing with, and frequently terminated by volcanism. Between the volcanic rocks, thick terrigenous sequences of fluvial conglomerates, arkoses, marine quartz arenites and shales are found. Non terrigenous sediments are present only in the Transvaal Supergroup where thick deposits of carbonate sediments accumulated and overlying these are substantial deposits of banded iron formation (Tankard *et al.* 1982).

At about 2 Ga, the craton stabilised and the focus of tectonic activity within southern Africa then moved to the Namaqua-Natal Metamorphic Province (NMP), an arcuate belt which is converged onto the western and southern margins of the Kaapvaal Craton (Figs.2.2 & 2.3), where both older crystalline rocks and their supracrustal cover were affected by this orogenic activity (Tankard *et al.* 1982; Wilson 1998). The NMP forms most of the crystalline basement of southwestern Africa in a triangular shape, the longer sides of which are 850 and 700 km in length. Most of the Namaqua Metamorphic Province is a complexly deformed heterogeneous group of gneisses and intrusions metamorphosed to medium and high grade, with the ages ranging from 1700-1900 Ma (Blignaut 1977). The main period of metamorphism, the Namaqua orogeny occurred between 1400-1000 Ma (Colliston & Schoch 2000).

During the late Precambrian the Pan-African orogeny was associated with the formation of a 3000 km long chain of geosynclines around the present south and west coast. The Gariep Supergroup, and its foreland basin, the Nama Group outcropping in the study area represents this time, ranging in age between 780-545 Ma (Frimmel *et al.* 1996).

All of the above rocks have been incised by the Orange River, and could potentially provide clasts to the river.

During the Early Palaeozoic, southern Africa, laying at the heart of Gondwana was bounded in the west by South America, in the south by the Falkland Plateau, and to the east by Antarctica. Abortive rifting on the present south and east coast resulted in the accumulation of continental and marine clastic successions of the Cape Supergroup (Tankard *et al.* 1982). The closure of this aborted rift resulted in the Cape Fold Belt event, converging onto the craton to yield a foreland basin cover, the Karoo basin, which covered the craton in a northward-thinning wedge. The Permo-Triassic Karoo basin, which covers about two thirds of South Africa, comprises glacial sediments at the base (Dwyka Group), overlain by open marine, then fluvial and deltaic successions and capped by aeolian dune fields (Veevers *et al.* 1994; Visser 1996). Extensive basic and acid lavas of the Lebombo and Drakensberg Groups, dated to 183 Ma (Marsh *et al.* 1997), terminate the Karoo Supergroup. At the close of Karoo volcanism, much of South Africa was blanketed by a lava carapace which thinned towards the west (Cox 1989; Marsh *et al.* 1997). The extrusion of these volcanics preceded the fragmentation of Gondwana, which took place in two phases. Between 190 and 160 Ma, separation on the east coast followed a line down the eastern flank of the Mozambique Ridge. The second phase of fragmentation, that along west coast, was accompanied by large-scale volcanism, the Paraná-Etendeka continental flood basalt and offshore extrusive complexes, giving ages of 137-127 Ma (Turner *et al.* 1994; Stewart *et al.* 1996). Rift initiation proceeded northwards, starting in the southern South Atlantic. Between 150-130 Ma, rifting had propagated to 38°S, between 130-126.5 Ma, rifting had advanced to 28° S, and by 118.7 Ma, rifting had propagated up to the Niger Rift (Nurnberg & Muller 1991). Other workers date rifting to between 154-115 Ma, depending on which magnetic anomaly is used (Gerrard & Smith 1984; Clemson *et al.* 1997). This period of rifting and sea-floor spreading was accompanied and followed by widespread anorogenic alkaline magmatism of the kimberlitic, carbonatitic and ring-complex types (Tankard *et al.* 1982), and was the period when the majority of the diamond bearing kimberlites were emplaced (Lynn *et al.* 1998). The post-breakup evolution of the western part of southern Africa is described in Section 2.4.

2.4 Geomorphology and Drainage Development

2.4.1 Geomorphology and Tectonic Setting

The geomorphology of southern Africa is dominated by several major landforms - an elevated central plateau, a major coast parallel escarpment known as the "Great Escarpment" and a series of broad planar erosion surfaces lying above and below the Escarpment (Fig 2.4) (King 1951; Partridge & Maud 1987). The elevated central plateau, underlain by much of the Kaapvaal craton, stands at more than 1km above sea level, in contrast to the average elevation of most worldwide cratons at between 400 and 500 mamsl (Nyblade & Robinson 1994). In addition, the Atlantic Ocean basin southwest of the continent is an area of unusually shallow bathymetry. This, together with the southern and eastern Africa plateau's have been referred to as a single topographic feature, the African superswell (Nyblade & Robinson 1994). When defined in relation to global mean continental and ocean floor elevations, its mean amplitude exceeds 500 m, and within large parts of the eastern hinterland of Africa, including southern Africa, the anomaly is greater than 1000 m (Nyblade & Robinson 1994; Partridge 1998).

Although this topographic anomaly has long been recognised, few workers can agree on the causes and timing of the uplift, as well as the timing and mechanisms responsible for the formation of the great escarpment.

2.4.1.1 Field-based Ideas/Models

The geomorphic history of southern Africa has traditionally been interpreted in terms of polycyclic erosional scenarios, following the works of Dixey (1955a; 1955b) and more especially King (1951; 1962). Several erosional surfaces were distinguished, the development of which was attributed to pulses of tectonic uplift and ensuing cycles of erosion. King proposed that the isostatic response to onshore erosion and offshore deposition is the mechanism responsible for the cycles of landscape evolution and it was suggested that an isostatic response occurred only when a threshold distance of escarpment retreat of ca.. 400 km had occurred, instead of progressing continually in response to denudational unloading (Gilchrist & Summerfield 1991; Gilchrist & Summerfield 1994). Partridge and Maud (1987; 2000) modified King's original schema, and recognised three major erosional surfaces: the "African" surface of Early Cretaceous to Miocene age and two "post-African" surfaces initiated in the Miocene (post African I) and late Pliocene (post African II), respectively. At the time of rifting, the margin of southern Africa possessed substantial relief, mainly due to rift flank uplift, which preceded the breakup of Gondwanaland (Partridge & Maud 1987;2000).

Evidence taken for this high relief at the time of break-up are the high rates of terrigenous sedimentation on the continental shelf during the immediate post-rifting period (early Cretaceous), which exceeded those of the Cainozoic by an order of magnitude (Dingle *et al.* 1983). In this period of erosion, the marginal escarpment, a pre-cursor to today's Great Escarpment, receded rapidly and was within 20km of its present position by the end of the Cretaceous (Partridge & Maud 1987). The so-called African surface, which comprises two planation surfaces, above and below the escarpment was cut during this time, recognised by a deeply weathered surface with a capping of laterite or silcrete which has protected it from later dissection (Partridge 1998). Erosion up to this time was aided by a warm, wet climate. The silcretes capping the deeply weathered African surface reflect a drying-out and cooling of the climate at the beginning of the Palaeocene (Partridge & Maud 1987; Partridge 1998).

In this scheme, further, modest uplift in the Miocene (100-200m), and then massive asymmetrical uplift in the Pliocene (100-900m) supports the late development of the southern African part of the superswell (Fig.2.5).

In contrast to the previous workers, Burke (1996) envisioned Africa, including southern Africa, as a low-lying continental area like Australia today, that was covered by a deeply weathered surface that formed between about 65-30 Ma after extensive Late Cretaceous seas had retreated from the continent (Fig.2.6). There is no evidence for these Late Cretaceous-Eocene incursions above the escarpment, where evidence would have eroded away but marine deposits are preserved on the coastal strip (Partridge & Maud 1987). In this scheme, surface uplift occurred at 30 Ma, preserving the weathered surface (African surface) on top of the southern African portion of the superswell. The great escarpment formed by erosion during the last 30 My (Burke 1996). Burke's strongest line of evidence is an Oligocene erosional unconformity in the offshore, as well as a great abundance of post-30 Ma deposits elsewhere off Africa. The offshore basins flanking southern Africa, however, do not do not contain large deposits of this age, a fact attributed to the onset of desert conditions in southwestern Africa (Burke 1996).

2.4.1.2 Modelling supported by AFTT/cosmogenic nuclide data

Apatite fission track thermochronology (AFTT) studies on the west coast (Brown *et al.* 1990; Brown & Gallagher 1999; Gallagher & Brown 1999a; Brown *et al.* 2000), suggested that during the Early Cretaceous, a phase of accelerated erosion associated with uplift in the early stages of rifting and breakup led to ca. 2-5 km of denudation. The depth of denudation generally decreases inland from maximum values at the coast, but is variable (Fig.2.7a). The fission track ages (Fig.2.7b) support both an Early Cretaceous phase of enhanced denudation, with the maximum between 120 and 130 Ma, as well as a later, more variable phase of accelerated denudation during the Late Cretaceous (Brown *et al.* 2000).

The bulk of the deltaic sediments in the Orange River basin were deposited between 112-90 Ma in the south, offshore from the current Olifants River, and between 93-65 Ma offshore from the Orange River mouth (Brown *et al.* 1995). Therefore according to the

AFTT data, the bulk of the denudation in this area occurred at the time of breakup (126.5 Ma-117.5 Ma), before there was any basin to receive it, or alternatively, the sediment was accommodated below the deltaic succession.

The interpretation of fission track ages in terms of thickness of rock denuded, is strongly dependant on the value of the palaeo-geothermal gradient at the time of closure. The value of that gradient is uncertain and this produces several kilometres of uncertainty in amounts of denudation (Gunnell 2000). In addition to this the geothermal gradients could be locally variable within the study area (evidenced by thermal springs at Ais Ais in the Fish River Canyon). Factors such as lithological changes (changes from soft Karoo sedimentary cover to harder basement) and drainage basin evolution changes, could also have marked effects on cooling histories and denudation rates without any tectonic episodes being involved (Gunnell 2000).

Matching these large amounts of denudation to onshore field evidence is challenging. The presence of crater facies preserved in Late Cretaceous to Early Tertiary kimberlite pipes located inland of Springbok, indicates that there has been minimal denudation inland of the escarpment in this area during the Tertiary (Smith 1986; de Wit *et al.* 1992), this in a region where Gallagher & Brown (1999) envisage ca. 1.7 km of denudation to have occurred in the Tertiary (Fig.2.7a). A similar situation exists in Australia where AFTT interpretations of kilometre-scale denudation levels disagree with conventional geomorphic studies (Hill 1999).

On the east coast, similar large amounts of Early Cretaceous denudation are implied by a combination of AFTT data, cosmogenic nuclides and numerical modelling, ranging from 1.7-4.5 km (Fleming *et al.* 1999; Brown *et al.* 2002; van der Beek *et al.* 2002). These amounts of denudation are in stark contrast to those obtained by zeolite zonation studies, which indicate that only 200-400 m has been eroded from the top of the Drakensberg basalt sequence since its formation (Dunlevey 1993). In the same region, Hawthorne (1975) estimated that only ca. 300 m of material has been removed from the Letseng le Terai kimberlite pipe in northern Lesotho since its eruption at ca. 90 Ma.

2.4.1.3 Development of the Great Escarpment

The lower Orange drainage cuts through the escarpment on the western margin of southern Africa. Combinations of AFTT and cosmogenic isotope analysis on both the west coast (Cockburn *et al.* 1999; Cockburn *et al.* 2000) and east coast (Fleming *et al.* 1999; Brown *et* *al.* 2002; van der Beek *et al.* 2002) have both been similarly interpreted. They indicate that the escarpment did not simply retreat from the coast, but that the initial location of the present escarpment was controlled by a major inland drainage divide This was located only a few kilometres oceanward of its present position, separating low-gradient interior drainage from the higher-gradient river systems flowing to the newly established margin. Initial high rates, followed by slow rates of scarp retreat have been controlled by pinning at the drainage divide, possibly enhanced by flexural rebound. These data match the models proposed by Gilchrist *et al.* (1994), and provide no evidence or reasoning as to why there should be major drainage divides located close to the present escarpment on both the east and west coasts. i.e. Were these drainage divides pre-breakup features or features formed as result of rift-flank uplift?

King (1951) believed that the great escarpment is an amalgamation of different palaeotopographic elements. Field evidence for this claim was provided by Martin (1953) from the northern Namibian sector of the escarpment, in the Koakoland, which is incised by major Dwyka-age valleys. These valleys, now occupied by rivers such as the Kunene, Endo and Hoarasib Rivers, still contain deposits of Dwyka tillite and thus are pre-Dwyka valleys, rendering the escarpment in this area a pre-Carboniferous landform.

Martin (1973; 1976), believed that answers to the geomorphological problems lay in the offshore stratigraphy, but was hampered by a restriction in quality and coverage in offshore data available at that time. Aizawa *et al.* (2000), following the lead of Martin linked the geodynamic evolution of the offshore and onshore portions of the Namibian continental margin. The offshore seismic data shows considerable longitudinal variation in syn-rift stratigraphy and structural style, late rift topography, post rift subsidence style, and post-rift geodynamic evolution. There is only a limited development of the type of fault-bounded rift margin, envisaged by Gilchrist and Summerfield (1990; 1991; 1994), thus the evolution of the margin cannot be related in any simple way to rift-related structural development. In this scheme, there are significant differences in geodynamic and topographic evolution of the northern and southern sectors of the Namibian continental margin, and the age and origin of the Great Escarpment is notably different in these two areas (Aizawa *et al.* 2000). In the northern sector, located offshore of the Ugab River (Fig.2.8), to the north of Walvis Bay, the initial escarpment was formed in the Early to Mid Cretaceous through shoreline erosion on a subsiding margin, which continued into the Late Cretaceous as the hinge migrated inshore, causing scarp retreat to its present position by shallow marine processes. In the southern sector, on the northern flank of the Orange

Basin (Fig.2.8), rotational subsidence in the Cretaceous was matched by the sediment flux of the Orange River. A significant uplift event occurred at the Cretaceous/Tertiary boundary where at least a kilometre of uplift is implied by the truncation geometries of Cretaceous topsets in the offshore. Shallow marine bevelling at the base Tertiary led to the initiation of the Great Escarpment in this southern sector (Aizawa *et al.* 2000).

2.4.2 Drainage Development

As in the case of the geomorphology, the drainage development of southern Africa, and in particular that of the Vaal/Orange system has generated a number of contrasting views. Cox (1989) and Moore and Blenkinsop (2002) suggested that the mantle plumes associated with the Karoo and Parana plumes created domes which played a major role in controlling the development of the post-Gondwana drainage systems with drainage flowing away from the domes. However, there was already topography capable of controlling the drainage prior to these plumes. During Permo-Carboniferous times, a belt of elevated ground, termed the Cargonian Highlands, extended from southern Namibia to the east-northeast across southern Africa, separating the Karoo and Kalahari Basins, the main depocentres for Dwyka glaciation in southern Africa. Glacial transport indicators show that these highlands, which are essentially the Kaapvaal and Zimbabwe Cratons were the source regions for Dwyka Group sediments in both the Kalahari and Karoo basins. Ice flow in the Karoo basin was dominantly from the north and northeast in Late Carboniferous times and similar southwest-directed ice-flow pattern have been demonstrated for the Kalahari Basin (Visser 1983;1987; Veevers *et al.* 1994).

Following breakup, much of southern Africa would have been covered by Karoo sedimentary rocks, and, in the eastern portion of the country, by the Karoo basalts (Marsh *et al.* 1997). This wedge of volcanics would have created a drainage divide close to its present position and a regional slope towards the west.

2.4.2.1 Constraints on west coast drainage development

The Orange River offshore sedimentary basin extends roughly from north of the Orange River to south of the Olifants River in a NNW-SSE direction (Fig.2.4), and has recorded events since break-up. Between 117.5-103 Ma, an outfall from the current Olifants River system was the dominant sediment supplied to the Orange River basin (Brown *et al.* 1995). The principal sediment supply shifted northwards to the Orange depocentre at 103 Ma where the ancestral Orange River system contributed sand-rich sediments at high rates, producing major progradational and aggradational sequences between 103-68 Ma. Thus at 103 Ma, the Orange River was already a large river with a well established drainage network, feeding an offshore delta (Brown *et al.* 1995). The high offshore accumulation rates recorded in the Orange Basin during the Cretaceous phase of sedimentation in contrast to the lower values measured in the Outeniqua Basin and Natal Valley attest to the relatively large drainage basin that the Orange system had at that time (Dingle *et al.* 1983). These offshore data, together with the fact that the Orange River is a superimposed river (du Toit 1910; Wellington 1955; Maske 1957; Wellington 1958; Ollier & Pain 1997; Ward & Bluck 1997) must indicate a large, long-lived drainage. All theories addressing drainage development need to conform to these strongly backed-up interpretations.

2.4.2.2 Drainage development theories

Gilchrist *et al.* (1994) proposed a revised conceptual model of landscape and drainage evolution for rifted margins. This model is essentially an integration of ideas outlined in Summerfield (1991) and Gilchrist & Summerfield (1990; 1991; 1994), invoking flank uplift at the time of rifting, which re-orientates the drainage systems into exterior (towards the margin) and interior (away from the margin) drainage. As the exterior drainage rivers incise, the divide retreats towards the continental interior. The largest drainage systems develop and grow headward most rapidly, and at some point their headwaters breach the marginal upwarp and begin to capture the deflected drainage of the continental interior. Capture of the Fish River drainage by the headward retreating Orange River is given as an example of this. Applying this model to southwestern Africa, they conclude that rifting has controlled the denudational chronology of the margin by the process of drainage re-organisation by the rift-flank uplift. Rust & Summerfield (1990) using a combination of isopach and borehole data, noted that in the Kudu 9A-1 borehole, located approximately 100 km to the north of the present Orange River mouth, the rate of sediment accumulation increases dramatically towards the end of the Cretaceous. They interpret this to represent the onset of sediment input from the newly integrated Orange drainage system, which had breached the marginal upwarp and captured the inland-directed drainages.

A late-Cretaceous integration of the Orange River drainage network is difficult to reconcile with both the offshore data outlined above, or the Orange River being a superimposed river. The model proposed by Gilchrist *et al.*(1994) requires rift-flank uplift. However it has been established that the Namibian margin is a volcanic margin, not a classic Red Sea type rift, and thus large amounts of rift flank uplift are not expected here (Corner *et al.* 2002; Menzies *et al.* 2002). The increase in sediment accumulation rate in the Kudu borehole in the Late Cretaceous, is more likely to represent increased denudation in the Orange River drainage basin in response to an uplift event. Evidence for Late Cretaceous uplift is present in the coast offshore deposits (Aizawa *et al.* 2000) and increased denudation rates obtained from AFTT in the Late Cretaceous could also be confirming this uplift (Gallagher & Brown 1999a).

Stevenson and McMillan (2004), postulate a major drainage exit point between the Groen and Buffels Rivers based on the existence of Coniacian age (88.5-86.6 Ma) incised fluvial valleys on the continental shelf. However, this interpretation contrasts with the data of most other workers in this area which indicate a large river located near the present Orange River exit feeding a deltaic succession between 103-68 Ma (Dingle *et al.* 1983; Rust & Summerfield 1990; Brown *et al.* 1995; Clemson *et al.* 1997; Aizawa *et al.* 2000).

A number of workers have attempted to reconstruct the post-Gondwana onland drainage evolution (Fig.2.9). Du Toit (1910) observed that the Vaal, Harts and Orange Rivers, in the Vryburg/Kimberley/Prieska area (Lower Vaal and Middle Orange reaches) are currently exhuming the pre-Dwyka drainage system (Fig 2.9A). These pre-Dwyka valleys still contain Dwyka Group tillite and shales, the intervening areas being composed of more resistant Archaean rocks, mostly Ventersdorp andesites. The main artery in this system extended down what he called the "Kaap" valley, a SW/NE trending valley, named after the Kaap Plateau (now the Ghaap Plateau), from Vryburg to Prieska and continuing southwards where it becomes untraceable under the cover of the main Karoo basin. He concluded that, in the north, drainage is controlled to a certain extent by the pre-Karoo topography, whilst in the south the drainage was able to establish itself independently. By noting the occurrence of jasper and BIF pebbles in the Britstown area, du Toit (1910) concluded that the river flowed up to 200 km to the south of its present position in the Middle Orange Reach (Fig 2.9A).

Dingle and Hendey (1984) presented a model where the outlet of the Orange River alternated between the modern exit at 28°S and a southern exit located at the modern Olifants River mouth (31°S). During the Late Cretaceous, the 28°S exit was used (B1 in Fig.2.9). The drainage switched to the 31°S exit during the Paleogene (B2 in Fig.2.9), based on sediment isopach thickness ratios, clay mineral species ratios and the location of the Cape Canyon. This southward shift was apparently made possible by uplift accompanying the extensive intrusion of melilitite and pseudokimberlite plugs in a zone between Gamoep and Pofadder in Late Cretaceous time (Moore 1979), presenting a physical barrier in the Koa Valley to the northward flowing Orange River. The Orange River switched back to the northerly 28°S exit in Late Oligocene/Early Miocene times, achieved by river capture across the Gamoep-Pofadder line, and a reactivation of major northward flow along the Koa Valley (B3 in Fig.2.9). The similarity of the freshwater fish in the Orange and Olifants Rivers is another line of evidence used (Jubb 1964; Dingle & Hendey 1984; Skelton 1986). However, Rust and Summerfield (1990) report that sedimentation in the Kudu borehole (located offshore the present Orange River mouth) continued into the Paleogene, and found no evidence for the southward shift of the exit point in the offshore data. Furthermore, no typical Vaal or Orange River clasts are present in the Koa River deposits, arguing against it being part of the Orange River channel. In addition, it is highly unlikely that a large river such as the Orange would be deflected by uplift across the Koa Valley, as recent world-wide datasets have demonstrated that large rivers are able to keep pace with virtually any amount of uplift (Burbank *et al.* 1996; Wohl 1999; Sklar & Dietrich 2001).

Malherbe *et al.* (1986) favoured a Late Cretaceous "Gamoep River" (C1-C4 in Fig. 2.9) draining the hinterland from Britstown westwards via Van Wyksvlei and Brandvlei and exiting into the Atlantic through the present day Swartlintjies River (the area in which Stevenson and McMillan (2004) proposed a large fluvial outfall). During the Tertiary, the Gamoep River was captured by the Koa, and the Orange followed its course during the Quarternary.

McCarthy (1983) proposed a major Trans-Tswana River that apparently flowed from central Africa across the Kalahari to join the palaeo-Orange in the Douglas/Prieska reach (Fig.1.1). Evidence for this system is the large amount of BIF that is introduced into the river in this reach from the Ghaap plateau (McCarthy 1983). A major problem with this interpretation is that the introduction of large amounts of BIF is readily explained by Early Cenozoic tributary input by small, steep gradient streams draining the high relief area to the north of Prieska (Asbesberge), thus, no large river needs to be invoked.

de Wit (1993; 1999) and de Wit *et al.* (2000) proposed two major westward draining rivers during the Cretaceous (Fig.2.10). These were a southern Karoo River which had headwaters in the vicinity of the upper reaches of the modern Vaal and Orange Rivers and entered the Atlantic Ocean via the modern Olifants River mouth, and a northern Kalahari River, with headwaters broadly corresponding to the modern Molopo River, that reached the coast in the vicinity of the modern Orange River. During the Eocene, tributaries of the

Kalahari River captured the upper part of the Karoo River system, creating the modern-day west-coast drainage (de Wit 1993). However, the offshore data of Brown *et al.* (1995) appear to limit this Karoo River to an earlier time period (ca. 117-103 Ma). Recently, Moore and Moore (2004) dispute the existence of the Karoo River, saying that it would have been flowing across a regional gradient created by the Cape Fold Belt and Karoo foreland basin.

Ward and Bluck (1997) suggested that by the Late Cretaceous, the Vaal-Orange system had networked into a major, mature and largely meandering drainage basin that was feeding into a major delta represented by the offshore Kudu area of the northern Orange Basin. The superimposed bedrock meanders and curved course of the Lower Orange across and to the west of, the escarpment within the study area are considered to be the legacy of this Late Cretaceous system that was incised following the Late Cretaceous - earliest Tertiary uplift that is recorded in the Orange Basin offshore (Dingle *et al.* 1983; Brown *et al.* 1995; Aizawa *et al.* 2000). Therefore the Orange/Vaal system evolved from a free-meandering, large drainage system with a fine grained (sand/silt) discharge in the Late Cretaceous-Early Tertiary to a superimposed, steeper, bedrock-confined, gravel transporting channel system since the Early Tertiary (Ward and Bluck, 1997). This change in fluvial character reflected the uplift of the subcontinent and the consequent downcutting and entrenchment of the Orange/Vaal drainage, a process that was initiated by Late Cretaceous - earliest Tertiary rather than in the Mid- to Late-Tertiary proposed by Partridge and Maud (1987, 2000). The major incision phase of the Lower Orange River had been completed by about 17-19 Ma, which is the palaeontological date of the aggradational Arries Drift Gravel Formation (Corvinus & Hendey 1978; Pickford *et al.* 1996a; Pickford 1998; Ward *et al.* 2002).

2.4.2.3 Onshore Evidence

Although the Cretaceous deposits are well represented in the offshore, evidence of Cretaceous drainage onshore is limited. de Wit *et al.* (2000) states there is only 'plausible, but indirect evidence' for the existence of the Karoo River. Cretaceous age gravels are preserved in the Lichtenburg area and at Mahura Muthla on the Ghaap Plateau. At Lichtenburg, the gravels are surface lags, or are confined to karst hollows in dolomites of the Transvaal Supergroup. Late Cretaceous pollen and wood fragments are found in these sinkholes. Although sparse, there is enough evidence to suggest that this part of central South Africa had a southward-flowing drainage during the Late Cretaceous (de Wit *et al.* 2000). At Mahura Muthla, a number of sinuous channel segments totalling 4.5 km is preserved which contain silicified tree-trunks of Late Cretaceous age (Bamford 2000; de Wit *et al.* 2000). In the Vaal River area near Kimberley, the Nooitgedacht and Droogeveld deposits occur between 75-100 m above the present river and both are spread across a pre-Karoo platform in Ventersdorp lava. The Nooitgedacht deposit is typically 10-20 cm thick, consisting of pebbly, subrounded to subangular resistant clasts consisting of quartz, quartzite and agate. The deposits are laterally extensive and drape bedrock irregularities with even thickness, suggesting that they are the weathered residue of an earlier diamondiferous deposit (de Wit *et al.* 1997; de Wit *et al.* 2000). The Droogeveld gravels, unlike those at Nooitgedacht, comprises identifiable fluvial channels contained in linear bedrock fractures, or "sluits"(Spaggiari 1993; Spaggiari *et al.* 1999). Both these deposits may be Late Cretaceous in age, although this is somewhat speculative (de Wit *et al.* 2000). Other deposits in this area that may fall into the same age bracket are colluvially distributed gravels mantling the slopes. These colluvially distributed gravels have been called "Rooikoppies" or "derived gravels" and contain elements of a variety of ages. In addition to these, deposits at Mafikeng and the older primary gravel in the western Transvaal and highest terrace remnants (>100 m) along the Middle Orange may be included in this group (de Wit *et al.* 2000).

All other onshore river gravel deposits are contained within a post-African surface landscape. The long-lived superimposition of the Vaal-Orange system is recorded by the terrace remnants preserved progressively down from higher (=older) to lower (=younger) levels within the drainage basin.

1. Miocene deposits (ca. 60 m above current river level) are found in the Holpan Terrace of the Lower Vaal Basin; Terrace B along the Middle Orange; Renosterkop upper potholes and Daberas potholes in the Augrabies area; Arries Drift Gravel Formation (incorporating the pre-Proto- and Proto-Orange River gravels) along the Lower Orange reach; Bosluis Pan and Gaalputs in the Koa Valley, tributary to the Orange River (de Wit *et al.* 2000; Ward *et al.* 2002).

2. Pliocene deposits (ca. 20-40 m above current river level): Proksch Koppie and Wedburg terraces of the Lower Vaal Basin; Terrace C along the Middle Orange; Meso gravels of Renosterkop and Lower Orange River Valley; the so-called 50 m and 30 m marine packages along the Atlantic coast; younger aeolianites in the Namib Desert derived from Orange River sands, including the Fiskus sandstone (de Wit *et al.* 2000; Ward *et al.* 2002).

3. Pleistocene to Holocene (<20 m above current river level): Rietputs and Riverton Formations in the Lower Vaal Basin and along the Middle Orange..

2.5 Longitudinal Profiles

2.5.1 Introduction

Due to the fact that the study area is situated at the end point of the Orange/Vaal system, the longitudinal profiles of both rivers were analysed to establish where the study reach fits into the system. The longitudinal profiles contain both geological and geomorphological information.

Longitudinal profiles of the Orange and Vaal Rivers were constructed using 1:50 000 scale topographic maps with a 20 m contour interval (Fig.2.11). The overall gradient of the Orange River is 0.00119 m/m. Upstream of the confluence, the Orange has a gradient of 0.00162 m/m whereas that of the Vaal is 0.000534 m/m. This difference stems from the Orange River source being close to the top of the Karoo Drakensberg basalts at 3100 mamsl, whilst the Vaal starts its journey mid-way through the Karoo stratigraphy at 1780 mamsl. Downstream of the confluence the overall gradient to the coast is 0.000749 m/m. Mid-way along this reach is the spectacular Augrabies Falls knickpoint, in which the channel drops a total of 180 m in ca. 15 km, in addition to two other impressive steep "knick-zones" on either side of this. Topographic profiles situated 30 km north of both channels (Fig.2.11) reveals the flat nature of the Vaal River landscape, in contrast to the high relief, highly incised upper reaches of the Orange River. In the middle reaches, near the confluence, the relief in the Vaal River valley increases substantially, as it drops into the glacially modified pre-Karoo Kaap Valley, flanked by the Ghaap plateau. After, the confluence, the height of the topographic profile gradually decreases towards the edge of the inland central plateau. There is a large increase in topography in the escarpment zone and then a decrease onto the coastal plain. Upstream from Augrabies Falls, the Orange River occupies a shallow valley in a subdued landscape whereas after the falls, it flows in an incised gorge in an otherwise subdued landscape.

Geological contacts from the 1:1 000 000 Geological Map of South Africa were used to define the underlying geology of both river profiles (Fig.2.12). Both profiles cut their way down through the Karoo Supergroup into the underlying Archaean to Palaeo-Proterozoic rocks of the Kaapvaal Province before meeting and flowing off the craton, onto the Meso-

Proterozoic Namaqua Metamorphic Province and then finally across the Late Proterozoic Gariep Province before entering the Atlantic Ocean (Fig.2.12).

2.5.2 Gradient index and DS Plots

A river is referred to as "graded" when gradient, width and depth of its channel are in equilibrium with its load (Mackin, 1948). The gradient of a graded river usually decreases downstream with the increase in discharge, and the bed particle size decreases, defining a concave up longitudinal profile that can be approximated in terms of a simple smooth mathematical function (Sinha & Parker 1996; Morris & Williams 1999). Hack (1973) modelled the equilibrium long profile as a semi-logarithmic relationship between elevation (normal) and distance (logarithmic). This is the most common form of the equilibrium long profile in use by workers analysing long profiles (Seeber & Gornitz 1983; McKeown *et al.* 1988; Goldrick & Bishop 1995; Bishop & Goldrick 2000). For a graded river flowing across lithologies of equal erosional resistance, the plot of elevation vs the natural log of downstream distance describes a straight line, the slope of which Hack (1973) defined as the stream gradient (SL) index. The average value of the SL index for any reach xy, is given by the following equation:

$$SL_{x,y} = (h_x - h_y) / (\ln d_y - \ln d_x) \qquad \text{or} \qquad SL_{x,y} = [(h_x - h_y) / (d_x - d_y)]\, d$$

Where h_x and h_y are the elevations and d_x and d_y are the downstream distances of points x and y, and d is the distance from source to the midpoint of the reach xy (Hack 1973). The units of the SL index depend on the units of elevation being used, and in the case of this study, are in gradient metres. The gradient index is a useful way of analysing anomalous gradients independent of distance along the profile. Departures from the equilibrium profile, usually steepening, being attributed to underlying lithological changes or relative base level change. Bishop and Goldrick (2000) define an alternative to Hack's semi-logarithmic form, where a plot of the logarithm of slope against the logarithm of distance downstream should describe a straight line, which they name a DS plot. As in the semi logarithmic plot, deviations from the straight line are deviations from the idealised long profile such as steeper, lithologically controlled reaches, knickpoints or knick-zones. The DS plot, however, is more sensitive to small changes in deviations from the long profile form (Bishop & Goldrick 2000).

When studying the conventional Orange and Vaal longitudinal profiles in figure 2.12, as well as semi-logarithmic (Fig 2.13) and DS plots (Fig 2.14) of the data, and gradient and gradient index data for each reach of the rivers (Tables 2.2 & 2.3), it is apparent that both profiles are far from graded concave up profiles. There is major steepening in the lower reaches of the river, below the Orange/Vaal confluence, with some minor steepening occurring above this confluence in both rivers. Gradients, as well as SL values increase, instead of decrease downstream. This anomalous profile can either be ascribed to the decrease in discharge downstream as the more arid western areas of the country are traversed, or alternatively, the profile could be explained by lithological differences or adjustments to relative base level/tectonic events.

2.5.3 Orange River Profile

The profile of the Orange River (Fig 2.12) has been sub-divided into 26 reaches, O_1 to O_{26}. These reaches have been defined on the basis of geological or gradient similarities. The long profile of the Orange River above the Gariep Dam (Reaches O_1 to O_4) displays a classic concave up graded profile. Between this point and the Vaal River confluence, the profile steepens significantly, especially when significant thicknesses of Karoo dolerite and Venterdorp andesite (Reaches O_5 and O_7) are encountered in amongst the relatively soft Karoo shales and sandstones. Reach O_7 is the lithologically controlled descent of the Orange into the southwesterly trending, glacially sculptured Kaap valley, where the river is locked into Ventersdorp Supergroup andesite (du Toit 1910; Wellington 1958; Helgren 1979b; Visser & Loock 1988). The semi-logarithmic plot of the Orange River profile (Fig.2.13) concurs that this is a graded reach by plotting Reaches O_1 to O_4 as a relatively straight line until point X_1. The lithologically controlled steeper reaches are marked by points X_1 and X_2. After the Vaal River confluence, the Orange profile flattens out significantly as it flows along the pre-Karoo Kaap Valley on Dwyka Group diamictites and shales (Reaches O_9 and O_{10} in Fig 2.11). The profile begins to steepen slightly in Reach O_{11}as the river traverses the resistant banded iron formation of the Transvaal Supergroup and then steepens dramatically at the start of Reach O_{12}, which is the start of the Namaqua Metamorphic Province (NMP) rocks. This point is demarcated by point X_3 (Fig.2.13). The extreme oversteepening of the river after point X_2 can be seen in the convex up nature of the profile on the semi-log plot. In Reaches O_{12} to O_{23}, the steeper reaches, knickpoints and knick-zones do not appear to be directly related to lithological changes or major structures, although they all are occurring on NMP lithologies, which are dominated by granitic gneisses in these reaches.

A DS plot of the Orange River (Fig 2.14) reveals a similar pattern, although the sensitive nature of this type of plot makes it less clear than the conventional and semi-log profiles..

The points X_1, X_2 and X_3 have been placed in equivalent positions to those in figure 2.13. The extreme disequilibrium displayed by the profile downstream of point X_3 is demonstrated by this plot (Fig 2.14).

2.5.4 Vaal River Profile

The Vaal River has been subdivided into 18 reaches (V_1 to V_{18})(Fig. 2.12). In contrast to the Orange, the Vaal River is a very low gradient river almost from its headwaters, punctuated by short, steep reaches that are lithologically controlled, either by Karoo dolerites or pre Karoo lithologies (Reaches $V_{7,\ 9,10,11,13,15,17}$). This was noted by earlier workers (du Toit 1910; Wellington 1958). The semi-log profile in figure 2.13 reflects this flatness and is slightly convex upwards from the start until point Y_1, after which the river steepens significantly. Point Y_1 corresponds to the start of Reach V_9 in figure 2.12, which is the start of its journey through pre-Karoo lithologies which are exposed in the Vredefort Dome impact crater. The profile flattens out again once on Karoo Supergroup lithologies until points Y_2 and Y_3, which are the significantly steeper reaches V_{15} and V_{17}, underlain by Ventersdorp Supergroup volcanics with the latter marking a descent into the pre-Karoo Kaap Valley, now occupied by the Harts River.

2.5.5 Profiles Summary

The Orange and Vaal Rivers, downstream of the confluence are not graded rivers. Projection of the graded Orange River profile above the Gariep Dam to the present coastline results in an elevation of between 700-900 m (Fig.2.12a), which is approximately the amount of uplift which is estimated to have occurred (Ward & Bluck 1997; Aizawa *et al.* 2000).

A gradient analysis of both rivers (Fig.2.15), above and below the confluence where gradient is plotted against percentage distance of the river at that gradient. The difference between the Vaal and Orange Rivers above the confluence is very clear, where, in the Vaal River, ca. 80% of the distance that the river flows is at a gradient of <0.6 m/km (Fig. 2.15b), as opposed to the Orange, where only ca. 20% of the distance is at this gradient (Fig 2.15a). Downstream of the confluence, although there are numerous steep reaches and knick-zones, ca. 65% of the distance is spent in the low gradients between <0.6 m/km (Fig. 2.15c), similar to the Vaal above the confluence, although generally steeper.

2.6 Synopsis

- The Orange River is unusual in terms of hydology: discharge decreases downstream, as the more arid western areas of the country are traversed.

- For a large river, the Orange is unusual in that in its downstream reaches, it has a steep, ungraded channel flowing through a high-relief landscape. Most large rivers start in high-relief mountain ranges, ending in low relief sedimentary basins.

- Geomorphology of southern Africa is unusual in terms of topography, with many varying explanations for the development of the geomorphology and drainage.

- Fission track data and numerical modelling indicates an Early Cretaceous phase of uplift 3-5 km (Brown *et al.* 1990).

- Geomorphological evidence indicates a period of peneplanation and deep weathering during the Cretaceous and Eocene, followed by two phases of uplift in the Miocene and Pleistocene (Partridge & Maud 1987).

- The most likely scenario, which can be recognised in the offshore data (Aizawa *et al.* 2000) and can be seen in some fission track data is ca. 1 km of uplift at the end Cretaceous/Early Tertiary, causing the Orange to incise, superimposing itself on the basement underlying the Karoo basin sedimentary rocks (Wellington 1958; Ward & Bluck 1997).

- The offshore Orange River basin records an outfall off the current Olifants River mouth between 117.5-103 Ma (Brown *et al.* 1995), which is likely to represent the Karoo river of de Wit (1993). Activity in this basin switched to the vicinity of the current Orange River mouth between 103-68 Ma, where a large river was already established near the beginning of this interval (Brown *et al.* 1995).

Chapter 3 Study area geomorphology and drainage

3.1 Introduction and Geomorphology

The study reach between Noordoewer and the Atlantic Ocean, is where the Orange River traverses the Great Escarpment. It is a highly dissected region with up to 1200m of relief (Figs.3.1 & 3.2a & b) A digital terrain model (DTM) of the study area (Fig.3.2) was constructed from the electronic versions of the 1:50 000 scale topographic maps of South Africa which have a 20m contour interval, obtained from the South African Department of Surveys and Mapping. For the Namibian side of the river, most of which is not covered by the South African maps, 100m contour intervals were digitised from 1:250 000 scale maps. The resulting contour data was imported into the Vulcan 3-d Modelling Software package, and a DTM was created.

The study area consists of a low relief coastal plain, and a generally high relief and elevated inland portion. The high relief inland portion when viewed in conjunction with a geological map (Fig.3.3), demonstrates that relief correlates with the outcrop of more resistant lithologies. In the inland portion, the lower topographic relief areas are underlain by softer sandstones and shales of the Karoo-age Nabas basin (Fig.3.1), in contrast to the higher relief of the areas underlain by the more resistant quartzite and Namaqua Metamorphic Province (NMP) lithologies. On the South African (south) bank of the river, prominent ranges are the Stinkfontein, Rosyntjieberg and Vandersterberg Mountains, all of which are built of the Rosyntjieberg quartzites and Stinkfontein quartz-rich schists (Figs.3.2&3.3). On the Namibian side, the Hunsberge and Namusberge (berg is Afrikaans for mountain) form the high lying ground in the northern part of the area and are composed of Namaqua Metamorphic Province (NMP) lithologies capped by flat-lying Nama quartzites and limestones.

The dominant geomorphologic feature in the study area is the Orange River and its network of tributary streams (Fig.3.2). The landscape has to a large extent developed in response to the incision of the Orange, which has entrenched its arcuate form across the escarpment and coastal plain. At Noordoewer, the drop from the surrounding plains (African Surface) into the Orange River valley is 500-700m. This is likely to be the minimum amount of incision that has occurred. However, from constraints in the offshore data, this incision is likely to be up to a kilometre in magnitude, the majority of which occurred during the Late Cretaceous/Early Tertiary (Aizawa *et al.* 2000).

Thus, the river has incised into the landscape from ca. 500-1000m above its present position and at this original time is thought to have flowed in the Karoo Supergroup sandstones and shales, in a relatively flat landscape resembling the Great Karoo area in South Africa. Incision into this plateau is thought to have proceeded through the Karoo sedimentary and intrusive pile and the resistant Nama Supergroup sedimentary capping (where present) and into the NMP basement (Fig.2.4). The study area constitutes the highly dissected portion of this incision. However even though highly dissected, the summit level of the majority of NMP ground is at the 500-600 mamsl elevation. This level coincides with the base of the flat-lying Nama Supergroup in the northern part of the study area that forms a resistant cap-rock to the NMP rocks. Thus the incision has stripped away the sedimentary cover in most places and dissected the underlying basement, but post-erosion lowering of the summits appears not to have occurred (Fig.3.4). The Rosyntjieberg quartzite ridge, is higher than this summit level summit level, and is inferred to be a pre-Nama topographic high, that the Nama basin would have onlapped. In addition, post-Nama granitic intrusions, such as the Kumoos-Bremen Suite, form outcrops above this summit level (Fig.3.1).

To the northeast of the study area, ca. 90 km up the Fish River from its confluence with the Orange River, the Nama Group sedimentary capping has not been stripped off and the Fish River has incised a spectacular canyon into the landscape (Fig.3.5). The various stages of incision can be followed from a very narrow, incised canyon, widening downstream until the Nama cover is removed, and an incised landscape, similar to that flanking the Orange is exposed (Fig.3.5a-c). It is envisaged that the Fish River incision in this landscape is analogous to the Orange River incision that is now in a state of relative maturity. The geomorphology we now see is a stage in the incision history of the Orange River.

The aims for this chapter are :

1. Within the study reach, to divide the river into reaches of similar characteristics and examine the geomorphology of each reach.
2. To document the profiles of the tributaries associated with these reaches.
3. To document the Orange River surface and bottom profile and analyse the controls on these profiles.
4. To determine the response level of tributary drainages to the Orange River incision within each reach.

Many of the characteristic features of the upstream reaches of the Vaal and Orange Rivers are also present in the study area. These include superimposition across basement structure, reaches where the river is influenced by basement structure, flowing in a pre-excavated Dwyka valley, and flowing through a basin in Karoo lithologies. The superimposed nature of the Orange River can be seen in the reach between Noordoewer and Aussenkjer where it follows a meandering route through rocks of the NMP and the Rosyntjieberg mountains, built of resistant quartzite, instead of the direct route through the softer Karoo lithologies (see A in Fig.3.6 as opposed to the easier A^1 route). In addition, the meandering form of the channel, crossing the strike is further evidence for superimposition. This form would have been inherited from the free-meandering ancestral river flowing in Karoo lithologies (Wellington 1958; Ward & Bluck 1997). These loops have been superimposed onto Gariep Supergroup schists and greywackes, whereas smaller scale loops are found superimposed onto NMP rocks in the lower reaches of the Fish River, a major tributary of the Orange River in the study area (B in Fig.3.7). The course of the Orange River has, however, been modified by the basement structure in places. Good examples of these are shown at C in Fig.3.7 where a straight reach has exploited the passage of a large dyke, and D in Fig.3.7 where the reach is sub-parallel to the foliation, which also controls the trend of a number of subsidiary tributary valleys. Downstream of Noordoewer, the river flows for ca. 500 m in a Dwyka glacial valley (B in Fig 3.6), with glacial striae on the channel sidewalls, together with tillite (Fig. 3.8).

In the study reach, the Orange River valley widens downstream (see Fig.3.2; 100 & 200 m contours and Fig.3.9). Intersection with the softer Karoo basin lithologies in the Nabas basin has resulted in a broader valley. A similar analogous situation exists in the Colorado and Green Rivers in the western United States which have incised the Colorado Plateau, where resistant sidewalls produce narrow, steep-sided canyons as opposed to the wide valleys produced by rocks of low resistance (Howard & Dolan 1981; Grams & Schmidt 1999). As the Orange River valley widens, the ability to preserve terrace remnants increases and larger numbers of terraces are preserved in the downstream reaches, and in those reaches where the valley has widened due to softer sidewall lithologies (see Section 4.4).

3.2 Tributary Drainage Network

A network of tributary streams has developed in response to the incision of the Orange River, which has acted as their local base level. The tributary network and their drainage basins (Fig. 3.10) range in length from hundreds of metres to a few hundred kilometres in length (Fish River). Most of the tributaries would have originated at the margins of the Orange River, cut their way back and down into the landscape in response to the incision of the Orange River, and hence join the trunk stream at a normal angle, much like a classical trellis drainage pattern. However, some of the larger tributaries are possibly remnants of the drainage that was in existence prior to the incision of the Orange River. The best example of this is the Fish River, which is ca. 400 km long and displays very well formed meanders in its lower reaches, presumably inherited from its ancestral stream (Fig. 3.11).

The tributary network is important in that it is the interface between the incising trunk stream and the landscape and is mainly responsible for the development of the landscape. To some extent, clues to some of the incision history of the area should be preserved in the tributary network, thus, an analysis of the tributary network was undertaken.

3.2.1 Geomorphic Reach Subdivisions

The Orange River in the study area was subdivided into 7 reaches (Fig.3.12), using the differences in the Orange River valley and channel characteristics, as well as landscape, relief and bedrock differences. This was done to facilitate comparisons between the tributaries connected to each reach. The overall reach characteristics are summarised in Figure 3.13. The reaches chosen represent different geomorphologic zones and are described as follows:

Reach 1 is situated between the edge of the Nabas basin (entrance of the Orange into the NMC lithologies) and Seven Pillars at the start of the Rosyntjieberg Mountains (Fig.3.12). The bedrock flanking the channel is dominated by Richtersveld igneous suite granites. The relief is steep and dissected, and the Orange River valley is narrow and has a steep gradient with abundant bedrock straths (Fig.3.13f).

Reach 2 is situated between Seven Pillars and the start of the Karoo-filled Nabas basin upstream of Grasdrif (Fig.3.12), and is dominated by crossing the Rosyntjieberg quartzite mountain barrier. The Orange River valley is narrow and the landscape relief is high, although both the river gradient and amount of straths flanking the channel are significantly lower than Reach 1 (Fig.3.13f).

Reach 3 is a combination of the reaches in which the Nabas Basin is intersected by the river and includes the reach between Modderdrif and Reach 1, as well as the reach passing Grasdrif (Fig.3.12). In this reach, the Orange River valley widens and the landscape relief and gradient of the Orange River is low, as the soft Karoo sandstones and shales are traversed. Dolerite sills form the elevated ground flanking the channel. This is a depositional reach of the river, indicated by the lack of bedrock straths and large numbers of sediment bars per kilometre (Fig.3.13f).

Reach 4 is the stretch between the downstream end of the Nabas Basin and de Hoop. The river enters NMC lithologies and the valley narrows, relief increases and the river gradient steepens. As in Reach 1, there are numerous bedrock straths, which confine the river to a narrow inner channel. This reach has large numbers of tributary input bars (Fig.3.13f).

Reach 5 is situated between de Hoop and the Boom River, near the northern-most point of the Orange course. Although the reach has a high relief in the NMC lithologies, it has been classified separately from Reach 4 by its wider valley and lack of exposed bedrock straths (Fig.3.13f) . The few exposed straths are partially covered by gravel.

Reach 6 is situated between the Boom River and Dreigratdrif. In this reach, the Orange is joined by a number of large, south-flowing, quartzite-bearing tributaries from the Huns and Namusberge Mountains. Although the landscape relief and river gradient are steep in this reach, the valley has widened further from Reach 5 (Fig.3.13f). In addition, a large number of big gravel bars are present in this reach which flank the channel and cover any potential bedrock straths. The number of Orange River bars have increased substantially from Reach 5.

Reach 7 is the coastal plain reach between Dreigratdrif and the Atlantic (Fig.3.12). The valley is wide and landscape relief and river gradient very low in this reach (Fig.3.13f). There are no bedrock straths, and very few tributary input bars from a subdued landscape. The channel is typically covered by sediment, and is usually flanked by large gravel bars.

3.2.2 Reach Comparisons

3.2.2.1 Overall Comparison

In a comparison of the overall geomorphological data from the reaches (Fig.3.13), the Orange River valley widens progressively from upstream to downstream (Fig.3.13a). An anomalous reach is Reach 3, the Nabas Basin (Karoo) reach which, due to its soft bedrock lithology, has a very wide valley.

The gradient of the Orange River is relatively constant for Reaches 1, 2, 4, 5 and 6 (Fig.3.13b). However, the river has a significantly lower gradient in Reaches 3 and 7, where the river flows across the softer lithologies of the Nabas Basin and coastal plain respectively. Variations and controls on the Orange River gradient are discussed more fully in Section 3.2.3.

In this comparison, only the tributaries that are actively contributing gravel to the Orange River were used (i.e. those with a tributary delta). The average tributary gradient for each reach (Fig.3.13c) gradually decreases downstream. Exceptions to this are Reach 3 and Reach 4, which have anomalously low and high tributary gradients respectively for their position. The low average tributary gradient in Reach 3 is a result of the lack of relief in a catchment dominated by soft lithologies. The Orange River gradient appears to be unrelated to the average tributary gradient, apart from the obvious correlation with Reaches 3 and 7.

The average tributary drainage area per reach is fairly constant for Reaches 1, 2, 4 and 5, but is larger for Reaches 3 and 7 and is anomalously large for Reach 6 (Fig.3.13d). The larger basins in Reaches 3 and 7 are what one would expect in softer lithologies where tributaries are able to cut back further and faster than in more resistant rock types, and amalgamate into larger basins.

Reaches 2 and 4 have significantly more tributary deltas than Reaches 1, 3 and 5, and the number of contributing tributaries per kilometre in Reaches 6 and 7 decreases further as the relief is reduced to the coastal plain (Fig.3.13e). In terms of the Orange River sediment bars, Reaches 3 and 6 have the highest number of bars per kilometre, and appear to be reaches where there is comparatively more deposition.

3.2.2.2 Drainage basin areas and tributary gradient

All tributaries have much steeper gradients than the trunk stream (Fig. 3.14 & Fig. 3.15), and the former tend to have a wide range in gradients (Fig. 3.16). The gradient of the tributaries (averaged over their entire lengths) is inversely related to their drainage area (Fig.3.17), a relationship which has been demonstrated by numerous workers (Hack 1957; McKeown *et al.* 1988; Kirby *et al.* 2003). Drainage area is a surrogate for discharge, and the relationship between discharge and slope is discussed in Section 2.5.2. Thus, analysing the gradient of tributaries independently of drainage areas is meaningless and the average tributary gradient value per reach (Fig.3.13d) needs to be analysed in conjunction with their drainage areas.

Tributary catchments within each reach are subdivided, on a logarithmic scale, into groups on the basis of area (Fig.3.18). The frequency percentage of the size of tributaries per reach, as measured by their drainage basin area reveals that there is a general increase in large drainage basins in a downstream direction. This is consistent with the fact that the incision of the Orange would have commenced at the coast and worked its way upstream, thus allowing the most downstream reaches more time to develop. Also, this downstream region is near to the region of base level change, and would be affected by marine planation on rising sea levels. Once again, Reaches 3 and 4 are exceptions to this trend, with Reach 3 giving an identical signature to the most downstream Reach 7, which can be explained by the relatively soft lithologies of Reach 3. Reach 4 has an anomalously high number of small drainages for its position downstream, which can be partially explained by its resistant bedrock lithologies, but not fully as the geology does not differ significantly from Reach 5.

Gradients of the tributaries were analysed per reach, within each catchment area subdivision. Graphs of the average gradient for each drainage area interval per reach (Figs. 3.19a-d) show that for the very small tributaries (Fig.3.19a), Reaches 2 and 4 have the highest gradients. For the drainage basin area 1×10^6-$10 \times 10^6 m^2$(Fig.3.19b), Reach 1 has the steepest gradient. For the drainage basin area 10×10^6-$100 \times 10^6 m^2$ (Fig.3.19c), Reach 2 has the highest gradient and for the very large drainage basins (Fig.3.19d), Reaches 5 and 6 have the highest gradients. Thus, the data can be summarised as follows (see Fig.3.20): as the size of the tributary drainage basins increase, the overall gradient of the tributary channels decrease. Gravel is supplied to the Orange River by many smaller, steep tributaries in the upstream reaches (Reaches 1, 2 and 4), whereas in the downstream reaches (Reaches 5, 6 and 7) and Nabas Basin (Reach 3), gravel is supplied mainly by the larger drainage basins. For comparable drainage basin areas, smaller basins have steeper gradients in the upstream reaches and larger basins have steeper gradients in the downstream reaches.

The controls on the differing gradients and drainage areas between reaches must be time (i.e. stage of development), lithological factors and the complications related to inherited drainage. The trend of increasing drainage basin area downstream is likely to be the effect of base level change generated at the coast. Although aspects of this trend can be explained by the softer bedrock present in Reach 7, the drainage areas of Reaches 5 and 6, which are in competent lithologies, are best explained by their proximity to the base level change. In these downstream areas, smaller tributary basins are scarce as they have had longer to develop and amalgamate. The small basins in the upstream reaches could be time-equivalents of the larger downstream basins. The large basins in the upstream reaches, which have lower gradients than equivalent size basins in the downstream reaches could well be old drainage, which was inherited from the pre-incision drainage.

3.2.2.3 Tributary profiles

To investigate if the shapes of the tributary profiles vary between the reaches, tributaries with similar sized drainage areas were plotted on the same set of axes to facilitate direct comparison. Features on the profile such as concavity, convexity or the presence of knickpoints were assessed. Knickpoints are convex breaks in the slope in the profile, which separate upper and lower graded reaches (Gardner 1983).

Tributaries were digitised from the 1:50 000 scale topographic maps and longitudinal profiles of these tributaries were drawn from the intersection of the tributary with the 20 m contours. To test whether this digitising method was representative of the actual tributary profile, five tributaries were surveyed with a Trimble differential GPS system. In a comparison between the two methods, for the most part, they compare favourably (Fig. 3.21). The smaller-scale knickpoints shown in figures 3.21d) & e) are not detected by the 20m contour intervals but the larger scale features and knickpoints are present in the digitising method, thus it was deemed satisfactory for the present investigation.

Profiles of the tributaries per comparable drainage area, which are colour-coded per reach are presented in figure 3.22A-L. The small inset graph present in each graph is the average gradient per reach for the data present on the graph. Profiles for the very small drainage basin areas, <1 000 000 m^2 (gradient data on Fig.3.19a) are represented in Figs.3.22A & B. The tributaries in this category are short and steep and are invariably straight or convex. They are only abundant in reaches 1, 2 and 4, where the Orange valley is very confined. Profiles for the drainage basin areas $1x10^6m^2 - 10x10^6m^2$ (Figs.3.22C-E) show the relative flatness of the tributaries in Reaches 3 and 7 and the steepness of the tributaries of Reach 4, especially within 2-3 km of the Orange River. Large knickpoints on the tributaries are rare. The knickpoint shown on tributary Bar AL (Fig.3.22C) is due to a transition from quartzite to mafic volcanic rocks whereas the knickpoint on tributary Bar BT (Fig3.22D) is due to another lithological change from Orange River Group felsic volcanic rocks to granite. The knickpoint on tributary Bar EQ (Fig.3.22E) is at a much more subtle lithological change, located in the Gariep schists and greywackes where a more resistant quartz-rich unit is encountered. Reach 7 is the only reach that displays knickpoints which are located close to the Orange River, which could indicate that recent incision in the Orange River has re-activated these tributaries in the lower reaches of the river. Photos of some of these knickpoints in Reach 7 are displayed in figure 3.23.

Profiles for the drainage basin area $10x10^6m^2$ – $100x10^6m^2$ are shown on figures 3.22F-I. The relative flatness of the tributaries in Reaches 3 and 7 can be seen as well as prominent knickpoints in the lower reaches of the tributaries of Reach 7. Another interesting feature is the behaviour of different tributaries in similar bedrock conditions. Tributary Bar AM (Fig.3.22F), which is situated in the same region as Bar AL (Fig.3.22C) and at the same lithological contact, has a completely different shape to its profile. The differences in these profiles could be highlighting the difference between an existing drainage (Bar AL) and one originating at the Orange (Bar AM). The profiles of these tributaries, as well as the position of the geological contacts are shown in figures 3.24a & b. A possible remnant of pre-incision landscape or a pre-existing drainage (prior to incision) is the profile of Bar AH, which is located in the Rosyntjieberg Mountains (Fig.3.22G). This profile is very flat in its upper reaches, and then the profile encounters first a major, and then a minor knickpoint. Neither reach between the knickpoints nor between the Orange and the 2nd knickpoint is as flat as the upper reaches. A projection of the profile of the upper reaches of Bar AH to the Orange, brings the height of interception with the palaeo-Orange to between 550-600mamsl (Fig.3.24c). Thus, the flat upper reach may be a remnant of the pre-incision landscape and drainage. Alternatively, the shape of the profile could be simply lithologically or structurally controlled but its unique shape relative to the other profiles in the area (Fig.3.22K) suggests different control. The likelihood of finding remnants of a pre-incision landscape should be the highest in the most resistant rocks, as they would be the slowest to change. Thus, it is probably no coincidence that the anomalous profiles are found in this area of the very resistant Rosyntjieberg quartzites. In addition to this, summits of the Rosyntjieberg Mountains appear distinctly planar (Fig.3.25). These planar features are not artefacts of flat-lying sedimentary cap-rocks, as the quartzites building the mountains are intensely folded, thus, they may well be remnants of an existing plateau, prior to incision by the Orange, or even a plateau which was planed by the Orange prior to or during incision.

Profiles for the tributaries with large drainage basin areas - $>100x10^6m^2$ (Figs.3.22J-L) show the shift towards the steeper profiles in the more downstream Reaches 5 and 6, as well as a knickpoint on the lower reaches of Bar EM (Fig.3.22J and Figs.3.23g & h).

Analysis of the percentage of graded profiles (smooth, concave-up profiles) per reach (Table 3.1), reveals that Reach 3, possessing the softest lithology has the highest percentage of graded profiles (36%). Reach 7, also in soft lithology has one of the lowest percentages of graded profiles (17%), due to the numerous knickpoints on these profiles, which are almost certainly a response to incision of the Orange. However, when this incision occurred is unconstrained and could range from Recent to Miocene times. Apart from these examples, there is no obvious pattern in the percentage of graded profiles.

3.3 Orange River Trunk Stream

3.3.1 Surface and Bottom Profile

Bedrock channels play a key role in landscape evolution. The ability of streams to incise through bedrock ultimately sets the rate of lowering of the landscape in the drainage basin, and therefore mass removal (Snyder *et al.* 2000; Whipple 2004). Thus an understanding of the controls on the bedrock channel and the development of the longitudinal profile is important.

The longitudinal profile of the Orange River in the study area was plotted with data from both the 1:250 000 and 1:50 000-scale maps (Department of Surveys and Mapping, South Africa). Due to the low gradient of the Orange River, the detail obtained from these profiles was insufficient for meaningful analysis, and thus a more detailed survey was done. Using a Trimble differential GPS mounted in an inflatable boat, an accurate survey of the water surface was done between Noordoewer and Oranjemund. The survey took place over ca. 300 km, thus it was necessary to use a cm-precision GPS base station located at various control points and to post-process the data. A reading was taken every 5 seconds, thus providing more detail in the steeper reaches which are more difficult to negotiate. A total of 10751 points were measured, ranging from 5 m between points in the rapids, to 50 m between points in the flat reaches. The relative precision of the survey was approximately 0.1 m in the horizontal and 0.3 m in the vertical. The difference in results between the three methods (Fig.3.26) demonstrates the greater detail obtained by the GPS survey. In addition to this, a separate survey was done with an Autohelm echosounder, coupled with a hand-held GPS to measure the bottom profile of the thalweg. The position of each depth measurement was noted and this depth was then used in conjunction with the GPS surface profile. Unfortunately, it was not possible to differentiate between bedrock and sedimentary cover whilst taking these measurements, thus it is not the true bedrock profile of the channel. Figures 3.27 & 3.28 illustrate the equipment used in each survey. The data for these surveys can be located in Appendix A.

The striking feature of the surface profile (Fig.3.29) is its stepped shape, especially in the "Richtersveld" reach between the entrance to the Richtersveld and Dreigratdrif. Here it consists of a number of very low gradient segments connected by higher gradient reaches, mostly in the form of rapids. A similar situation is seen in the Grand Canyon reach of the Colorado River, in the western United States, where 50% of the drop in elevation takes place in only 10% of the distance (Leopold 1969; Dolan *et al.* 1978). A common characteristic of most bedrock channels is a downstream alternation between relatively narrow, high gradient reaches and wider, low gradient reaches that commonly have an alluvial veneer or fill (Wohl 1999). The striking feature of the bottom profile (Fig.3.29) is its rough "saw-tooth" appearance, where depths of up to 26 m were encountered. In some cases, the deep pools have an equivalent depth to the surface of the river approximately 30 km downstream. In figure 3.30, the geomorphic reaches 1-7 (as discussed in Section 3.2.2) are shown on the profile, together with the average surface gradient of each reach.

Steepening of the local channel gradient can result from one of several causes, including, more resistant bedrock strata; the introduction of a coarser or larger bedload; tectonic activity; or the effect of past events, notably a fall in base level (Knighton 1998).

3.3.2 Geological controls on river gradient

To evaluate the bedrock geology controls on the gradient, the river profile was plotted along with the major geological contacts (Fig.3.31). On a broad scale, the gradient of the profiles matches the geology well. When the river enters the NMP rocks, at the entrance to the Richtersveld reach, a significant steepening occurs. At the end of the Richtersveld reach, at Dreigratdrif, when Gariep Supergroup metasediments are encountered, the river profile flattens off significantly. Within the Richtersveld reach, significant flattening of the profile occurs when Karoo Supergroup sandstones and shales and Rosyntjieberg Group quartzites are encountered, the softest and hardest lithologies in the study area respectively. Within the NMC rocks and intrusive granitic rocks of the Richtersveld igneous complex and Kuboos-Bremen intrusives, the changes in gradient are more difficult to distinguish visually on the profile. However, when analysing the numerical gradients and stream gradient indexes from each geological reach, it becomes apparent that there is significant correlation of gradient with lithology (see Fig.3.32b & c, as well as the plan view indicating the location of the geological reaches). Gradients and SL values (gradient index

– see Section 2.5.2) of the de Hoop subgroup metavolcanic rocks are consistently the highest of all the lithologies. At the low gradient end of the scale are the sedimentary rocks of the Karoo, Nama, Gariep and Rosyntjieberg Groups. In between are the granitoid rocks of the Vioolsdrif, RIC and Kuboos-Bremen suites.

The connection between lithology and gradient has been well established, with steeper gradients being associated with harder lithologies. Hack (1957; 1973) established that SL values and hence gradient increased markedly whenever more resistant bedrock was encountered on many rivers in the Eastern United States. Goldrick and Bishop (1995), working in southeastern Australia confirmed these findings. In a study of transverse rivers in the Himalayan region, Seeber and Gornitz (1983) found good correlation between erosion-resistant rocks and high river gradients, and between easily erodable rocks and low river gradients. However in the Himalayas, the role of differential erosion due to lithological changes is secondary to the role of tectonics in shaping the profiles of the rivers (Seeber & Gornitz 1983). In the Vaal and Upper Orange River catchments, the connection between gradient and lithology has been well established (du Toit 1910; Wellington 1958; Tooth *et al.* 2002) and demonstrated in Section 2.5. To maintain equilibrium, incising rivers tend to steepen, when encountering a reach of more resistant bedrock relative to the less resistant reaches. Sufficient power is consequently gained to eliminate the obstruction so that a graded profile can be maintained, and that relative base level fall can be transmitted along the entire length of river (Merritts & Vincent 1989; Merritts *et al.* 1994). In incising rivers, rocks are eroded by a combination of plucking, abrasion and cavitation, depending on the properties of the bedrock involved. Plucking dominates when rocks are well bedded, jointed or fractured on a sub-metre scale, whereas abrasion and cavitation tend to occur in more massive rocks (Whipple *et al.* 2000). In an experimental study of rock erodability, Sklar and Dietrich (2001) established that rocks with a higher tensile strength erode the slowest, in the absence of pervasive jointing. In their study, quartzite, andesite, greenstone, welded tuff and granite were the hardest rock types, as opposed to the softer mudstone, sandstone, limestone and greywacke (Fig.3.33). Thus, the findings from this study correlating the steepest gradients with metavolcanic and granitoid rocks are to be expected (Fig.3.32c & d). The extremely low gradient of the reach through the highly resistant, but well jointed Rosyntjieberg Group quartzites may point to the importance of erosion by plucking in this reach. Other geological variables influencing the gradient of the river could be the direction of flow relative to structure. The dominant orientation of the structural fabric in the study area is roughly north-south. In the Richtersveld reach, reaches were separated into north-south (N-S) and east-west (E-

W) components. The average gradient in the N-S reaches is 0.00065 m/m, whereas in the E-W reaches, where the river flows across the structural grain, the gradient is 0.00096 m/m (Fig.3.34). An example of the effect of direction of flow on the gradient within one lithological type comes from the Richtersveld igneous complex (RIC), which has an overall gradient of 0.00086 m/m. The N-S flowing sections of this reach have an average gradient of 0.00077 m/m, whereas the E-W sections have an average gradient of 0.00115 m/m. This pattern is repeated in Reach 22 (Fig.3.32b) where a gradient of 0.00123 m/m is achieved by the E-W flowing Vioolsdrif Suite gneisses as opposed to the average gradient of 0.00069 m/m for this lithology which is predominantly N-S flowing in its other reaches. These results are in agreement with those of Goldrick and Bishop (1995), who found that gradients increased when the river flowed across the grain of bedrock. However, the metavolcanic lithologies in this study tend to have steep gradients regardless of the direction of flow.

3.3.3 Tributary input control on river gradient

The water surface profile has a stepped shape with many small-scale "knickpoints" (Fig.3.29). In reality, the majority of these steps are sites of turbulence or rapids that the river is removing to achieve grade. Although there is good overall correlation of gradient with lithology, changes in lithology are generally not responsible for the steps in the profile. The few lithological contacts, directly responsible for sudden changes in gradient, are illustrated in figure 3.35 (green lines below the profile). Many more rapids are associated with tributary input bars (deltas) where the much steeper tributaries join the main channel. The red lines above the profile (Fig.3.35) are the tributaries responsible for rapids or steps in the profile. Although the tributaries do not contribute water to the trunk stream on a regular basis, many of them deliver large quantities of coarse gravel to the Orange River channel on the occasional flash flood. These coarse gravels are deposited in the main channel, where they constrict or even temporarily block flow. Most of the gravel is redistributed by the main channel, however, the large immovable blocks remain or are moved only small distances and will remain on the bar until broken down into smaller, more easily movable clasts. Tributaries in arid areas, with the large amount of sediment available, tend to yield debris flows on flooding and are able to transport much coarser clasts than the trunk stream due to a combination of their steeper slopes and high viscosity of fluid. The small density difference between boulders and their transporting fluid medium assists the flow in carrying otherwise immovable blocks (Webb *et al.* 1988). An eye witness account of a tributary in flood in the study area (Fig.3.36a) described it as a loud roar which continued for a full day whilst boulders knocked against each other in the high density medium (Botes, Pers. Comm., 2002).

This situation is comparable to the canyon rivers in the Colorado Plateau region, where much work has been done describing and analysing the interactions between tributary deposits and the trunk stream on the Colorado River (Dolan *et al.* 1978; Howard & Dolan 1981; Kieffer 1985; Webb *et al.* 1988; Rubin *et al.* 1990; Schmidt 1990; Schmidt & Rubin 1995; Webb *et al.* 1996; Cerling *et al.* 1999), Green River (Graf 1979; Grams & Schmidt 1999) and Yampa River (Hammack & Wohl 1996). Leopold (1969) suggested that in the Grand Canyon, the Colorado River exhibits a uniform morphology along its length, including regular spacing of pools and riffles and that these are a necessary part of the equilibrium of the river. Graf (1979) investigating the occurrence of rapids in the Colorado and Green Rivers concluded that the spacing of rapids is essentially random, and that the majority (92%), but not all rapids investigated were created by tributary fans or mass movement processes. Dolan *et al.* (1978), presented data from the Grand Canyon relating most rapids and deep pools to the positions of tributary debris fans, which partially block the main channel. The tributaries themselves exploit structural weaknesses in the bedrock and thus the fluvial organisation is ultimately structurally-controlled (Dolan *et al.* 1978). Grams and Schmidt (1999), working on the Green River in the eastern Uinta Mountains, concluded that all rapids are formed by constrictions that are caused by debris fans, or downstream gravel bars derived from the fans, but that not all debris fans are associated with rapids (Grams & Schmidt 1999).

In the Colorado River, the recurrence interval of the debris flows that created and maintain the formidable Lava Falls rapid is 15-60 years for the smaller flows, whereas the large flows recur at a frequency of several hundred to several thousand years. The highest and oldest debris flow which blocked the river completely and aggraded the bed ca. 30 m for an unknown length of time, yielded a cosmogenic date of ca. 3 Ka (Webb *et al.* 1988; Cerling *et al.* 1999). Although the main river re-works these deposits, boulders that form the core of the rapids are essentially immovable, as evidenced by historical photographs (Webb *et al.* 1996). Cosmogenic surface exposure ages from Hawaii indicate that large boulders remain below knickpoints for hundreds of thousands of years (Seidl *et al.* 1997), and, although not directly comparable to the Colorado or Orange Rivers, these give an indication of the residence time of very coarse gravel on the bars. On the Colorado River, dams have reduced the frequency of flooding, thus numerous rapids are becoming steeper and more treacherous as tributaries are actively depositing material on the debris fans and are no longer being re-worked to the same extent by the main river (Kieffer 1985).

One of the principal differences between the Orange River and the canyon rivers is that bedrock is present in most rapids in the Orange River (Fig.3.36b), whereas in the Grand Canyon and Green Rivers, only boulders from the tributaries are present (Webb *et al.* 1996; Grams & Schmidt 1999). In the Green River, boreholes confirm that the river does not flow directly on bedrock, but 12-45m above bedrock, thus it is presently in a state of aggradation (Grams & Schmidt 1999). In the Grand Canyon, evidence from lava flows which dammed the canyon, indicate that the size and shape of the canyon has not changed for more than a million years. Some of the remnants of the lava dams extend down to present river level, thus the river removed the lava barrier down to its original profile, but no further (Hamblin 1994; Dalrymple & Hamblin 1998). Thus, these rivers are expending their erosional energy in abrading and removing boulders, not in eroding bedrock (Webb *et al.* 1996). Hence, for those rivers that are not in direct contact with bedrock, bedrock type does not control the channel gradient directly. Steep gradients occur where there is a high abundance of debris flows, which in turn are only found in canyons where the bedrock is most resistant and the fans are composed of resistant boulders, thus, resistant bedrock influences channel gradient indirectly. In these cases, channel slope has increased so that streamflow competence matches the characteristics of tributary supply (Grams & Schmidt 1999).

Thus, the introduction of coarse gravel is the first way in which tributaries control the gradient. In alluvial gravel bed rivers, equations relating to channel slope commonly include relationships between slope, discharge and bed material size (Knighton 1998). Mean grain size of bed sediment is commonly proportional to channel slope (Leopold & Wolman 1957; Prestegaard 1983; Bridge 2003). Mackin (1948) predicted the following :

> "if a graded stream receives an influx of debris then part of this influx is deposited at the point of influx. This causes the channel to be steepened immediately downstream allowing increased transport of load, some of which is deposited in the next segment downstream, and so on. This down valley wave of deposition causes a general increase of slope until it is everywhere adjusted to the transport of all the debris delivered to it" (p.493).

Alluvial rivers are more able to respond and adjust to conditions which are imposed on them than are bedrock dominated rivers, and their gradients are likely to more sensitive to coarse clastic additions. Steepening of the gradient of the Lachlan River, a bedrock river in eastern Australia, occurred after the addition of coarse clasts at two localities (Bishop *et* *al.* 1984). Hack (1957), in the eastern United States found no simple relationship between channel slope and median bed material size. Two examples where the addition of coarse gravel appears to influence the slope of the Orange River are illustrated in figure 3.37. In both these examples, the flattening of the profile well after the contact with softer bedrock is controlled by the occurrence of large coarse gravel bars in the river. The profile flattens out with the fining and disappearance of gravel bars in both instances. Downstream fining of gravel is discussed more fully in Section 5.

The second and main control that the tributaries exert on the river is that the tributary input bars are inhibiting bedrock incision at the site of the tributary delta. On both the canyon rivers and Orange, tributary input points cause most of the active sites of turbulence, thus, before the river can attain a smooth profile, it needs to eliminate the material present in these rapids. However, this situation will persist whilst the tributary is still supplying coarse material to the rapid. The continuous supply of gravel mantles and protects the bedrock, and once it is removed, is replaced by subsequent debris flows, thus the gravel behaves like "renewable" bedrock. Only when the tributary slopes have flattened out sufficiently to prevent the transport of coarse, clastic material to the rapids, will the river be able to eliminate the boulders in the rapids, and then eliminate the bedrock step in the channel. The fact that bedrock is ubiquitously present in the Orange River rapids which are caused by tributary deltas and absent in adjacent channel reaches is proof of the protective role of tributary delta gravel, and that the bedrock step has not been eliminated by the river. The end result is a set of mini "knickpoints" that the river has to deal with whilst the tributary supplies gravel. This is an end-member of the "coverage effect" as outlined by Sklar and Dietrich (2001) where partial burial of the bedrock substrate beneath sedimentary deposits limit incision rates. Thus, the extremely flat reach whilst the Orange River is traversing the Rosyntjieberg quartzites can also be explained by the presence of a large tributary delta knickpoint at the downstream end of the reach. This "holding up" of the knickpoint by this tributary is likely to continue until the tributary supplying the coarse material ceases to do so. Two examples of these "tributary knickpoints, where the relationship of the long profile shape to the position of the geological contacts and tributary input points is shown in figure 3.38.

3.3.4 Controls on bedrock incision

Depth data (Fig.3.32c & e) tends to mirror lithological strength. The deepest average depths and highest depth variation (standard deviation) are encountered in the Rosyntjieberg quartzites, followed by the granitic and metavolcanic rocks. The shallowest depths occur in the softer Karoo and Gariep sedimentary and metasedimentary rocks. This data correlates well with that of Howard and Dolan (1981) from the Grand Canyon where the softer areas in shale yielded flat, shallow reaches and the harder crystalline rocks yield an uneven bed topography with numerous deep pools. This is consistent with observations made in palaeo-deposits, where mining has exposed large areas of palaeo-strath and channel deposits. Softer lithologies tend to be planed off, whereas harder lithologies display higher bedrock relief.

In the study reach, there is up to 25 m of relief on the channel bed. Deep holes are scoured into the bedrock whenever the channel width is restricted. This generally occurs in reaches of resistant bedrock (where the channel takes on the form of a linear inner channel), on the outer bend of meanders, or in association with tributary input bars. There are a total of eleven areas which have scoured holes deeper than 16m in the study area. Five of these are associated with inner channels, three with tributary deltas and three occur on meander bends, but also have tributaries in attendance.

Inner channels are deep, narrow bedrock channels that are flanked by bedrock straths on one or both sides of the channel (Wohl 1999). Inner channels have been produced experimentally under conditions of steep gradients and their development follows a progression from longitudinal lineations, ripples and potholes that enlarge into prominent grooves which coalesce to form a narrow, deep inner channel, with deeper channels being produced at higher gradients. This inner channel eventually conveys the entire flow, leaving part of the previously scoured channel floor above the water surface (Shepherd & Schumm 1974; Wohl & Ikeda 1997). Inner channels have been described from a variety of areas and bedrock types, and are always found in steep reaches (Baker & Pickup 1987; Wohl 1999; Wohl & Merritts 2001). This could be due to these reaches having more resistant bedrock, where the river maximises stream power per unit area by incising deep, narrow inner channels (Wohl 1992). However, the position of inner channels do not always correlate with rock strength, as evidenced by Wohl (1993), but also signal actively incising reaches, initiated by base level change, where the mechanism of incision is the creation of inner channels. In general, the presence of differential erosion at the cross-sectional scale implies that a bedrock channel is incising (Wohl 1999).

In the Orange River, deep inner channels are only found in steeper reaches and in areas of more resistant metavolcanic, granitic or gneissic bedrock and are flanked by well-defined bedrock straths (Fig.3.39). Thus the presence of inner channels could be signalling either the presence of incising reaches, or the adjustment of the river to the varying bedrock conditions. Another complication when dealing with resistant bedrock is that relict erosional features may persist along a channel long after the conditions which created the forms have changed (Wohl 1999). Thus, the reaches where inner channels are present may not be indicating current incision, but could be recording past incision events.

Closer inspection of the inner channel in the Sambok area (Fig.3.40) reveals numerous features of incision such as small and large scale fluting, ripples, incipient potholes and potholes similar to those described by Whipple *et al.*(2000) from the incising Indus River. Further evidence that this reach is undergoing incision, and that the knickpoint has migrated upstream is indicted by the freshness and abundance of these features at the upstream side of the inner channel, and lack of them further downstream, away from the knickpoint.

According to Whipple *et al.* (2000), the discontinuous strath terraces which are commonly preserved along bedrock channels (Wohl 1992;1993) are a result of the interaction of plucking and abrasional processes. Well-jointed reaches are dominated by plucking processes, which erode more rapidly than adjacent more massive reaches, which are eroded by slower abrasional processes. The downstream ends of these more massive rock ribs form prominent knickpoints, and the reach upstream of this knickpoint is bevelled into a wide strath until the knickpoint cuts a narrow inner channel in the strath and retreats upstream, leaving a discontinuous strath which converges with the channel bed in the upstream direction. The 3-dimensional distribution of these massive rock ribs will limit the rate of channel lowering and the equilibrium slope (Whipple *et al.* 2000). However, this model may be applicable to smaller bedrock rivers but is questioned for this study reach as very few of the present knickpoints appear to be lithologically related. Large gravel-bearing rivers such as the Indus tend to act as "bedrock saws"(Burbank *et al.* 1996) and the Orange River appears to do likewise.

In the Colorado and Green Rivers, most deep holes are associated with the rapids caused by tributary bars (Dolan *et al.* 1978; Howard & Dolan 1981; Grams & Schmidt 1999). The channel constriction causes an increase in hydraulic energy (Kieffer 1985), which results in scouring downstream of the constriction. The same effect is observed in the Orange River, however the association is not as strong as in the more confined canyon rivers. (Fig.3.41).

3.4 Geomorphic Reach Table

Data for both tributaries and trunk stream per Geomorphic reach is summarised in Table 3.1. The values included in this table are colour-coded, that red and purple indicate high values and grey and yellow low values. The link between the surrounding landscape and trunk stream properties within each Reach is clear. Reaches 1, 2 and 4, which are underlain by resistant lithologies consistently display high values (tributary and trunk stream gradient, depth, clast size, no. of tributaries and strath percentage), whereas 3 and 7, underlain by less competant lithologies display low values. Exceptions are in the categories marked in black fill (drainage area, valley width, percentage terraces), where low values are obtained by the more competant lithologies. The mixture of high and low values obtained for Reaches 5 and 6, also situated in competant lithologies, can be explained by their downstream position, within the influence of eustatic change.

3.5 Summary

The study area consists of a low relief coastal plain and a high relief and elevated inland portion, forming part of the southern African Great Escarpment. The geomorphology of the area is dominated by the Orange River trunk stream and its system of tributaries. The Orange River is a superimposed river (Wellington 1958), as evidenced by meanders cross-cutting basement structure and a preference for following tortuous routes through rugged terrain, where easier alternatives would have been taken by a back-cutting river. However, some reaches display modification by basement structure and one short reach flows in a Dwyka-excavated valley. The incision of the Orange River, most likely to have occurred at end-Cretaceous/Early Tertiary, is envisaged to have activated a system of tributaries that would have originated at the Orange River and cut back into the low relief landscape as incision proceeded. However, some of the large tributaries, such as the Fish River would have undoubtedly existed prior to the incision event.

The Orange River incision is illustrated by downstream changes in its largest tributary in the area, the Fish River, which hosts a spectacular canyon 500-600 m deep in places. Upstream of the canyon, the Fish River flows in a broad, shallow valley, analogous to the Late Cretaceous Orange River before incision, whereas the canyon reach is analogous to the early incisional phase of the Orange River and the reach between the canyon and the Orange River represents the stage of relative maturity that the landscape flanking the Orange River channel is currently in.

The Orange River valley responds to changes in the surrounding geology, but generally widens downstream. To investigate changes within the study area, the trunk stream was sub-divided into 7 geomorphic reaches, enabling a comparison to be made between the trunk stream, as well as the tributaries associated with these reaches (Table 3.1).

Comparison of the reaches reveals that tributary gradient declines gradually downstream, and is lowest in areas underlain by softer rocks. Apart from areas of softer lithologies in the Nabas Karoo basin and on the coastal plain, there is no apparent connection between the gradient of the trunk stream and the tributary gradient in a reach. Drainage basin size tends to increase in a downstream direction and, as gradient and drainage area are related (Fig.3.17), it was necessary to analyse gradients per drainage area (Fig.3.18, Table 3.1). This analysis reveals that the reaches possessing high relief which are situated in resistant rocks have the highest gradients in the small tributaries, whereas in the large tributaries, the downstream reaches have higher gradients. These data appear to be consistent with base level changes moving up the valley, reaching the downstream areas first, hence the relative lack of small-sized tributaries (which have cut back, expanded and amalgamated) in these reaches. The time equivalent tributaries in the upstream reaches are still small, with steeper gradients. The large tributaries in the upstream reaches with flatter gradients may have been inherited from the ancestral, pre-incision drainage.

Tributary profiles from the reaches were compared (for similar sized drainage areas). Knickpoints in the profiles in the upstream reaches are fairly rare, and only occur at lithological changes, whereas in the coastal plain reach, knickpoints are more common, occurring at seemingly much less significant changes in lithology, thus, these coastal plain profiles appear to have responded to a recent incision of the Orange River. There is no pattern in the percentage of graded profiles per reach (Table 3.1), although the Karoo basin has the highest percentage, followed by Reaches 2 and 5. In the Rosyntjieberg Mountains, two tributaries respond in very different ways to the same geological contact, which may be indicating the difference between an inherited tributary and one that has worked its way back from the trunk stream. In the same area, a possible remnant of the pre-incision drainage was identified which grades to a level of approximately 600 mamsl (Fig.3.24c), which would indicate that the Orange River has incised ca. 500 m in this area, the minimum estimate of incision that is likely to have occurred.

In surface and bottom surveys of the modern Orange River channel, it is found that the river gradient is controlled by changes in the lithology, as well as structure. The more resistant, massive lithologies such as the mafic and felsic volcanics, as well as granite and gneiss have the highest gradients, whereas the sedimentary and metasedimentary rocks, including the very resistant Rosyntjieberg quartzites have the lowest gradients, possibly indicating plucking as the dominant erosional mechanism. Gradients are generally higher in reaches that flow across the tectonic fabric of the NMP. Relating the morphology of the present river to geology, however, is difficult as it ignores the inheritance history, or the effect of recent events such as the passage of a knickpoint through the reach or the breaching of a resistant unit.

On a smaller scale, most of the knickpoints and changes of gradient, as well as sites of turbulence in the river are located at tributary input points. The tributary inputs in the study area differ from those of conventional river confluences, as described by Best (1986), in that negligible amounts of water, but large amounts of coarse sediment are contributed by these local tributaries. They influence the Orange River by adding large amounts of coarse material which causes an increase in the gradient downstream of the input (Mackin 1948). More importantly, the tributaries inhibit the erosion of bedrock at the site of the tributary confluence by covering up the bedrock. The river uses a great deal of erosional energy in removing these clasts, which are replaced during subsequent tributary flooding events. The bedrock in these areas remains covered until the tributary stops delivering coarse sediment, through a flattening of its profile (Fig.3.41). A similar association between rapids and tributary input bars has been recognised in the canyon rivers of the western United States, where most of the sites of turbulence (rapids) in these river occurs at tributary confluences, however, their applicability to the role that these processes play in limiting the rate of landscape evolution have not been considered.

Depth data tend to mirror lithological strength with the hardest lithologies having the deepest average depths and highest bedrock roughness, consistent with observations made in the ancient deposits. Deep holes are scoured into the bedrock whenever the channel is confined, through either lithological control or downstream of tributary input points. In the Richterveld reach, the profile is not graded and the channel is incising the bedrock and inspection of an inner channel in the Sambok area indicates very recent incision. This incision, as well as the general continued incision of the river such a long time after the initial uplift event can be conveniently explained by active tectonics affecting small blocks. Andreoli *et al.* (1996) have documented the historical seismic events which have occurred in southern Africa, and although stable tectonically, there has been a surprisingly large amount of seismic activity. Further speculation on the effects of recent tectonics is derived from reports of "earth tremors" in the Rosh Pinah area, near Sendelingsdrif (Fig.3.1) (Saaiman, B, Pers.Comm., 2004). However there is an absence of any field evidence, such as the existence of fault scarps or faults that have been active recently, thus active tectonics cannot be considered seriously as an explanation for the current profiles of the Orange River and its system of tributaries at this time.

The aridity of the landscape has probably contributed to the present geomorphology, by preserving the relief. Rates of weathering of the granitic and metavolcanic lithologies would be significantly higher in a wetter climate, and rates of abrasion in the trunk stream would be higher than present. However, rates of sediment delivery to the trunk stream is likely to increase initially, and then decrease in wetter climates as vegetation becomes established (Schumm 1993). Due to the current aridity, it is difficult to quantify the timing involved in the tributary response to incision in the Orange River. The knickpoints present on the coastal plain reach may not be representing response to recent incision but fossilisation of the tributary network in the arid climate that has prevailed since at least the Plio-Pleistocene, and perhaps longer (Ward *et al.* 1983).

3.6 Synopsis

- In the study area, the Orange River displays reaches that are clearly superimposed, as well as reaches that have been modified by basement structure. The tributary system consists of a combination of the pre-existing drainage, prior to incision, and those tributaries that developed in response to the incision of the Orange River, and cut their way back into the landscape. It is not possible to differentiate between the two types of tributaries, apart from the Fish River which contains superimposed meanders in its lowermost reaches and is clearly inherited.

- The Orange River valley responds to changes in the surrounding geology, but generally widens downstream. Likewise, tributary drainage area increases downstream as well as in softer lithologies.

- Analysis of tributary gradient per equivalent drainage area reveals steeper gradients in the upstream reaches for small drainage basins whilst large drainage basins have steeper gradients in the more downstream reaches.

- The coastal plain tributaries display the highest proportion of knickpoints, indicating response to Orange River incision generated from the mouth. However, when this incision occurred is difficult to quantify, as the arid climate is suspected to have preserved the landscape.

- Orange River gradient is the highest in reaches underlain by resistant metavolcanic or granitic rocks, and in reaches flowing across bedrock strike. Low gradients in the highly resistant, but highly fractured Rosyntjieberg quartzites indicate plucking as the dominant erosional mechanism.

- Most sites of turbulence or knickpoints are located at tributary input points, where tributaries supply large volumes of coarse gravel and impede bedrock incision in the Orange River channel.

- Deep scours result when the channel is confined, either in a narrow inner channel, on the outside of a meander bend or downstream of a tributary delta.

- The role of neotectonics being active in some reaches is suspected, but cannot be proved.

4 Drainage Development

4.1 Introduction

The history of drainage development is partly gleaned through straths and terraces on the continent and the sedimentary pile in the offshore. There are no Cretaceous fluvial sediments preserved onshore in the study area, despite the presence of a well preserved Cretaceous delta offshore (see Section 2.4.2). The only possible onshore signature of this Cretaceous river lies in the geometry of the meander loops which have been branded into the Precambrian rocks during a 500-1000 m phase of incision (Ward & Bluck 1997).

The oldest sediments preserved onshore, within the study area, that are linked to the Vaal/Orange system are found some 140-180 km north of Oranjemund, where marine deposits yield a Middle Eocene age of c.42 Ma (Kaiser 1926; Stocken 1978; Siesser & Salmon 1979; Corbett 1989). These deposits contain clast of agates, yellow chalcedonies, jasper and some BIF, all of which have a Kaapvaal provenance, and are prominent in Vaal and Orange River gravels (Ward & Bluck 1997). Near the mouth of the Orange River, trenches reveal fluvial gravels with a similar siliceous clast assemblage to the Eocene deposits, and are accepted as Eocene in age (Ward *et al.* 2002). Unfortunately, within the study area there is no upstream equivalent of this Eocene type gravel preserved to obtain a longitudinal profile of the Orange River in Eocene times. It would have then been possible to calculate long term rates of bedrock incision since Eocene times.

A series of terraces ranging in age from pre-Early-Middle Miocene to Recent, generally decreasing in both age and height, flank the modern river. Their presence indicates that base level lowering has occurred, through either tectonic or eustatic adjustments and similar deposits have been used in other studies to unravel geological histories or quantify rates of incision (Merritts & Vincent 1989; Pazzaglia & Gardner 1993; Merritts *et al.* 1994; Pazzaglia & Gardner 1994; Burbank *et al.* 1996; Repka *et al.* 1997; Lave & Avouac 2001; Pazzaglia & Brandon 2001).

The aim of this section is to characterise, classify and interpret the sequence of terraces using bedrock strath levels, longitudinal profiles and the overall geometry of the sedimentary packages.

4.2 Previous Work

De Villiers and Sohnge (1959), working primarily on the south bank of the Orange River, recognised four groups of terraces, based on the relative height of the terrace surface above modern river level. These are situated at 115 m, between 36.5-76 m, at 30.5 m and between 9-21 m above modern river level. van Wyk & Pienaar (1986), also working on the south bank, recognised five different terrace levels in some localities, although their classification scheme included only four categories. Their classification scheme which was also based on terrace surface height, included an Upper, Intermediate I, Intermediate II and Lower Terrace. These authors specifically avoided making use of the height of the bedrock surface underlying the terraces as a criterion for classifying the terraces, due to its irregular nature (van Wyk & Pienaar 1986). Fowler (1976; 1982) divided the terraces between Sendelingsdrif and the mouth into two broad groups: an older, higher-lying series to which the course of the modern river is markedly discordant, called the Proto-Orange River terraces, and younger, lower-lying Meso-Orange River terraces, with a palaeo-course which is concordant with the modern Orange River, and this classification has been used and expanded on here. The Proto terraces generally correspond with the Upper terrace of van Wyk & Pienaar (1986), although, in places, their Intermediate I terrace is included in the Proto terraces.

4.3 Terrace Suites

4.3.1 Classification

In classifying terraces, more reliance was placed on the height of the bedrock surface underlying each terrace and the material present in the alluvial fill, than the height of the terrace surface itself. The terrace surface height is usually a later erosional feature and may be unrelated to the age of the underlying sediments (although terraces at higher elevation are **generally** composed of older sediment). The old system of classification based on terrace surface height misidentified a number of terraces. Downcutting rivers, during a pause in the downcutting cut laterally, carving flat bedrock straths. When incision resumes, these straths are incised and abandoned (Burbank *et al.* 1996). Thus, although the bedrock surface is variable in nature, fairly flat bedrock straths could be identified at each terrace studied in detail.

The Proto-Orange deposits comprise the Arriesdrift Gravel Formation SACS (1980) and have been assigned an Early-Middle Miocene age, some 17.5-19 Ma, on the basis of macrofauna found at two localities, Auchas Major and Arrisdrif (Corvinus 1978; Corvinus & Hendey 1978; Hendey 1978; Pickford 1987; Pickford *et al.* 1996b;a; Morales *et al.* 1998; Pickford & Senut 2002). The Meso-Orange deposits have not been adequately dated but are assigned a Plio-Pleistocene date (5-2 Ma) from correlation with the littoral deposits of the so-called 30 m package described by Pether (1986). The distribution of Proto and Meso suites of terraces is seen in figure 4.1, and an oblique aerial view of adjacent Proto and Meso terraces is seen on figure 4.2.

Recent work by Namdeb and Transhex geologists has identified a range of deposits; two pre-dating the Proto, one Intermediate between the Proto and Meso, and three discreet deposits within the Meso. Pre-Proto and Proto-Orange River deposits together constitute the Proto Suite of terraces, and Intermediate and all sub-divisions within the Meso-Orange deposits constitute the Meso Suite of terraces (Jacob *et al.* 2001). An idealised cross section showing these subdivisions and their field relationships is illustrated in figure 4.3. Very few localities host the entire suite of terraces.

4.3.2 River Courses

The Proto suite of terraces are those deposits which are discordant with the course of the present Orange River (Fowler 1976), and include the pre-Proto deposits. The distribution of Proto suite terraces between Reuning and Arrisdrif, as well as the inferred Proto and Meso river courses for this coastal plain reach is shown in figure 4.4. The striking feature of the Proto Orange River is its highly sinuous course, which meanders across the present valley, connecting cut-off deposits at the apex of the meanders. Examples of this are found at Reuning-Dreigat-Mehlberg in the Sendelingsdrif area, Bloeddrif North-Auchas Annexe-Bloeddrif South-Auchas Major, where three cut-off loops are fully preserved and Kheis-Auchas Lower-Koeskop (see Fig.4.4). The present valley in the Sendelingsdrif and Auchas area is lithologically/structurally controlled, running roughly north/south parallel to both the foliation in the bedrock and to the resistant ridges of dolomite. The highly sinuous course that the original, Proto-Orange River followed cross-cuts all of these features and breaches a thick (ca. 40 m) unit of dolomite several times in the Bloeddrif/Auchas Major area. This sinuous palaeocourse is perhaps the best evidence for superimposition of the early Orange River.

In the Meso course the channel has straightened to the "easier" lithologically controlled course, although significantly, the reach between Sendelingsdrif and Daberas has remained sinuous from Proto-Orange times until present. This particular area is underlain by more resistant greywacke, and possesses the highest topographic relief of the coastal plain, thus the river was not able to escape from its earlier, sinuous, course, remaining "locked into" its bedrock course. Sinuosity is defined as the ratio of river length along the thalweg to the straight line distance between end points (Leopold & Wolman 1957; Schumm 1963). Sinuosities calculated for the Proto, Meso and Modern Orange River, allowing for changes in the valley direction, are 1.92, 1.48 and 1.61 respectively, confirming the very serpentine nature of the Proto-Orange course. The modern river has increased in sinuosity, after the straightening of the Meso course. Much of this increase is due to the river winding its way around the Meso Orange deposits. Both the Proto and modern Orange would be classified as meandering channels [sinuosity >1.5 (Leopold *et al.* 1964)] and the Meso-Orange would be on the borderline between meandering and straight, although the position of the Meso channel is uncertain and is likely to have a sinuosity between 1.48 and 1.61.

4.4 Terrace Types

"A flood plain is defined as a strip of relatively smooth land bordering a stream and overflowed at times of high water", whilst "a terrace is an abandoned flood plain" (Leopold *et al.* 1964). However, the definition "river terraces are landforms that were at one time constructed and maintained as the active floor of a river but are now abandoned" (Merritts *et al.* 1994) is more applicable to the terraces preserved in the study area, where up to 6km of "active river floor", with attendant channel and bar deposits have been abandoned and preserved at various localities. In a downcutting environment, such as the post-Cretaceous Orange River system, sections of older river can only be preserved if the active channel shifts laterally or changes position before downcutting resumes. Thus, looking at the distribution of terraces, it becomes apparent that they are preferentially preserved in Reaches 3 and 7, where the wider valley has allowed some lateral movement of the channel. The increase in terrace preservation along reaches with a wider valley is commonly observed in other studies (Wisniewski & Pazzaglia 2002; Starkel 2003). In contrast, in the more confined reaches, the river has incised vertically, obliterating any record of its earlier history. As such, there are 4 different mechanisms of preservation (abandonment) present in the area:

Type 1 - Terraces preserved due to straightening of the channel on downcutting. These are usually cut-off meander loops (Fig.4.5a).
Type 2 - Terraces preserved due to lateral movement of the channel subsequent to downcutting – resulting in unpaired terraces. These are usually found in meandering reaches (Fig.4.5b).
Type 3 - Terraces preserved downstream of tributary input points. This style is prevalent in the more confined, upstream reaches (Fig.4.5c).

Type 4 - Terraces preserved astride the modern channel. i.e. dissected on downcutting – resulting in paired terraces. This type of terrace is limited to the younger Meso terraces (Fig.4.5d).

The percentage of the channel length preserved as terraces in the study reach, as well as the representation of each terrace type and in each of the Reaches 1 to 7, was calculated, and only 18% of the original Proto course, as opposed to 49% of the Meso course has been preserved (Fig.4.6). The value (18%) obtained for the Proto deposits is a minimum value as there may be additional Proto deposits present below the Meso deposits. No Proto terraces are present in the more confined Reaches 1, 2, 4 and 5. Small remnants of Proto terraces are found in Reach 3 (8%) and 6 (16%) but the best preserved terraces are found on the coastal plain in Reach 7 (35%). No Meso terraces are present in Reach 1, and few of these terraces occur in Reaches 2 (17%) and 4 (13%). As with the Proto terraces, the wide valley of the Nabas Basin (Geomorphic Reach 3) has preserved more deposits than the adjacent confined reaches (49%), and Reaches 5, 6 and 7 have preserved 39, 69 and 74% respectively, increasing as the valley widens downstream. In terms of preservation style, the Proto terraces display an overall combination of Type 1 (58%) and 2 (39%), with minor Type 3 (3%), whereas the Meso terraces display a combination of Type 2 (59%) and 4 (36%), with minor Type 3 (5%). However, when analysed per reach, it becomes apparent that terrace preservation style changes down the valley. Large cut-off meander loops of the Proto only occurs on the coastal plain in Reach 7, whereas Type 3 terraces are only found in the more upstream reaches for both Proto and Meso and the dissected Type 4 terraces are better preserved in the downstream reaches in the Meso terraces.

4.5 Terrace Profiles

4.5.1 Introduction

A longitudinal correlation of bedrock straths underlying each terrace deposit was done for both the Proto and Meso suite of terraces. To avoid the pitfalls of biased correlation, where one assumes the river is maintaining a nearly constant profile shape through time (Merritts *et al.* 1994), the use of relative height above the modern channel was avoided, with absolute heights in mamsl used instead. Accurate mapping of all terraces on the north bank and some south bank deposits were done using a differential global positioning system (DGPS). The terrace surface height (in mamsl), together with the bedrock strath height and deepest point of bedrock scour was measured in each case. For the deposits downstream of Dreigat, use was made of Namdeb's extensive drilling, prospecting and mining dataset, which was modelled using Vulcan software. For the upstream deposits in the Augrabies Falls area, heights were measured with a hand-held *Silva Windwatch* device that measures height based on barometric pressure. Calibrations to local trigonometric beacons were done to minimise error, but the accuracy of these readings is limited to ca. 1 m. For deposits on the south bank which were not measured directly, old reports were consulted (Stocken 1971a;b; Pienaar 1977; Steyn 1982), and information was kindly supplied by the geologists of Transhex and Alexkor. The data, which can be located in Appendix B, was projected onto the modern profile (Fig.4.7).

As with variable amounts and rates of downcutting along the course of a river, the cutting of straths and the abandonment of the river course to leave terraces must differ in timing along its course. Thus, correlating terraces of similar age, without precise age control is difficult as there may be a continuum of ages preserved for each suite of terraces. In the Aussenkjer/Grasdrif area, there is an extensive record of deposits that have been preserved (Fig.4.7). In this area the river flows on relatively soft Karoo shales and sandstones, upstream of a narrow gorge composed of resistant Nama Group and NMP rocks. The rate of downcutting in the Nabas Basin is controlled by the more resistant rocks in the downstream reach (Miller 1970; Grams & Schmidt 1999; Tooth *et al.* 2002). The softer bedrock has also allowed the river to migrate laterally in this reach, thus, after periods of downcutting, new terrace deposits are more likely to be preserved in this broad valley than in a narrow gorge. Consequently this area has provided the most complete record of terraces in the study region. Apart from this area, the remainder of the data does fall into broad categories (Fig.4.7). These categories will be addressed separately.

4.5.2 Proto Suite

There are two terraces in the study area, the bedrock to them being substantially higher (40-45m) than the surrounding deposits. These two deposits, although undated, are interpreted to be the oldest surviving terrace remnants in the study area and have been termed the "Pre-Proto 1" deposits. These are small remnants that were preserved adjacent to the younger Proto terraces (Fig.4.2b). The upstream deposit at Sendelingsdrif (locally named Snake after the trigonometric beacon situated on its surface) is 3.5 m thick, whereas the downstream deposit at Koeskop (locally named Skilpadsand) is 8 m thick. The longitudinal profile of these deposits is given in figure 4.8.

Shaded contour maps of the bedrock surface elevations of several Proto-Orange deposits obtained from drilling and mining operations are presented in figure 4.9a-e. The longitudinal profile for the Proto-Orange River straths (Fig.4.8) is fairly smooth, assuming that these straths were cut at the same time. In the Grasdrif/Aussenkjer area, three possible strath levels could be used to construct the longitudinal profile. In this case, the strath chosen is the one underlying the largest and thickest terrace deposit in the area. In some terrace deposits, remnants of higher strath levels are present, and at Sendelingsdrif, two older strath levels are present (Fig.4.9a). As with the modern Orange River, the profile of the Proto River contains flat reaches separated by steeper reaches, with small "knickpoints" situated at Aace, Obib and Auchas Lower/Koeskop. It is uncertain whether these are actual knickpoints in the profile or if this profile could have been produced by abandonment of the different reaches at different times. The presence of higher strath levels still evident at Sendelingsdrif (Fig.4.9a) and Arrisdrif (Fig.4.9e), immediately downstream of the Aace and Auchas Lower knickpoints, respectively, may be remnants of a more extensive strath which was abandoned after migration of the knickpoints upstream. Significantly, the highest grade diamond deposit in the area, Reuning Mine is situated in the steep reach downstream of the Aace knickpoint. The Proto profile has a relatively steep gradient through the Richtersveld section (0.00087 m/m) and a shallower gradient across the coastal plain (0.00038 m/m). The modern river, has gradients of 0.00076 and 0.00022 m/m for the equivalent reaches, in parallel with the Proto profile. However, overall, the gradient of the Proto profile is steeper than the modern profile (slope is 0.000688 vs 0.000585 m/m).

The most spectacular features of the Proto profiles are the deep incisions that have occurred at numerous sites (Fig.4.8). These scours are generally 10-30 m below average strath level but can reach depths of 40-50 m below strath level. One of these scours reaches mean sea level at a site 67 km upstream of the present river mouth. As well as discrete scours, linear inner channels are also present at two localities. Within entrenched systems, peak flood discharge is accommodated through increased water depth over the channel and deepening of the channel by greatly increased scour of the bed (Leopold 1969). These scours are similar to those described from active, confined reaches of modern rivers such as the Colorado and Green Rivers, where many scours reaching depths of up to 35 m are encountered (Leopold 1969; Dolan *et al.* 1978; Grams & Schmidt 1999), and the Potomac and Susquehanna Rivers in the eastern United States of America, where scours between 30 m and 35 m deep occur in the steep reaches traversing the Piedmont crystalline basement (Reed 1981; Pazzaglia & Gardner 1993). The most spectacular example, is described by Bretz (1924) from the Dalles area of the Columbia River, where scours of up to 50m below low water depth are reported, reaching up to 35m below sea level, a distance of 217 km from the coast. These scours are described as "mid channel rock basins without waterfall or eddy" and are ascribed to erosion of the jointed basalt by plucking. Shepard and Schumm (1974) working in a laboratory flume, produced similar smaller-scale-scours and inner channels, the "bedrock" of which was scoured to below base level and predicted that similar features would be found beneath the alluvium of natural rivers. Fortunately, diamond prospecting and mining operations have exposed several kilometres of palaeo-river bed in the study area. Figures 4.10 & 4.11 display the modelled and actual bedrock surface of two Proto Orange River deposits, showing features of the bedrock such as bedrock straths and scours. Scours are associated with meander bends, changes in lithology, structural features such as faults and intersecting joint sets and tributary input points.

Deposits filling the scours consist of thin (<3 m) units of well-packed gravel and minor amounts of sand on the bedrock surface, infilling the irregularities. These gravels and sands often have a yellow/brown colouration from a limonite coating and are older than the overlying gravels, which truncate them (Fig.4.12). As a consequence, the scouring is referred to as "Pre-Proto" scouring and the degradational gravel filling the scours is referred to as Pre-Proto 2 gravel. This is distinct from the higher lying Pre-Proto 1 gravel remnants and must represent a younger pre-Proto unit.

The scours, pre-Proto deposits and Proto bedrock straths are buried by gravel and sand deposits, with thicknesses varying from locality to locality but generally increasing in thickness downstream in a wedge-shape (Fig.4.13). Terraces situated furthest from the modern river (in cut-off meanders) or that have been buried by sheetwash derived from local tributaries tend to be the thickest, for a given position along the profile. A date of 17.5-19 Ma was derived from the lower portion of this aggradational fill (Pickford *et al.* 1996a), and is referred to formally as the Arries Drift Gravel Formation (SACS, 1980), although this date is unrelated to the earlier (Pre-Proto) cutting of the bedrock strath. The gravel in these deposits tends to fine downstream through the study reach, and upwards at each locality. The wedge of sediment thickens from 6 m at the upstream end, through 30 m in the middle reaches to approximately 100 m at the downstream end, with gravel thickness declining in the distal portion (Figs.4.13 & 4.14).

4.5.3 Meso Suite

The Meso suite contains all the terrace deposits younger than the Proto suite and includes the Intermediate Terrace, as well as two distinct lower level terraces.

Although frequently referred to in old reports (Fowler 1976; Pienaar 1977), true Intermediate terraces could only be identified at five sites (Fig.4.15). Most of the so-called intermediate terraces referred to in these reports are actually underlain by Proto-level straths and contain gravel characteristic of the Proto suite of terraces, thus illustrating the ineffectiveness of the classification scheme based on terrace surface height alone. Significantly, the Intermediate terraces that were identified in this study, only occur in a very restricted zone between Swartpoort and Jakkalsberg, a distance of approximately 30 km which represents the transition between geomorphic Reaches 6 and 7, i.e. end of the incised reach/start of the coastal plain. Constructing a longitudinal profile using these four deposits shows that the Intermediate terrace at Lorelei West does not appear to match the other Intermediate terraces in terms of bedrock strath height, and it must represent a terrace that was abandoned and preserved at a later time to the other Intermediate terraces. Once again, correlating terraces with the "continuum" of terraces in the Aussenkjer/Grasdrif area is uncertain, as two options can be taken (Fig.4.15). Gravel deposits range from 7m thick at the upstream end to 15m thick at the downstream end and consist of coarse gravel with minor sand interbeds.

Meso terraces are generally underlain by two bedrock strath levels, and are defined as the Upper and Lower Meso deposits. In the Xarries North area, three distinct Meso bedrock levels can be differentiated on the inner bend of a meander, younging towards the outer bend and modern channel (Kruger, Pers. Comm., 2002). However, this is an exceptional case and only one or two bedrock levels are usually present. When two bedrock levels are present, the terraces which they underlie either forms two distinct terraces, or they share a common sloping top surface. When the two levels lie adjacent to each other, the Lower Meso terrace is more often found alongside the modern river. In the longitudinal profile of the Upper Meso terrace (Fig.4.16), strath levels form a much smoother profile than for the Proto strath. The gradient of the Upper Meso profile is steep through the Richtersveld reach, flattening between Reuning and Auchas, and then steepens again to the mouth. Overall, the gradient of the Upper Meso bedrock profile from Noordoewer to Oranjemund is 0.000599 m/m, which is slightly steeper than the modern river (0.000585 m/m). As with the pre-Proto profile, the Meso profile displays numerous large scour features, reaching much greater depths than the modern river. The most spectacular scour, reaching sea level (35-40 m below strath level), is situated 106 km upstream of the present mouth. Although these scours could be re-occupying scours cut in pre-Proto times, a number of them (e.g. At Sendelingsdrif and Bloeddrif South) are situated off the course of the pre-Proto river and must be attributed to a phase of deep incision during Meso times. In the present mouth area, drilling of the Meso terraces has intersected sections of the valley, up to –71 mbmsl. Once again, it cannot be ascertained when these scours were cut. The Upper Meso package is relatively thin and ranges from 6-23 m (upstream - downstream areas) above strath level, and up to approximately 85 m in the overdeepened areas (Fig.4.16). In contrast to the Proto depositional wedge, the Meso deposits retain their gravel in the upper units and distal portions of the deposit (Fig.4.17). Unfortunately, very little is known about the scour deposits underlying these terraces as none have been excavated during mining operations. All the information gained from them thus far is from drill logs.

The longitudinal profile of the Lower Meso bedrock straths (Figure 4.18) are the smoothest of all the profiles, and almost defines a straight line from the Fish River to the mouth, diverging with the modern profile in the Aace to Obib reach, where the modern profile has a concave up shape. The Lower Meso terraces are generally thin, ranging from 8-20 m above strath level.

4.5.4 Correlation with deposits flanking Augrabies Falls

Gravel terraces appear on either side of the Augrabies Falls; ca. 12 km upstream from which a prominent hillock (Renosterkop) has two distinct gravel deposits within large (10-30m diameter) potholes. The higher-lying gravels, ca. 6 m thick are situated about 42 m above modern river level. Two outcrops of lower lying terraces are found in this area, separated by ca. 2 km. They are between 2-7 m thick and lie ca. 22 m above the modern river level. Approximately 70 km downstream of Augrabies Falls, three small outcrops of gravel are found - filling shallow potholes, 2-3 m deep at about 100 m above modern river level. These are situated on the farm Daberas and hence are named the Daberas potholes. No lower terraces are present at this locality. The clast assemblage of the Daberas and Renosterkop gravels are very similar (Section 6). Based solely on the level of the bedrock above the modern river, both these deposits would be correlated with the Proto deposits in the study area, although such long distance correlations are dubious, without proper age control (for further discussion see Section 6). The lower lying terrace at Renosterkop has been correlated with the Meso deposits of the study area (Ward *et al.* 2002), by virtue of the fact that it is at a lower elevation. The profiles from Renosterkop to the mouth (Fig.4.19) show a substantial steepening in the Proto profile between Renosterkop and Daberas indicating that there was some form of major steepening or knickpoint in the Augrabies Falls area during these times. The Daberas potholes are situated far higher above Modern River level than those at Renosterkop so, if they are of similar age, then substantial incision has occurred downstream of Augrabies in the modern channel in post-

Proto times. Alternatively, the Daberas potholes could be older than the Renosterkop gravels, correlating more with the pre-Proto gravels in the study area. Stolzenfels kimberlite is located approximately 35 km downstream of the Daberas potholes, and 2.5 km north of the river (Fig.4.19). The early Orange River must have flowed over the pipe depression as typical Orange River gravel deposits are preserved there. These gravels are situated approximately 135 m above modern river level, the highest level of any gravel relative to the modern river in the Lower Orange region, and may well represent the oldest surviving remnant of inland river gravels.

4.6 Sequence of Events

The sequence of events based on the evidence presented in the section on Figure 4.3 can be interpreted in a number of ways, summarised in Figs.4.20 & 4.21:

4.6.1 Progressive Incision and Aggradation

In response to an uplift event at the end Cretaceous/Early Tertiary, recorded in the offshore sediments (Aizawa *et al.* 2000), the Orange River has incised between 500 and 1000 m into the landscape. In the initial stages, it is predicted that the river would have been flowing across sandstones and shales of the Karoo Supergroup as the Late Cretaceous kimberlites in Bushmanland harbour xenoliths of Lower Karoo sediments (Ward pers. Comm., 2000). The course of this early Orange River was then superimposed onto the Proterozoic rocks in the study area from this cover, although in some places the course has been modified by basement structures (Fig.4.20A&B). The only record of this initial incision is recorded by a small patch of Orange-Vaal River gravels preserved in the Stolzenfels kimberlite some 60 km below Augrabies Falls. However, for the most part, there is no record of this initial incision and either the river incised vertically, leaving behind no record of straths or terrace gravels, or these have subsequently been removed. The initial incision event is thought to have occurred rapidly. Data from modern rivers responding to uplift indicate astounding rates of incision (0.2 to 12 km/Ma), easily keeping pace with uplift (Burbank *et al.* 1996; Whipple 2004). This early incision is probably analogous to the cutting of the Grand Canyon region by the Colorado River which incised the uplifted Colorado plateau at 6 Ma very rapidly at a rate of between 470-800 m/Ma (Paton *et al.* 1991; Hamblin 1994).

The Pre-Proto 1 deposits at Snake and Skilpadsand (Fig.4.20C), located approximately 80 m above the modern river, are the oldest deposits in the study area, and are loosely constrained to a post-Eocene but pre Mid Oligocene age, based on the level of incision and the clast assemblage (Section 6). At this stage, the Orange was deeply incised into the Richtersveld landscape, and the initial 84-92% of the incision had taken place without leaving any sedimentary record onshore. This Pre-Proto 1 strath and veneer of gravel was preserved as the river began to widen its valley.

The next event (Fig.4.20D) to have taken place was the cutting of the Proto strath (in Pre-Proto 2 times). In rivers experiencing long term incision, straths are apparently cut during a pause or slowing down of the uplift/incision, enabling the river to cut laterally and bevel a platform (Burbank *et al.* 1996). Alternatively, in areas within the influence of the coast, straths are cut during sea level transgressions, nulling or reducing the effect of uplift during the period of transgression (Merritts *et al.* 1994; Pazzaglia & Gardner 1994). The pre-Proto scours, incising the Proto strath, appear to represent a phase of incomplete incision, unable to remove the material present between scours. The scours could also represent local overdeepening due to channel constriction in a confined, bedrock valley, like those in the modern Orange River (as well as other bedrock rivers mentioned in Section 3.3.), and thus represent a specific style of river. This implies that an incising bedrock river such as the Orange never cuts a perfect strath. Deep scours and inner channels in the Richtersveld reach of the modern river are present on the coastal plain reach of the Proto-Orange River, although it is unknown to what extent the bedrock in the modern river is covered by sediment. In addition to this, tributaries in Reach 7, currently carrying sand, were delivering coarse gravel to the Proto-Orange River. This indicates that the early valley was more confined than the modern river and that the tributaries were steeper, flowing on a higher relief coastal plain than present.

The timing of this phase of deep incision is thought to be the widespread mid-Oligocene regression (Fig.4.22) reported by many workers, when uplift-induced incision would have acted together with an eustatically-induced sea level low. The magnitude of this regression differs widely in the literature, and is estimated as 400 m (Vail *et al.* 1977), 130 m (Haq *et al.* 1987) and 30-90 m (Miller *et al.* 1991) to non existent (Kominz *et al.* 1998). These estimates differ so widely because eustatic effects on the stratigraphic record are complexly intertwined with other processes such as basin subsidence and changes in sediment supply (Hallam 1992; Miller *et al.* 1998). In southern Africa, a major regression spans all or most of Oligocene time (Siesser & Dingle 1981), based on a widespread Oligocene unconformity on the continental shelf in seismic records (Dingle 1971). The incised valleys meandering across the coastal plain mapped by Hoyt *et al.*(1969) may represent the expression of this incision event on the continental margin. Furthermore, the well-dated Early Miocene (ca. 20 Ma) fluvial to lagoonal sediments of the Elizabeth Bay Formation near Lüderitz (ca 250 km north of the Oranjemund) indicate aggradation within valleys cut during an older, probably Oligocene sea level low stand (SACS 1980; Pickford & Senut 2002).

Incision in the pre-Proto channel ceased and both the undulating bedrock profile and the Proto straths were covered by gravel and sand deposits, up to 55m above strath level (Fig.4.20E). Aggradation occurs when a river is no longer able to transport the sediment supplied to it. The causes of this aggradational event could be one or more of the following:

1. Sea level rise causing reduced depositional gradients and backfilling in the valley.
2. Changes in climate with a period of aridification causing more sediment to be released by the landscape (with less vegetation), followed by wet phases (Cotton 1945; Schumm 1993; Inman & Jenkins 1999), which would flush the stored sediment into the river. Also, aridification would decrease the trunk stream's ability to transport the sediment supplied.
3. Natural expansion of the tributary network, in response to trunk stream incision, eventually supplying too much sediment for the trunk stream to cope with resulting in aggradation – the great thickness of coarse gravel preserved in the Proto terraces support this.

The presence of polychaete serpulid worm tubes (cf. *Mercierella*), indicative of brackish water conditions (usually present in estuaries), in the upper deposits at Arrisdrif (Corvinus 1978; Corvinus & Hendey 1978), currently ca. 34 km from the coast imply that the coastline was much closer to Arrisdrif. These data support the case for sea level rise, although estuaries can extend up rivers for a considerable distance, and it is difficult to estimate the actual height of sea level based on these worm tubes (Dingle *et al.* 1983).

Studies on small streams indicate that base level rise causes a wedge of sediment deposition to migrate upstream for a short distance, and when the deposition is complete, the gradient of deposition is less than that of the original channel (Leopold & Bull 1979; Leopold 1992). Thus, the thick units of sand and silt overlying the gravel package in the distal regions of the Proto depositional wedge (Fig.4.14), which extends for between 75-100 km upstream (value depends on which river course is used) support the case for aggradation due to base level rise. Global sea levels indicate a general transgression, following the Oligocene regression, which peaks in the middle Miocene (Vail *et al.* 1977; Haq *et al.* 1987). In southern Africa, evidence for a middle-late Miocene transgression is found on the continental shelf, which continued to a maximum in the early Pliocene

(Siesser & Dingle 1981; Dingle *et al.* 1983)(Fig.4.22), supporting the case for aggradation due to sea level rise.

The abundant fossils (>10000 fossils) preserved at Arrisdrif and Auchas indicate subtropical to savannah conditions (Pickford *et al.* 1996b;a; Pickford & Senut 2002). A similar fossil assemblage of Late Middle Miocene age has been found at Bosluispan in Namaqualand (Senut *et al.* 1996), thus this was the first significant wet (pluvial) period since the drier Oligocene climate (Dingle *et al.* 1983). However, feldspathic grits present in the gravels (which breakdown and are absent in humid climates) as well as semi-arid sedimentary settings in the nearby Grillental (Stocken 1978; Corbett 1989) suggest an arid climate, and that the fossil assemblages are more likely to be derived from the linear oases formed by the Lower palaeo-Orange at that time (Ward *et al.* 1983). Conditions dried out further during the late Miocene when thick aeolianites accumulated in this region under arid to hyper-arid conditions (Ward *et al.* 1983; Ward & Corbett 1990), probably initiated by the commencement of upwelling and the cold-water Benguela current (Siesser 1980). Thus, the Proto aggradation was initially accompanied by a wetter phase, drying out in the later phases indicating that climate changes also played an important role in this aggradational event. It should be noted that sea level change is itself related to climate change (Bridgland *et al.* 2004).

The preservation of thick coarse gravel deposits, especially in the Sendelingsdrif-Auchas reach (Fig.4.14) could also be explained by aggradation due to oversupply of gravel by tributaries, in reaction to the mid-Oligocene incision event. The knickpoints present in the Proto profile could also be explained by the "top down" supply of gravel from upstream (Wisniewski & Pazzaglia 2002). As a slug of gravel moves progressively downstream, it covers up the bedrock, thus inhibiting further bedrock lowering in the covered areas. Bedrock lowering can occur in the downstream areas, until it is also covered by gravel.

The best explanation for the aggradation problem probably lies in a combination of all three factors, and will be explored further using clast assemblages in Sections 5 and 6.

Following the dated, Middle Miocene aggradational phase, the river resumed incision, flowing on its own alluvium, 50-90 m above bedrock level at the downstream end (Fig.4.21F). This occurred either in response to Plio-Pleistocene uplift of the sub-continent (Partridge & Maud 1987), or as a response to the regression event recognised in the global sea level curves in Late Miocene (Fig.4.22), and locally in the offshore in Late Pliocene times (Siesser & Dingle 1981). Due to the relatively unconfined nature of the river and soft footwall, the river incised rapidly, straightening its course on this down-cutting phase. Consequently, the Proto meanders in the areas of relatively soft micaceous schist were cut out and abandoned, whereas in the areas of slightly more resistant rock, such as the metavolcanics at Arrisdrif, and the greywackes in the Daberas/Obib area, the earlier meandering pattern was retained (Fig.4.4).

The earliest record of this new phase of incision is recorded in the Intermediate terraces, the strath levels of which are ca. 13m below the Proto strath levels. The restricted longitudinal occurrence of these Intermediate deposits could either suggest that some special factors were involved in their preservation in this reach, or it indicates local aggradation at that time, when the reaches both upstream and downstream were incising. The longitudinal position of these Intermediate deposits in the transition zone between reaches 6 and 7 is noteworthy in that it is a position that currently favours the deposition of large gravel bars after exiting a relatively confined reach. In this locally aggrading reach, the river was able to shift laterally and when incision resumed, the deposits were left as an Intermediate terrace. This local aggradation could represent a pulse of gravel, in response to the post-Proto incision, that would have re-activated the tributaries supplying coarse gravel. This locally aggrading reach is analogous to the modern river in the Aussenkjer/Grasdrif area where large gravel bars have been deposited downstream of the Rosyntjieberg gorge. This local reach of deposition is flanked by actively downcutting reaches. Note that the Intermediate deposits are not represented in Fig.4.20 due to their very limited areal extent.

Intense downcutting resumed, with a pause, to cut the Upper Meso strath (Fig.4.20F). As with the Pre-Proto 2 scours, the scours underlying the Meso deposits were either cut at the same time as the Upper Meso strath, or following it. This phase of incision was followed by aggradation of the Upper Meso terrace (Fig.4.20G). This aggradational event could also be linked to sea level rise (Siesser & Dingle 1981), or wetter phases during the Pleistocene. Wetter phases are known to have occurred further inland during this time (de Wit *et al.* 2000), and from calcareous travertines deposited in pans and at spring sites in the southern Namib (Pether *et al.* 2000). Another alternative is from increased input of tributary material following an incision event.

The Late Neogene Upper Meso aggradation was followed by downcutting to the Lower Meso strath level (Fig.4.20H), a pause to cut the strath before aggradation of the Lower Meso deposits (Fig.4.20I). These terraces were locally incised by the modern river

(Fig.4.20J) to leave the Type 4 paired Meso terraces. This interpretation of events requires alternations of bedrock erosion and alluviation and also requires the older terraces to remain largely intact during the >20 Ma time it took for the cycle of degradation to reach present river course level. It must also be noted that the above interpretations apply only to the terraces on the coastal plain reach. As mentioned previously, the terraces in the upstream reaches may have been abandoned at entirely different times to those further downstream due to the compartmentalisation of bedrock rivers. This would result in differing ages of terraces at similar heights above river level.

4.6.2 Incision and Large Aggradation Event

An alternative interpretation of the sequence (Fig.4.21A-I) is that there was a long phase of incision, with pauses in the downcutting during which the Proto, Upper Meso and Lower Meso straths were cut culminating in the cutting of the deep pre-Proto scours, some of which reach modern river level and lower (Fig.4.21A-G). A long period of aggradation ensued (>20 Ma) which filled the valley with sediment up to 90 m above the channel. Gradual entrenchment into this alluvial fill followed to preserve the different terraces. Hence the modern meander loops present between Sendelingsdrif and Daberas are occupying the original inherited incised meanders.

Problems with this interpretation are 1. The presence of deep Meso scours underlying the Meso deposits: some of which could be occupying scours cut in pre-Proto times, although, a number of them are situated off the course of the original superimposed pre-Proto river. 2. The general lack of paired terraces, and the overall dominance of unpaired Type 2 terraces situated on meander bends displaying a younging of terraces towards the modern river supports the interpretation in figure 4.20, with a gradual lateral movement of the river with progressing incision and aggradation. This implies that the meander loops in the Sendelingsdrif to Daberas reach are modified versions of the original Cretaceous superimposed loops, and not the original meander loops (cf. Ward & Bluck, 1997).

4.7 Summary

The series of terraces flanking the modern channel in the study area are consistent with a long period of incision, punctuated by periods of aggradation (Fig.4.3). These terraces can be subdivided into 2 suites, where the older Proto Suite of deposits span a long period of uplift induced-incision, aided by a global sea level lowstand which incised the landscape up to a kilometre below the planated African surface. This was followed by a long period of alluviation, induced by a combination of sea level rise, climatic fluctuations or the natural tributary response from the incision event. The younger Meso suite of deposits also represent a period of active incision which straightened the course of the river and scoured deep holes, followed by aggradation and then gradual incision to the present river, punctuated by at least another aggradation event. However, the majority of the overall incision of the Mid to Lower Orange River was complete by mid-late Oligocene (Pre-Proto 2 time) and since that time the river has incised very little.

As observed by previous workers elsewhere (Pazzaglia & Gardner 1993; Merritts *et al.* 1994; Formento-Trigilio & Pazzaglia 1998; Wisniewski & Pazzaglia 2002), the gradient of the straths generally decreases from the old to the modern river (Fig.4.23), and converge downstream. In the Matthole River in northern California, converging profiles have been explained by backfilling of the valley by a wedge of sediment in response to sea level rise, causing the gradient of deposition to decrease. This in turn would have an effect on the adjacent upstream reaches which are cutting bedrock (Merritts *et al.* 1994). In the Lower Susquehanna River on the east coast of the USA, the divergence of terraces upstream is thought to result from a long term isostatic uplift affecting the inland portion, whereas the hinge zone separating this from offshore subsidence is located where the terraces converge (Pazzaglia & Gardner 1993). Formento-Triglio & Pazzaglia (1998) measured downstream converging profiles on the Jemez River terraces in the southern Rocky Mountains, western USA, attributing this to an offset across a major fault. In a nearby area, Wisniewski & Pazzaglia (2002) observed similar strath convergence downstream, attributing it to epeirogenic arching over the Jemez lineament. Unlike some of these examples, downstream convergence of the straths in this study cannot be easily explained by local tectonic effects. The most likely comparison would be from the east coast of the USA where slow, long term regional isostatic uplift is affecting the upstream reaches.

In the study area, since the initial aggressive incision and large aggradational event, the development of the terraces is consistent with a river incising into a slow, isostatically uplifted margin, much like the conditions envisaged on the East coast of the USA (Pazzaglia & Gardner 1993). Straths are cut in times of stability (probably during base level rise when this balanced the isostatic uplift), incising the straths and abandoning terraces in times of base level fall, and aggrading due to a flush of gravel from the local tributaries, either from wetter phases, or from natural tributary incision, following trunk stream incision events.

4.8 Synopsis

- More terraces have been preserved in the downstream and wider sections of the valley, whereas in the confined reaches, terraces have only been preserved downstream of tributary input points.

- Two suites of terraces are distinguished on their river course, bedrock strath height and overall geometry. The Proto terraces are thick accumulations of sediment, built on a high level strath, with a very sinuous palaeo-course. The Meso terraces in contrast are thinner, laterally extensive deposits built on lower strath levels with a significantly straighter palaeo-course.

- The older Proto suite, incorporating the Pre-Proto and Proto deposits represents a long, post-Eocene phase of incision, followed by a long period of aggradation up to 90 m thick.

- The younger Meso suite of deposits, incorporating the Intermediate and Meso deposits represents shorter phases of incision and aggradation.

- In latest Miocene time, Augrabies Falls was in existence near to its present position, or a steep knick-zone was in place.

- Bedrock straths are converging through time, but with the Richtersveld reach between Noordoewer and Sendelingsdrif remaining steep through time.

5 Clast Assemblage – Maximum Clast Size

5.1 Introduction

In a fluvial system, smaller clasts travel furthest, and so have the most diverse provenance, and the potential to provide information on the evolution of the entire drainage basin (Chapter 6). Larger clasts (>90mm), being locally sourced, provide information relevant to the study reach and on processes operating there. The coarsest material present may be related directly to hydraulic mechanisms (e.g. competence and abrasion), and is generally considered to exert the greatest influence on channel roughness (Rice & Church 1998). The largest clasts are also responsible for bedrock cutting, and have the most information pertinent to the evolving tributary network and on the potency of the system to break-down labile rocks and concentrate the tougher lithologies.

The gravel size and clast type in the study reach between Noordoewer and the mouth is modified considerably by the addition of clasts from tributaries. In order to understand the response of clast type to flow along the river together with the effects of tributary input within the study reach, changes in grain size, roundness, shape and composition of the maximum clast size (MCS) population are recorded and analysed. Emphasis is placed on clast assemblages in the modern river system, which is intact and processes responsible for variations can be observed. Comparisons are made with the terrace deposits and the insights gained from the modern system are used to interpret the dataset from the terrace deposits.

Clasts tend to fine in a downstream direction, hence at any one site, those from a remote source have the potential to be finer than those from a more proximal source. This decline in size may be both a function of abrasion and progressive sorting, each being more significant in different circumstances. In this section, the relative significance of progressive sorting and abrasion are assessed.

5.2 Previous Work

The reduction of clast size in gravel bed rivers (Pettijohn 1957) has been attributed to a combination of abrasion during transport (Sneed & Folk 1957; Whiteman 1986; Werritty 1992; Kodama 1994a), abrasion of gravel in place (Schumm & Stephens 1973), *in situ* weathering whilst sediment is stored in bars (Bluck 1964; Bradley 1970; Jones & Humphrey 1997) and selective entrainment and/or enhanced transport of finer sediment sizes, generally called sorting (Brierley & Hickin 1985; Paola & Seal 1995; Hoey & Ferguson 1997; Seal *et al.* 1997; Rice & Church 1998; Gasparini *et al.* 1999; Rice 1999; Gomez *et al.* 2001).

The relative importance of abrasion and sorting processes are difficult to separate, although there is general agreement that sorting processes are more effective than abrasion processes. However abrasion (incorporating splitting, crushing, chipping, grinding and sandblasting) is thought to have more importance in specific settings such as a degrading river (Kodama 1994a). This conclusion is due, in part, to the results obtained from early experiments on abrasion which have been unable to replicate field downstream fining rates (Krumbein 1941; Kuenen 1956; Sunamara *et al.* 1985). However, Lewin and Brewer (2002), in reviewing experimental abrasion studies concluded that field abrasion rates have been seriously underestimated, as the processes involved are only partially represented by the experimental equipment used. In addition, the translation of weight loss abrasion coefficients (from experiments) into those of particle size create problems for the comparison with existing laboratory and field data, thus, the results obtained experimentally are capable of misinterpretation when related to field trends (Lewin & Brewer 2002). The experiments also take no account of the effects of weathering of gravel stored on bars (Bradley 1970; Jones & Humphrey 1997), or the effect of abrasion *in situ*, by a passing bedload and by collisions of clasts "vibrating in place" without downstream movement (Schumm & Stephens 1973). Kodama (1994b) managed to replicate diminution coefficients of Japenese rivers in his particularly vigorous experiments, achieved by free-fall of particles onto water-submerged ones in a rotating drum.

Laboratory flume studies (Seal *et al.* 1997), numerical modelling (Parker 1991a;b; Hoey & Ferguson 1994; Hoey & Ferguson 1997) and field measurements (Smith & Ferguson 1996) indicate that rapid downstream fining can develop through selective sorting alone. Selective sorting at the bar scale results in large grain size changes on individual bars (Bluck 1982;1987), however, relating the mechanisms involved in bar scale to system scale changes in clast size is difficult (Hoey & Bluck 1999).

Disruption in downstream fining patterns, and the proportions of clasts in the different size fractions occurs in response to sediment supply by tributaries and other lateral sources such as tills, fans and terraces (Knighton 1980; Dawson 1988), however, not all tributaries have a significant impact on fining profiles (Rice 1998; Rice & Church 1998). The size distribution of material supplied by tributaries is constrained by lithological characteristics that determine the initial clast size and shape, the climatically influenced weathering rate and style and the hillslope/sediment transport process and length of travel to the trunk stream (Sklar & Dietrich 2001). The rate of sediment supply from tributaries is important in fining studies. The rate of downstream fining is reduced with increases in sediment supply rate (Hoey & Bluck 1999) as the supplied grain size is dispersed over a greater length (Seal *et al.* 1997). In addition, high rates of sediment supply often result in aggradation which in turn affects rates of downstream fining, although there is no simple relationship between fining and aggradation rate (Hoey & Ferguson 1997).

Clast roundness is known to increase with distance travelled, initially proceeding rapidly over short distances, and then slowing down (Wentworth 1922; Sneed & Folk 1957). However, roundness also varies with size, shape and lithology. Roundness generally increases with particle size (Pettijohn 1957), up to a certain size limit, after which a decrease in rounding occurs (Bluck 1969b). Variations in rounding with clast shape also occur, although separating the influence of clast shape and lithology is difficult, since clast shape is often related to subtle changes in lithology. Bluck (1969) found that spheres are generally better rounded than disks for all size fractions, with the disparity increasing at large grain sizes. Roundness is very sensitive to lithological variations. Sneed and Folk (1957) observed that limestone reached its maximum roundness within the first few kilometres, quartz rounded much more slowly but eventually attained a limiting roundness and chert only rounds very slightly downstream, this being in agreement with the results of earlier workers (Krumbein 1941; Plumley 1948; Kuenen 1956).

Numerous studies have investigated clast shape changes with distance. Bluck(1982) demonstrated that the coarse size fractions show a downstream decrease in discs and an increase in spheres, however, shape segregation on individual bars has a greater effect than many kilometres of transport of the bulk gravel. Different lithologies behave differently with increased downstream transport: Vein quartz becomes more rod-like, chert becomes more bladed and limestone is unchanged with distance travelled (Sneed & Folk 1957). Bradley *et al.* (1972) found that quartzite and greywacke became more elongated and foliated rocks more platy downstream, whereas Huddart (1994) found no change in particle shape with distance travelled for all lithologies studied.

Whilst studies of rates of gravel fining in existing active channels are many, few workers have sought to determine the rate of fining in both the active channel and flanking terrace deposits. Toyoshima (1987) found less rapid downstream fining of gravel in Pleistocene terraces in comparison with the modern Osarube River in Japan. He related this difference to downstream progressive degradation which occurred during the formation of a strath terrace in Pleistocene times. In an alluvial fan sequence, Bluck (1964) compared an earlier mudflow to a younger stream deposit, and found that the mudflow had a faster rate of grain size decline, but a consistently larger particle size. Ward (1984) working in the Kuiseb River valley at the northern boundary of the Namib sand sea, found a lower maximum clast size in Miocene age gravel deposits than Pleistocene age deposits, and used this to infer deposition of the younger deposits in a higher energy, incised, canyon-like fluvial system. Spaggiari (1993) studying the palaeo-Vaal River deposits also found the maximum clast size tended to decrease with increasing age. However, in these studies, comparisons of the downstream rate of grain size change through time were not done.

Plumley's (1948) classic study of the Black Hills gravels, although working in terrace gravels, concentrated on the trends within one terrace and thus did not make any comparisons through time.

5.3 Methodology

A sampling campaign was undertaken to measure and collect data from every exposed modern bar between Noordoewer and the mouth. To ensure that no bars were excluded by being submerged, the exercise was done during consistent low flow conditions over a period of three weeks. A canoe was used to access the bars in the turbulent, Richtersveld reach between Modderdrif and Sendelingsdrif, whereas an inflatable boat with a small engine could be used in the less turbulent coastal plain reach between Sendingsdrif and the mouth. Each bar was classified into a bar type, after which all three axes of the largest 20 exposed clasts in a 2x2-m area were measured (Bluck 1987) and lithologically identified. A visual estimate of the roundness was made using the scheme of Powers (1953), where each clast was assigned an estimated roundness value between 0 and 1. Samples from clasts that were difficult to identify in the field were collected for later examination. A typical measuring site and roundness scale is illustrated in figure 5.1. The MCS per bar was calculated using the average of the 20 measurements taken. The dataset can be located in Appendix C.

Maximum clast size is used for analysing the grain size changes, as opposed to size distribution of a bulk sample, because maximum clast size more than likely represents the direct response of the largest grain sizes to the maximum flood events (Bluck 1982). Particles of intermediate and fine grain sizes represent clasts that move at lower flow stages, and are consequently mixtures of different grain populations with different transport histories, and thus cannot provide a meaningful set of data when analysed as a single population (Bluck 1982). The coarsest active material present has been the focus of many previous fining studies (Pettijohn 1957; Dunkerley 1994; Rice & Church 1998). The details of data measurement employed by various authors differ somewhat, ranging from the measurement of the long or intermediate axes of the largest 10-50 clasts that can be measured within either a set period of time (Bradley *et al.* 1972), within a certain distance along the channel (McBride & Picard 1987), within a restricted area (Bluck 1987), at a locality or stratigraphic section (Ward 1984; Matheys 1990; Spaggiari 1993), or a set number of clasts per available lithology (Werritty 1992). In this study, a fixed number of clasts were measured in a restricted area to facilitate unbiased downstream comparisons in average maximum clast size (MCS), roundness and shape per bar and per lithology. In addition to this, the relative proportions of lithologies present in the MCS population could also be compared per bar or per reach.

Due to the fact that substantial local size and shape sorting is evident on bars and within channels (Bluck 1982), care was taken to sample the bars consistently with respect to position. Thus, the upstream area of the bar, the bar head, was always targeted, as this contains the coarsest clasts and is the most stable part of the bar (Bluck 1971; Church & Kellerhals 1978; Bluck 1982;1987). Considerable noise can be introduced to downstream fining data if sampling is not carried out carefully with respect to bar type and position of each sample on each bar (Hoey & Bluck 1999).

Three bar categories were noted: tributary input bars, normal bars and strath bars (Fig.5.2a-c). Tributary input bars or tributary input deltas are the deposits of gravel that are found at the point of intersection (mouth) of many tributaries with the Orange River. They are equivalent to the debris fan deposits of Grams & Schmidt (1999). Normal bars consist of point bars, lateral bars (Bluck 1975) or medial/mid-channel/braid bars (Leopold & Wolman 1957; Bluck 1975). Strath bars are defined as bedrock-attached bars, usually thin gravel accumulations on a bedrock strath surface, some of which are protected from being swept off the strath by bedrock protruberences on the strath surface. The location of the MCS samples within the study reach can be seen on Figure 5.3.

To investigate maximum clast size variations on individual bars, the long axis of 20 clasts were measured within a 2x2m area, and this was repeated every 100m from the upstream head, to the downstream tail of the bar or until the gravel-sand transition. Four bars in the

Sendelingsdrif area and one in the Modderdif area were measured up in this way (see Fig.5.25).

For comparison with the deposits preserved in the terraces, MCS measurements were done at each clast locality used for clast assemblage determinations (see Fig.6.4a). In these cases, the long axes of the 20 longest clasts in a 2x2m vertical section of gravel was measured and their roundness estimated. Due to the fact that the clast was buried in the face, only the long axis was measured. Although not directly comparable to the modern bar surface, where significant surface armouring appears to have occurred in the degradational setting of the modern river, it does provide a minimum value of MCS for the terrace deposits.

5.4 Clast Lithologies

The study reach traverses a large variety of rock types, all of which could potentially contribute clasts (Fig.3.3). Due to the wide variety of lithologies, where possible, an attempt was made to group clast types according to their source geology. However, this was not always possible, as in the case of granite where large variations within each suite prevented separation into the different suites. Figure 5.4 is a summary diagram relating the MCS clast types to the stratigraphic column.

5.4.1 Mafic and Felsic Metavolcanics

The mafic and felsic metavolcanic rocks are sourced from the de Hoop Subgroup, being the oldest unit of the Orange River Group. They consist for the most part of mafic lavas with occasional felsic lavas, the relative proportions being roughly 3:1. All of the lavas are metamorphosed and sheared to some extent (de Villiers & Sohnge 1959; Ritter 1980; SACS 1980).

Mafic Volcanics (Fig.5.5a) are typically green, although some are dark grey to black, and are generally andesitic, and rarely amygdaloidal. Where fresh, the mafic lavas are dense, fine grained rocks in which individual minerals are difficult to distinguish. They typically have equigranular textures but may have hornblende or plagioclase phenocrysts in some areas. Some of the sheared lavas are chlorite-epidote-sericite schists (de Villiers & Sohnge 1959; Ritter 1980), although these sheared varieties breakdown very quickly once in the river.

Felsic volcanics (Fig.5.5b) consist largely of quartz-feldspar porphyry and are pale, dense fine grained rocks, although some darker grey and green varieties are present. Quartz and plagioclase form phenocryst phases, whereas the groundmass contains quartz, K-feldspar and plagioclase. They may have a massive to a strongly anisotropic texture. When sheared, quartz-sericite schists are produced and in some areas, replacement by epidote makes it difficult to distinguish these rocks from the sheared mafic lava. However, clasts that displayed visible quartz in hand lens examination were generally classified as felsic volcanics (de Villiers & Sohnge 1959; Ritter 1980).

5.4.2 Rosyntjieberg Quartzites

Quartzites (Fig.5.5c) are sourced from the Rosyntjieberg Formation, which conformably overlies the de Hoop Subgroup (de Villiers & Sohnge 1959). The formation forms the core of the Rosyntjieberg mountain range, which strikes northwest to southeast (Fig.3.2a). The quartzite is very pure and distinctive being composed mainly of white to pale pink quartz, in which cross-stratification and ripple marks may be observed occasionally. In addition to the quartzite, minor amounts of ferruginous quartzite, chloritic quartzite and schist are found, although very few of these were identified in the gravel deposits. The quartzites have been metamorphosed, and both coarse and fine grained varieties are present. The quartzite is generally well bedded in the dm to m range, however, it can be massive and apparently unlayered (Ritter 1980). The distinctive features of the Rosyntjieberg quartzites are their white colour and their purity, often resembling vein quartz.

5.4.3 Namaqua Metamorphic Province Lithologies (NMP)

The NMP lithologies comprise the Vioolsdrif Suite [1900 +-30 Ma to 1730 +-20 Ma (Reid 1982)], which was originally the "grey gneissic granite" of de Villiers & Sohnge (1959), and is found in the north-eastern and eastern parts of the Richtersveld, and stretches east to beyond Goodhouse (SACS 1980). It is intrusive into the volcano-sedimentary succession of the Orange River Group. In all areas, this suite is composed of spatially associated intrusives varying widely in composition. Each lithology contains a regional foliation, and has a characteristic history involving the initial intrusion of more mafic followed by more felsic lithologies. The Vioolsdrif Suite is composed of mafic and ultramafic rocks, together with diorite, tonalite, granodiorite, granite and leucogranite (Blignaut 1977; Ritter 1980; Visser *et al.* 1984). Due to the large variety of lithologies present in the NMP, the colloquial term "Basement" was used. This encompassed both felsic and mafic intrusives with a metamorphic fabric, most commonly a gneissose one. For the MCS study, diorites and granodiorites without fabric were classified separately to the foliated basement (Fig. 5.5d).

5.4.4 Granite and Syenite

There are a number of post Namaqua granitic intrusions, each displaying a wide variety of compositions and grain sizes. For this study, all felsic intrusive rocks without a fabric were included in the granite category (Fig. 5.5e). The sources of these granites are outlined below.

The Richtersveld Suite [833 +-2 Ma to 771 +-6 Ma (Frimmel *et al.* 2001)], formerly known as the Richtersveld Igneous Complex (de Villiers & Sohnge 1959), consists of a large batholith and two smaller plutons, which are intruded into the Orange River Group and Vioolsdrif Suites in a roughly trending northeast/southwest line (Fig. 3.3). The plutons consist of alaskitic granite, porphyritic microgranite, porphyritic syenite, granular syenite and local veins of rhyolite (de Villiers & Sohnge 1959). Syenite clasts are derived uniquely from the Richtersveld Suite.

The Kuboos-Bremen Suite [ages between 500 and 550 Ma (Allsopp *et al.* 1979; Frimmel 2000)] comprises a chain of plutons making the Kuboos-Bremen "line" which was emplaced in five linear clusters (Tankard *et al.* 1982) (Fig. 3.3). They vary in composition from granitoid to syenitoid and include carbonatite diatremes and sheets in the extreme north, although only rocks of granitic composition were yielded to the river.

In addition to the intrusions mentioned above, unfoliated granites are also derived from the Vioolsdrif Suite, although they constitute a very small minority of the granitic population.

5.4.5 Gariep Supergroup Metasedimentary Rocks

The Gariep Supergroup provides relatively few clasts to the modern river, despite flanking the river across the entire coastal plain from Dreigratdrif to the mouth. This is mainly due to a combination of subdued topography and relatively labile lithologies. All of the Gariep Supergroup lithologies encountered are classified with the Metasedimentary group (Fig.5.5f), which includes mainly schist and greywacke, although impure (dirty) quartzites and feldspathic quartzites, as well as diamictite, are also present in small amounts. Other lithologies from the Gariep Supergroup, not included in the metasedimentary category are dolomitic limestones from the Hilda Subgroup (categorised with the limestones) and metabasalt from the Grootderm Formation (being classified with the mafic volcanic category). Quartzites belonging to the Stinkfontein Group, the basal unit of the Gariep Supergroup could have also contributed to the quartzite population, although the river is not in direct contact with them and no tributaries are actively supplying any of this material to the river at present. However, they may have been a more important contributor of clasts in the past. These quartzites are generally pale cream to blue grey coloured arenites, and are difficult to distinguish from the Rosyntjieberg quartzites in places (von Weh 1993).

5.4.6 Nama Group

The most important lithology to be sourced from the Nama Supergroup and most ubiquitous in the Orange River gravel deposits is Nama quartzite (Fig.5.5g). This quartzite has a large variety of colours, ranging from black, green, grey, green and white in places, grain sizes range from pebble conglomerates to rare fine grained varieties and compositions vary from arkosic to quartz arenite. Most quartzites are sourced from the Kuibis Subgroup which contains two thick quartzite horizons: a basal clastic member and an upper clastic member. There is a general increase in mineralogical maturity (less feldspar) and quartz cement content towards the upper member, as well as a change in colour to white, often with a red oxidised surface (Germs 1972;1974;1983). Nama quartzites are distinguished from the white, pure Rosyntjieberg quartzites by their colour, as well as their relative immaturity. However, the pure white Nama quartzites in the upper clastic member are difficult to distinguish from the Rosyntjieberg quartzites if there are no red oxidised surfaces present in the Nama quartzites.

Thick units of limestone and dolomitic limestone are found in the Nama Group. These also contribute to the limestone group of clasts (Fig.5.5h).

5.4.7 Karoo Supergroup

The Karoo Supergroup, outcropping in the Nabas Basin connecting Noordoewer and Aussenkjer provides shales and sandstones to the river. Due to their fissile nature, the shales are only found in the finer grain sizes, whilst sandstones enter the river as large, flat slabs (Fig 5.5i). The sandstones have a yellow/brown colour and are easily distinguished from the quartzites by their colour and relative incompetence.

Unaltered dolerite, present as the sills constituting the "Tandtjiesberge" in the Nabas basin can also be identified in the gravels. The dolerite is generally medium grained, and feldspar and pyroxene can be distinguished in hand-specimen.

5.4.8 Vein quartz

Vein quartz (Fig.5.5j) is provided from the many quartz veins and pegmatites present in the NMP rocks, as well from quartz veins in the Gariep Supergroup.

5.4.9 Basement quartzites

This category of quartzite, which are fairly rare, are quartzites which have a distinct foliation and are quartz-mica schists (Fig.5.5k). These are highly resistant, but it is not known from which set of rocks they originate, hence the name basement quartzite, as it is likely that they are derived from the NMP or the Stinkfontein Group of the Gariep Supergroup.

5.5 Results – Modern River

5.5.1 Maximum Clast Size and Roundness

Changes of maximum clast size with distance in the study reach are demonstrated by using either one of the clast dimensions (Fig.5.6a & b) or the clast volume (Fig.5.6c). Although there is a general size decline downstream, there is considerable variability in the maximum clast size (MCS) between bars in the reach between Modderdif and Sendelingsdrif, reflecting the disruptive influence of clasts supplied by the tributaries on the clast size decline of the main Orange River. Downstream of Sendelingsdrif, apart from a few disruptions by tributaries, size decline occurs at a fairly constant rate (Fig.5.6a). Displaying the data per bar type (Fig.5.6b & c), illustrates the disruptive influence of the tributaries on the material present on the normal bars in the reach between Modderdrif and Sendelingsdrif. In the reach between Sendelingsdrif and the mouth, where there are less tributaries, the size decline is less variable. The use of clast volume (calculated as the product of the length of a, b and c axes) illustrates the large difference in size of clasts present in the Richtersveld reach compared with those present on the coastal plain (Fig.5.6c). For all parameters in Fig.5.6 the maximum clast size per bar shows a general increase in sorting downstream (Fig.5.6d), in agreement with the results of Bluck (1982). Material present on the tributary input bars ranges widely in size in comparison with the clasts worked onto the normal and strath bars. This large increase in sorting over very small distances illustrates that the river is well capable of sorting out the diversity of grain sizes delivered to it. This is complicated by the fact that river competence changes downstream as its gradient changes. However, it would appear from the sorting data that river competence is ample in all reaches to re-distribute the sediment supplied to it, and the lack of tributary bars in some of the steepest reaches may actually be indicating that the river is too competant for the grain size supplied.

In the study area, the supply of material from tributaries is the principal control on grain size, although the Meso and Modern rivers may have derived some of their gravel from pre-existing terraces flanking the channel. However, this contribution is thought to be relatively minor in comparison to the very active tributaries. The grain size of the material supplied by the tributaries is clearly an important control on the grain size in normal bars. It is therefore important to investigate the possible controls on the grain size supplied by the tributaries. The clasts supplied by tributaries are a function of tributary gradient, drainage basin area/channel length, source lithology and climate. There is no correlation (r^2=0.0499) of clast size with tributary slope (Fig.5.7a) or clast size with drainage basin area (r^2=0.0117) (Fig.5.7b). One would think that there needs to be a balance between a slope steep enough, and enough water to transport the clasts, and, as tributary slope and drainage basin area are inversely related (Fig.3.17), an optimal size of tributary drainage area would be expected to deliver the largest clasts.

In contrast to gradient and drainage basin area, there is much stronger lithological control on the size of clasts found in the tributary bars (Fig.5.7c). Tributary bars dominated by granite, granodiorite or gneiss clasts (NMP basement) tend to have large average clast sizes. All of the tributaries with extremely large average clast sizes in Fig.5.7a and b are dominated by one of these lithologies, which tend to be more massive in outcrop and weather into large blocks, in comparison to the well bedded and fractured sedimentary rocks and well jointed metavolcanic rocks.

In spite of all the disturbances to the system by the input of coarse clasts from the tributaries, four well defined fining profiles can be recognised within the normal and strath bar dataset (Fig.5.8a and b). These reaches which appear to be relatively undisturbed by tributary inputs are referred to as sedimentary links (Rice & Church 1998). The position of these links within the study reach is illustrated in Figure 5.8c. In terms of the tributary reaches defined in Section 3 (Fig.3.12), Link 1 correlates exactly with Reach 1. Link 2 incorporates Reaches 2, 3 and approximately one half of Reach 4. Link 3 incorporates the other half of Reach 4 and Reach 5 whilst Link 4 includes Reaches 6 and 7.

When clasts of a certain size enter a river, if there are enough of them they become established in the bed as the dominant coarse-grained fabric that controls the size of clasts which may then accumulate near them: clasts of a similar size are accepted and others are rejected (Bluck 1987). Thus, to redefine the grain size of the throughput population in a river, the load brought in by tributaries must be sufficiently voluminous or sedimentologically distinct. Tributaries which are able to modify the grain size of a trunk stream radically are referred to as significant tributary sources (Rice 1998). Clast sizes in the four links in the study reach fine down to approximately 30 cm, after which, the next link begins. Control on the position of the start of the links must be grain size, volume and composition of clasts supplied mainly by the tributary streams. Grain size alone is clearly not a fundamental control on the position of the four links, as evidenced in Fig.5.8a, where large grain sizes are supplied to the river by tributaries within all of the four links, yet very few are significant enough to start a new link.

The volume of material supplied by tributaries of the right size is undoubtedly very important in the location of links. However, it is extremely difficult to quantify the volume of material supplied by individual tributaries over an interval of time in an arid area such as the study reach where the recurrence interval and magnitude of significant events is unknown, and difficult to measure. The only clue to the activity or inactivity of a tributary is the presence or absence of a tributary delta at the intersection with the trunk stream. However, this evidence can be very misleading in areas where the tributary feeds into a high energy reach, and is quickly removed by the trunk stream. There is no evidence supporting increased volumes of gravel being supplied by the tributaries, such as larger or more active tributaries, at the start of the four links in the study reach.

In the study reach, the principal control on significant tributaries and the position of Links 1 to 4 is the composition of material supplied to the trunk stream by the tributaries. The coincidence of the beginning of a link with the input and survival of hard lithologies on the normal and strath bars is illustrated in figure 5.9, although this must also be coupled with significant volume additions of the resistant lithologies. In figure 5.9, the percentage of lithologies in the MCS population in the tributary bars is shown juxtaposed with the rock types forming the MCS population in the normal and strath bars. Initiation of Links 1 to 4 are coincident with the input of Nama quartzite, Rosyntjieberg quartzite, felsic (silicic) volcanics and Nama quartzite respectively. An alternative way of presenting this data spatially is by using pie charts superimposed on the satellite image of the study reach (cf. Figs.5.10 & 5.11). Separating the entire maximum clast size population into tributary bars and normal bars (comprising normal and strath bars), the whole dataset was analysed in terms of percentage lithology by number of clasts (count %), average roundness estimate, average clast length and clast length standard deviation (Fig.5.12). The tributary input bar composition is assumed to represent the assemblage of clasts entering the river via the tributaries. As noted previously, this assumption may not hold true as some tributaries are undoubtedly more active than others and thus their composition (active tributaries) should hold more weight. Also, some tributaries actively supply gravel, but do not have a tributary delta, and thus are not represented in the dataset in Fig.5.12.

All of the data is spatially sensitive, and is better analysed in shorter reaches, however, a few important trends can be recognised. Resistant lithologies such as Nama and Rosyntjieberg quartzite, felsic volcanics and vein quartz tend to increase in abundance after being worked onto the normal bars, whereas softer lithologies such as limestone, basement, granite, granodiorite, dolerite, mafic volcanics, metasedimentary and Karoo sedimentary decrease in abundance once in the river (5.12). Granite, which is the most abundant clast supplied to the river (based only on the percentage abundance present in the tributary input deltas), suffers the highest reduction and Karoo dolerite does not occur in any of the normal bars. As expected, the average roundness of all lithologies increases from the tributary to the normal bars and some lithologies such as diorite and granodiorite become very well rounded whereas lithologies such as the vein quartz and metasedimentary clasts remain subrounded. There is a reduction in clast size (length) and standard deviation from the tributary to normal bars in all lithologies, although, these data are particularly sensitive to position in the study reach (due to overall downstream fining), and is better analysed on a link by link basis.

Subdividing the dataset in Fig.5.12 into the four links (Figs.5.13-5.16), analysed for the whole reach length:

5.5.1.1 Link 1

Link 1 (Fig.5.13) is dominated by inputs of granite from the Richtersveld Suite (50%) and, to a lesser extent, Nama quartzite (19%), with lesser amounts of syenite, dolerite, felsic volcanics, mafic volcanics and Namaqua basement. The Nama quartzite clasts are added to the 1st half of the link, whilst the granites are added throughout the link (Fig.5.11). The normal bars in Link 1 see a large increase in Nama quartzite (47%) and decrease in granite (29%) through the complete link length. Basement, syenite and felsic volcanics show slight, insignificant increases in abundance after re-working whilst mafic volcanics show a slight decrease and limestone, dolerite, vein quartz and Karoo sandstones and shales disappear altogether from the MCS population.

Granites are the largest clasts supplied to the river in this reach. In terms of roundness, most of the lithologies move from the sub-angular to sub-rounded category in tributary bars to the rounded category in the normal bars, exceptions being basement and felsic volcanics which remain sub-rounded, and dolerite, which is already rounded in place on the tributary bars.

5.5.1.2 Link 2

Link 2 (Fig.5.14) is the link dominated by the Rosyntjieberg quartzite, which increases in abundance from 32% on the tributary bars to 79% on the normal bars. This increase is at the expense of all other lithologies in the reach, except for Nama quartzite that shows a slight increase. Once again the amount of granite is reduced significantly from 21 to 8%. Other lithologies reduced are basement, mafic volcanics, felsic volcanics and Karoo sandstone, whereas limestone, dolerite, granodiorite and vein quartz are completely removed from the MCS population.

The largest clasts supplied to the river in this link are granodiorite, granite and Karoo sandstones, whereas Karoo sandstones and the resistant Rosyntjieberg quartzite are the coarsest lithologies present on the normal bars. The presence of large Karoo sandstone boulders on the normal bars can be explained by them only being found in close proximity to the tributary bars which supply them, as evidenced by their relative lack of rounding compared to the other lithologies on normal bars in this reach.

5.5.1.3 Link 3

Link 3 (Fig.5.15) is the link dominated by felsic volcanics, which shows no change in abundance from the tributary to the normal bars. Interestingly, the tributaries bringing in large amounts of coarse felsic volcanics were not identified in this exercise as no such deltas exist at present. This provides proof that the lack of a tributary delta by no means excludes the tributary as a possible contributer of clasts. Nama quartzite, which is supplied as an already well-rounded population by the Fish River near the downstream end of the link increases from 6% in the tributary bars to 22% in the normal bars. Thus, the downstream end of this link is swamped by Nama quartzite, (Figs.5.9 and 5.11), but neither the size or the volume of Nama quartzite boulders added is large enough to re-set the texture of the bars and start a major new link. Only one clast of well rounded Rosyntjieberg quartzite is present in a tributary delta in this reach, and this is likely to have been supplied from upstream, emplaced in a flood. Nonetheless, Rosyntjieberg quartzite occupies 8% on the normal bars, despite having last been added at least 25 km upstream, demonstrating its high durability in the river system. In this link, the lithology which is severely depleted is granodiorite (25 to 5% count).

Granodiorite, diorite and basement are the largest clasts supplied to the river in this reach, with diorite, granite and mafic and felsic volcanics representing the largest average clast sizes on the normal bars. In terms of clast roundness on the normal bars, the mafic and felsic volcanics are the least rounded. Both the Nama (via the Fish River) and Rosyntjieberg quartzites (from upstream) are supplied to this reach as pre-rounded populations.

5.5.1.4 Link 4

Link 4 (Fig.5.16) starts at the point where large volumes of coarse, Nama quartzite and basement (from the area to the north of the study reach) are added by two large tributaries. After this point, the link is dominated by Nama quartzite, which increases from 26 to 68% from tributary to normal bars. Lithologies which are reduced substantially in this reach are basement (26 to 5%), granite (13 to 2% count), and metasedimentary clasts (21 to 14%). There are slight disturbances to the fining profile in Link 4, especially at ca. 200 km downstream, after an influx of metasedimentary clasts, however, not enough to reset the grain size completely.

Basement, diorite and granite are the largest clasts supplied to the river in this reach, whereas basement, Nama quartzite and mafic volcanics represent the largest clasts on the normal bars. Metasedimentary and vein quartz clasts remain fairly poorly rounded in comparison to the other lithologies in this reach.

5.5.1.5 Geomorphic Reaches

The lithological and grain size/roundness data for tributary and normal bars can also be analysed for each geomorphic Reach, as defined in Chapter 3 (Figs.5.17 & 5.18). Although some of the reaches represent only part of a link (Reaches 2, 3, 5, 6 and 7) or a combination of 2 links (Reach 4), it is important to consider as it is comparable to the data in Chapter 3 and represents smaller lengths of river than the links.

Reach 1 (Fig.5.17a) and Link 1 are identical.

Reach 2 (Fig.5.17b) is dominated by inputs of Rosyntjieberg quartzite with some minor volcanic and granite inputs, whereas the normal bars are completely dominated by Rosyntjieberg quartzites.

Reach 3, which traverses the Nabas basin, is dominated by inputs of a mixture of Rosyntjieberg quartzite, Karoo dolerite and Karoo sandstones (Fig.5.17c). The Karoo lithologies are almost eliminated leaving nearly 90% dominance of the Rosyntjieberg quartzite on the normal bars.

Reach 4, which has the highest number of tributary deltas per kilometre is dominated by inputs of granite, with slightly less amounts of felsic volcanics and granodiorites (Fig.5.17d). Felsic volcanics become the dominant lithology present on the normal bars (start of Link 3), with lesser amounts of granite and Rosyntjiebeg quartzite (from the tail end of Link 2) also present.

Reach 5, which is the tail end of Link 3, is dominated by inputs of basement, granites and felsic volcanics (Fig.5.18a), with minor proportions of Nama quartzite from the Fish River. On the normal bars, Nama quartzite is the dominant lithology, something that wasn't highlighted in the Link data. Granites and basement are reduced significantly in abundance, with felsic volcanics being slightly reduced. Remarkably, Rosyntjieberg quartzite still makes up a significant proportion of the lithologies present on the normal bars (11%).

Reach 6, the first part of Link 4, is dominated by inputs of Nama quartzite and basement and lesser amounts of mafic volcanics (Fig.5.18b). These later lithologies are virtually extinguished from the MCS population on the normal bars with Nama quartzite constituting 86% of the population.

Reach 7, the tail end of Link 4 is completely dominated by inputs of metasedimentary clasts (Fig.5.18c), as the river traverses the Gariep Supergroup. Nama quartzite, however is still the dominant lithology present on the normal bars, with metasedimentary clasts being the other significant lithology.

Interestingly, the abundant bedrock straths present in Reaches 1 and 4 (Section 3.2.1) coincide with the links where quartzite is not prominent. Links 2 and 4, dominated by Rosyntjieberg and Nama quartzite respectively, both lack exposed bedrock straths, thus, it would appear that the gravel composition present within a link exerts some control on the channel type present.

5.5.1.6 Downstream Changes per Lithology

The average percentage clast length reduction for key lithologies between tributary bars and normal bars was calculated for each Link (Fig.5.19). Granite and basement show the most clast length reduction for all of the links, an exception being granite in Link 3, which is probably related to the relative lack of quartzite in this link, allowing the granite to survive. Felsic volcanics show low decreases in clast length decline, always lower than the mafic volcanics. Rosyntjieberg quartzite has the lowest percentage reduction where present and Nama quartzite displays low percentage reductions when abundant in Links 1 and 4, but moderate reductions in Links 2 and 3, when not as abundant. Interestingly, apart from Link 1, the dominant lithology in the normal bars for each Link shows the least percentage reduction in clast length. Rosyntjieberg quartzite, which dominates the clast lithologies of Link 2, has the lowest clast length decrease for this Link. This is repeated in Links 3 and 4 with felsic volcanics and Nama quartzite respectively. The fact that this pattern is not observed in Link 1 may reflect the fact that the MCS is composed of a combination of large clasts of granite and smaller clasts of Nama quartzite, and not one dominant lithology, or that granite is not very competant in a river system.

Changes in clast size and roundness with distance travelled per lithology on each bar are illustrated in Figs.5.20 and 5.21. The fact that the dominant lithology in each reach controls the grain size and roundness is illustrated in these diagrams, where the average clast length and roundness per bar matches that of the dominant lithology in Links 2, 3 and 4. In Link 1, the average clast length in the 1st part of the link matches that Nama quartzite, and diverges after an influx of coarse granite. A more detailed version of figure 5.20, which includes the clast length data per lithologies for tributary input bars is shown in figure 5.22.

Nama quartzite is found as relatively small clasts in Links 1, 2 and 3 (mostly 30-40 cm), whereas in Link 4 it occurs as large clasts on the normal bars (40-60 cm). In Links 1, 2 and 4, it is supplied as large clasts (Fig.5.22), whereas in Link 3, it is supplied as relatively small clasts by the Fish River. The difference between Links 1, 2 and 4 is the volume of large clasts supplied in Link 4. This difference in volume can be attibuted to the fact that in Links 1 and 2 Nama outcrops at river level, whereas in Reach 4, the base of the Nama group is approximately 500 m above the Orange River level. Thus the tributaries in the latter area are able to bring more large-sized Nama quartzite clasts to the river from much further distances than in Links 1 and 2. The selection and rejection of clasts on bars, as described by Bluck (1982; 1987) is illustrated well where, after an influx of metasedimentary clasts in Link 4 (at Daberas), there is an increase in the average grain size. At this point, there is also an increase in Nama quartzite grain size, which, up to this point was steadily decreasing in size. Thus, the passing population of Nama quartzite clasts which have a coarser grain size than that present on the bars before Daberas do not fit into the fabric on the bar and are rejected, whereas, after Daberas, the fabric on the bar (provided by the metasedimentary clasts) is coarse enough to trap the passing population of larger Nama quartzite clasts. It must be noted that clasts that do not fit into the fabric of bars are rejected into the channels and pools that the current sampling did not reach (Bluck 1979).

In terms of clast roundness, the Nama quartzite population shows a steady increase in rounding in Link 1. This is disturbed in Link 2, after which the roundness increases until a maximum in Reach 3, after which it is disturbed at the beginning of Link 4, but increases again quickly (5-10km), after which it decreases steadily, contrary to expectation. This apparently anomalous decrease in roundness with distance travelled can be explained by the concomitant decrease in grain size with distance travelled. Roundness generally increases with increasing grain size, up to a limit, after which it decreases (Pettijohn 1957; Bluck 1969b). In the study reach, this decrease in roundness is also partly due to the accumulation of many damaged or broken clasts of quartzite in these distal regions (Fig.5.23).

Rosyntjieberg quartzite (Figs.5.20 & 5.22) shows a steady decrease in size in Link 2 on the normal bars. The clast size of Rosyntjieberg quartzite supplied by the tributaries also decreases towards the distal end of this reach. These are tributaries that have traversed the Nabas Basin and are thus further from the Rosyntjieberg quartzite outcrop, hence deliver smaller clast sizes. The grain size of the Rosyntjieberg quartzite increases markedly at the start of Link 3, this, once again being another good example of the selection and rejection of clasts on the normal bars. For the remainder of the river, Rosyntjieberg quartzites generally decrease in size and are a slightly smaller size than the average clast size per bar.

Rosyntjieberg quartzites enter the river in Link 2 as subangular clasts and steadily increase in roundness to the end of the Link (Fig.5.21). At the start of Link 3, the roundness drops off and then increases towards the distal end of the Link, and this repeated at the start of Link 4. This decrease in rounding at the start of the Link is either related to the change in grain size, or is due to damage caused by the larger framework clasts present on these bars. The low roundness values of Rosyntjieberg quartzite in the distal part of Link 4 reflect input of this lithology by the Annis River, a tributary downstream of Auchas Major. Although no tributary delta is present at its intersection with the Orange, terraces and bars on this tributary reveal the presence of Rosyntjieberg quartzite. The extreme distal end of Link 4 displays a well rounded population of Rosyntjieberg quartzite clasts.

In Link 1, felsic volcanics maintain a fairly even grain size, however, they display a relatively rapid decline in clast length in Link 3. Once again, at the start of Link 4 the grain size increases, this either being another example of clast selection, or there is a tributary supplying this material which was not detected in the tributary bar dataset. Clast roundness of felsic volcanics is fairly haphazard, but generally increases in the distal parts of Links 1 and 3, and then slowly declines throughout Link 4, with decreasing grain size (Figs.5.20 & 5.22).

Granites and Namaqua basement, represent the most frequent inputs along the Orange River channel in the study reach and have the widest scatter of clast lengths, although, distinct fining can be recognised in each of the reaches. Roundness data for these lithologies is particulary haphazard, this being due to the high turnover of material which is supplied, rapidly rounded and destroyed. In the distal part of Link 4, granite is better rounded than basement, and this may be due to the foliated nature of the basement (Figs.5.20 & 5.22).

5.5.1.7 Variations within the dataset

Some closely spaced bars, away from the influence of tributary inputs, display large contrasts in clast size or roundness consistently throughout all lithologies present on those bars. These variations can be seen in the data in figures 5.20 and 5.21. A similar "seemingly random" variation of roundness values between bars was observed in the Lower Colorado River (Sneed & Folk 1957), which was ascribed to local variation in stream characteristics. The position of bars relative to the active channel and the passing bedload must determine the size and roundness of material present on the bar to some extent. An example of this is from Link 4 where a significant difference between a bedrock attached strath bar and adjacent point bars can be seen (Fig.5.24). The clasts on the strath bar are locked into a tight fabric and are able to withstand higher energy flows as opposed to the point bars, whose largest clasts are not locked into any fabric and are less mobile.

5.5.2 Maximum Clast Size Variations on Individual Bars

The bars chosen for clast size decline analysis are located in the Sendelingsdrif area and downstream of Modderdrif (Fig.5.25a). These localities were chosen as the Sendelingsdrif area is essentially out of the influence of tributary input, yet the bars are gravel-dominated. Downstream of this area, the bars become pebble and sand dominated fairly close to the bar head. The Modderdrif site was chosen to provide an example from a smaller, coarser bar. In all cases (Fig.5.25b), maximum clast size decreases from the bar head towards the tail. The variations in the data, especially from Bar DZ are from the presence of small unit bars (Bluck 1987) which are migrating across the bar surface. The high rate of size decline on these individual bars is evident, when compared with the dataset per bar from the entire study reach (Fig.5.25c). The size decline on each of these bars is equivalent to between 30 - 50 km of downstream fining along the channel. This size decline along these bars is thought to be achieved by the process of selective sorting (Bluck 1982).

5.6 Results – Ancient deposits

Typically, the MCS in the ancient deposits is considerably less than that in the modern system in the Richterveld reach between Noordoewer and Sendelingsdrif (Fig.5.26a). These results are anomalous as the modern system has traditionally been viewed as a less competant system than the ancient systems (Jacob *et al.* 1999; Ward *et al.* 2002). However, as explained in Section 5.3, the value of MCS obtained from the terrace deposits is a minimum value and may not be directly comparable to the value obtained from the modern river bars. However, downstream of Sendelingsdrif, even this minimum MCS obtained from both the Proto and Meso systems is coarser than the modern system. Thus, the rate of downstream fining in the modern river is higher than in the terrace deposits, as illustrated by the linear trends applied to each dataset (Fig.5.26a). Most of this downstream fining occurs in Link 4, after the addition of coarse Nama quartzite. To effect a more direct comparison, without the effects of tributary input, only the Nama quartzite data for Link 4 was plotted for each terrace suite (Fig. 5.26b). Once again, the modern river displays the highest rate of fining followed by the Proto and Meso suites, which have very similar rates of downstream fining.

The composition of the MCS between the modern and terrace deposits also differs somewhat through the study reach (Fig. 5.27). In the Richtersveld reach, between Noordoewer and Sendelingsdrif, the MCS of the modern river has a more mature clast assemblage in terms of quartzite content. At Noordoewer, the proportion of Nama quartzite increases from the Proto towards the modern river. In Link 2, Seven Pillars, Grasdrif and Gamkab, there are higher proportions of Rosyntjieberg quartzites present in the modern river, and at Grasdrif, the Proto has the least amount of quartzite. In Link 3, at Block 6, the proportion of felsic volcanics is greater, and at the Fish River, a higher percentage of Nama quartzites is found in the modern river. In Link 4, at the Boom River, the proportion of Nama quartzite present is roughly the same through the ages. At Lorelei East and Lorelei West, the Pre-Proto and Proto deposits have a much more diverse MCS composition than the Intermediate, Meso and modern deposits, whilst at Sendelingsdrif, the MCS composition of the Pre-Proto, Proto and modern deposits are very similar.

Downstream of Sendelingsdrif, the opposite trend occurs: the modern river displays a more diverse MCS assemblage whereas the terrace deposits are more mature with respect to the percentage of Nama quartzite. Metasedimentary clasts which are locally sourced from the Gariep Supergroup constitute a large proportion of the MCS population in the modern river, whereas they are virtually absent in the MCS populations of the Proto and Meso deposits. Apart from at Daberas, where the Proto deposit has a more diverse composition, the Proto and Meso MCS compositions are similar.

5.7 Discussion and Summary

5.7.1 Modern River

In the study reach between Noordoewer and Oranjemund, lateral sediment sources from tributaries are the primary control on clast size variations on the normal and strath bars present along the channel. Although there is a balance between tributary gradient and drainage basin size, the primary control on clast size in the tributary input bars is lithology, with the coarse crystalline varieties (granite, granodiorite, gneiss) yielding the largest clasts.

Within the study reach, there are four well defined sedimentary links, where, within each of these links the average MCS declines to ca. 30 cm, after which the grain size is re-set by fresh tributary input. The position of the start of these links is determined by the input of a large enough volume of big durable clast types (Nama quartzite, Rosyntjieberg quartzite and felsic volcanics). These resistant lithologies tend to dominate the clast assemblage of the MCS population, thus the fining of these populations essentially determines MCS fining within the link (Fig.5.22). Tributary bars (representing clasts entering the Orange river) have a minority resistant suite of clasts, thus the dominance of resistant lithologies remaining in the MCS population indicates much faster breakdown of other softer lithologies. The process of abrasion is likely to be more responsible for this MCS compositional dataset than selective sorting. Abbott & Peterson (1978) recorded faster rates of abrasion when clasts of different durabilities were tumbled together in their experiments, and it is thought that this is analogous to what is happening in the study reach.

Selective sorting is a very effective process in the downstream fining of gravel (Hoey & Ferguson 1997), as evidenced in the data by the fining of the MCS down individual bars (Fig.5.25c), as well as in how effectively the MCS population is sorted from the tributary input to the normal and strath bars (Fig.5.6d). The sudden coarsening of lithologies at the start of a link, when no tributary input is bringing them in (e.g. Rosyntjieberg quartzite at the start of Link 3), can also be best explained by selective sorting by the process of selection and rejection of clasts (Bluck 1982;1987).

The resistant lithologies experience the least reduction in clast length between the tributary and normal bar MCS population (Fig.5.19). Unfortunately, the dataset does not allow the analysis of detailed size decline with distance downstream per lithology. To achieve this, the largest 10 or 20 clasts per lithology would have needed to be measured within a given area. The dominance of the resistant lithologies in the present dataset results in exclusion of the softer lithologies from the MCS population.

Although disrupted severely by tributaries, within Links 1 to 3, the clast roundness of the overall and dominant lithology tends to increase with distance along the links (Fig.5.21). In Link 4, the anomalous decrease in roundness with distance of most lithologies can be explained by the concomitant decline in grain size.

It appears that the composition of the gravel within a link exerts some type of control on the channel type present. Reaches dominated by quartzite lack inner channels flanked by bedrock straths (Table 3.1).

5.7.2 Comparison with the Terrace Deposits

The MCS measured from the Proto and Meso suites of terraces is considerably finer grained than that of the modern river bars in the Richtersveld reach between Noordoewer and Sendelingsdrif. However, the rate of downstream fining in the ancient deposits is considerably lower than in the modern system and thus, after Sendelingsdrif, where the grain size is roughly equal, the modern MCS declines steadily so that by Arrisdrif the terrace deposits are much coarser than the modern system (Fig.5.26). In addition to the size differences, the lithologies of the MCS populations differ from the modern to the ancient deposits. Between Noordoewer and Sendelingsdrif, the modern river has a more mature MCS population with respect to the amount of resistant lithologies present, whereas between Sendelingsdrif and the mouth, the opposite is true and the terrace deposits have more mature MCS populations (Fig.5.27).

Reasons for the higher rate of downstream fining in the modern system could be either lower stream competence in the modern system than in the ancient deposits, or less tributary gravel is being supplied to the modern river. In general, changes in water discharge affect downstream fining: with less fining occurring at greater discharges, when sediment transport capacity is increased (Hoey & Ferguson 1997; Hoey & Bluck 1999). However, the rate of downstream fining is reduced when rates of sediment supply are increased (Hoey & Ferguson 1997; Hoey & Bluck 1999), although not all workers found this to be the case (Seal *et al.* 1997). Bluck (1987) related the fining rates on alluvial fans to the fan radius, i.e. the bigger the fan, the lower the fining rate. Although not directly transferrable to gravel bed rivers (Hoey & Bluck 1999), the simple process of sediment dispersal of tributary gravel by a trunk stream is analogous to the alluvial fan case: the more sediment supplied, the further it goes (Bluck 1987).

Increased river competence and larger supply of gravel by tributaries may both be responsible for the lower rate of fining in the terrace deposits, although the lithological differences in the MCS population may assist in differentiating the relative importance of each. The relative diversity in lithologies of the MCS in the terraces relative to the modern river between Noordoewer and Sendelingsdrif points to increased amounts of clasts entering the river via the tributaries in former times. If anything, more water in the ancient river system would have lead to increased rates of abrasion and a more durable MCS population. This in turn could imply that there is more water in the modern river than the Proto or Meso-Orange Rivers which is in contrast to most evidence which indicates aridification of this area from Miocene to Present (Section 4.6.1). The more mature MCS population present in the modern river in the Richterveld reach more than likely indicates a relative lack of supply of gravel by tributaries, due to the flattening of tributary profiles and maturing of the landscape, allowing a very coarse, mature population of coarse clasts to populate the various bars. So, although the river was likely to be more aggressive in former times, it was not aggressive enough to cope with the increased sediment supply.

The more mature MCS assemblage present in the terrace deposits downstream of Sendelingsdrif relative to the modern bars would appear to contradict the above argument. However, it is still compatible with increased tributary supply in the ancient deposits as the tributaries that supply Link 4, essentially supply Nama quartzite to the system, thus large amounts of coarse Nama quartzite would be expected to populate the bars downstream of these inputs. A longer tail of Nama quartzite boulders would result in more abrasion occurring further downstream on the coastal plain, thus less metasedimentary rocks are found in the MCS populations from the terrace deposits. However, the relative lack of Gariep Supergroup metasedimentary clasts in the ancient MCS populations also supports a more competant river with higher amounts of abrasion in the older rivers.

The longitudinal profiles of the Proto straths appear to be grading to a coastline situated further offshore, and the lowest point of the Mid-Oligocene lowstand certainly supports this (Siesser & Dingle 1981). Thus, the lower rate of size decline in the older deposits may be reflecting a coastline further out than the present.

As discussed in Section 4.6.1, the very fact that gravel deposits were built up and abandoned as terraces indicates that at times, sediment supply exceeded transport capacity, thus increased sediment supply is thought to be the primary control on the rate of downstream fining through time. The changes in composition and rate of size decline of the MCS population through time lend weight to the argument that aggradation is due to increased sediment yield from the tributaries.

5.8 Synopsis

- Tributary inputs are the primary control on grain size in the Orange River bars within the study reach.

- Clast size in the tributary bars is not a function of tributary gradient or drainage basin area, but clast lithology. Tributaries dominated by granite, granodiorite or basement clasts have the largest clast sizes.

- In the study reach, there are 4 sedimentary links, the position of the start of a new link being controlled by significant additions of durable clast types, namely quartzite and felsic volcanics, with the Orange River bars in the remainder of each link being dominated by these lithologies.

- Clast abrasion is likely to be responsible for the dominance of resistant lithologies on the normal bars, although examples of selective sorting are also evident in the dataset.

- Links dominated by quartzite clasts lack any bedrock straths, thus clast lithologies exert some type of contol on the channel present.

- The rate of downstream fining in the Proto and Meso deposits is significantly lower than in the modern river. This is indicating greater volumes of tributary input in the older systems, a more competant river and a coastline offshore the present coastline.

- The composition of the MCS population indicates more tributary input, as well as a more competant river in the older deposits.

6 Clast Composition - Changes with Grain Size

6.1 Introduction

Smaller clasts travel further in a fluvial system, so they have the potential to provide provenance information for the entire Vaal/Orange drainage basin. Thus, analysing the terrace record through time should provide data on the evolution of the entire Vaal-Orange basin.

Preliminary observations (van Wyk & Pienaar 1986; Ward & Bluck 1997) have shown that the gravels in the Orange River course and along the Atlantic coast record a changing temporal clast assemblage documenting the changing rock types available to the drainage during its period of down-cutting and evolution. The principal aim of this part of the study is to record and explain the change in clast composition over time. The modern river, and the series of terraces preserved in the Lower Orange valley (Section 4.3), are examined with a view to reading, with greater precision, the Cainozoic evolution of the drainage network. Concomitantly, changes along the channel within all of the terrace suites are analysed. However, it is acknowledged that by the time the oldest of the terraces were deposited, considerable erosion into the basement and cover had already occurred (Section 4.6), thus, a large portion of the early incision history is missing from this record. The Eocene to Plio-Pleistocene age of the majority of the deposits makes them too old for commonly used dating methods such as cosmogenic (up to c.2 Ma) or luminescence (1 Ka to 1 Ma) dating (Burbank & Anderson 2001). With the exception of the Proto deposits, no dateable fossils have been found thus far. The methods remaining to characterise the various deposits are the correlation of strath heights (Section 4.5) and clast assemblage/petrography. The use of clast composition provides a check on the strath-based terrace classification scheme.

Clasts in the Lower Orange River deposits could have origins anywhere within the current drainage basin (Fig.2.2). Clasts could also be derived from outside of the current drainage basin, either brought in by the glaciers responsible for the many glacial deposits, or by the process of drainage capture. In addition, there is the added complication of inheritence of clasts from the older deposits present in the younger deposits. The large time difference between the Proto and Meso suites of deposits should minimise these effects, but the modern deposits could be contaminated with clasts which have been derived from the Meso deposits.

6.2 Previous Work

In attempting to reconstruct drainage basin and provenance changes, several authors have resorted to the use of a combination of strath and terrace levels and clast composition. In the Lower Susquehanna River terraces, Pazzaglia and Gardner (1993) used petrography and elevation to distinguish two groups of terraces in a study investigating the late stage evolution of the U.S. Atlantic passive margin. In the Canadian River, New Mexico, terraces were correlated according to soils, elevation and clast type in a study of the long-term landscape evolution of the Canadian River and Canadian escarpment of northeastern New Mexico (Wisniewski & Pazzaglia 2002).

Bluck (1969a; 1980) and Haughton & Bluck (1988) used clast assemblages per size fraction to unravel the complexities in the conglomerates of the Old Red sandstone determining that the material was sourced from opposite margins of the basin at different times. Abbott & Peterson (1978) used the differing clast assemblage in 2 suites of conglomerate in California to infer the existence of a major river in that region.

In the Middle Orange River, ca. 30 km downstream from the Vaal/Orange confluence, McCarthy (1983) used clast assemblages in older and younger terraces to infer the existence of a large river that he called the Trans-Tswana River. On the Lower Vaal River terraces, Spaggiari (1993) and Spaggiari *et al.*(1999) analysed the clast composition of the Droogeveldt gravels but could not find a distinct pattern, which they attributed to the input of clasts from the Dwyka tillite. However, their analysis led to the discovery of a banded iron formation-carrying tributary entering the Vaal system on the right bank.

On the Lower Orange River terraces, two attempts have been made at quantifying the clast composition of the Proto and Meso terraces. Starke (1965) and Fowler (1976) did clast counts of 200 and 4000 clasts respectively, and obtained roughly similar results. The Proto gravels were distinguished from the Meso gravels by their greater concentration of quartz, igneous and metamorphic clasts, together with a higher proportion of unspecified "exotic" clasts, whereas the Meso terraces had higher concentrations of banded iron formation (BIF) clasts. These observations were confirmed by observations by van Wyk & Pienaar (1986).

In many studies of terrace deposits, much reference is made to clast assemblages and the presence or absence of particular clast types (Partridge & Brink 1967; Helgren 1979b; de Wit *et al.* 2000; Pether *et al.* 2000; Pazzaglia & Brandon 2001), but very few systematic, quantitative clast assemblage studies have been done. Most detailed clast studies have concentrated on modern river gravels in order to understand basin-wide sediment dynamics (Hack 1957; Jones 2000), pebble morphogenesis (Wentworth 1922; Sneed & Folk 1957) or the problem of downstream fining of gravel and unravelling the relative role of sorting vs abrasion (Krumbein 1941; Werritty 1992; Huddart 1994; Kodama 1994a; Jones & Humphrey 1997; Rice 1999).

6.3 Methodology

Due to the selective transport and deposition of clasts, locally sourced varieties are more likely to occur in the coarser fraction (+90 mm), whereas smaller clasts can be sourced from further afield. In addition to this, clast abundance is generally inversely related to clast size (Bluck 1987). Thus, clast assemblages are grain size dependant, and any analysis needs to take this into account.

Methods of analysing clast assemblages differ. Pazzaglia & Gardner (1993) used clasts in the size range 20-100 cm to characterise the assemblage, although the number of clasts counted at each site was not specified. Wisniewski & Pazzaglia (2002) used 100 randomly selected clasts from each site, although the size range of the clasts sampled were not specified. McCarthy (1983) did clast counts of 150 per site, and clast sizes ranged from 3 cm to boulders. In this case, the resultant clast type proportions are sensitive to the grain size distribution sampled, and thus the results are significantly flawed. The studies done by Starke (1965) and Fowler (1976) at Daberas within the study reach did clast counts of 200 and 4000 clasts, respectively. However, as in McCarthy's study, both studies sampled the full size range from boulders to pebbles without differentiating between the different size classes and consequently their results are not very meaningful, as a bias toward any size range will bias results drastically.

On the Lower Vaal River, Spaggiari (1993) and Spaggiari *et al.*(1999) recognised the bi-modal nature of the gravel and analysed the assemblage of the coarse and fine clasts separately. The framework clasts were made up exclusively of locally derived Ventersdorp lavas, whilst the smaller clasts contained a variety of lithologies. This detail would have been missed if all sizes had been analysed together. Bluck (1980) did clast counts of >300 clasts per size fraction, and up to 6 size grades per site in his studies of the Old Red Sandstone.

In this study, to avoid the pitfalls mentioned above, the following size categories were used: +300 mm, 200-300 mm, 90-200 mm, 40-90 mm, 25-40 mm, 16-25 mm, 8-16 mm and 3-8 mm. These intervals were designed to match the data from bulk size distribution studies done for diamond prospecting, hence the standard phi scale intervals were not used. The sampling procedure that was used for the terrace deposits is as follows: Only excavated sections through the deposit were sampled, as opposed to natural terrace deflation surfaces, which could bias the results to clasts that are resistant to weathering. An area of relatively homogenous gravel was analysed, generally not exceeding 2 m thick and 4 m wide. Initially, the long axis of the largest 20 clasts within a 2x2-m area was measured and their lithologies were noted. A visual estimate of the roundness (between 0.1 angular and 1.0 very well rounded) was made for each clast, using the scheme of Powers (1953). The additional axes were not measured and clast shape not estimated as the majority of clasts were embedded in the face. After the initial 20 clasts were measured, the lithology and roundness of each clast was assigned a size category and this continued until all the required clasts were noted, or the specified sample area had yielded all of its eligible clasts. This was done for the +300, 200-300 and 90-200 mm size fractions. For the remainder of the size classes, material within the 2x2-m area was dug out and sieved into the various size fractions. Material from the sieves was split in a standard splitter until the required number of clasts was obtained, after which these clasts were sorted into lithological classes. Samples were bagged, labelled and stored in case further information was needed from them.

For clasts in the 200-300 mm and +300 mm classes, between 20 and 30 clasts were analysed, depending on availability of that size in the sampled horizon. For the intervals 25-40 mm, 40-90 mm and 90-200 mm, a total of 50 clasts, and for the remaining smaller size categories, a total of 100 clasts were analysed per size fraction. During the analysis of the 3 finer fractions, tests were done on the representivity of the sampling and it was found that the proportions of the various lithologies were relatively fixed after counting between 40 and 60 clasts. However, the full amount of clasts were always counted because, although the percentages remained fairly constant, the possibility of encountering rare clast types would have been reduced if the number of clasts counted had been reduced. A typical sampling face, together with the equipment used in the sampling is shown in figure 6.1, and a summary of the sampling methods and amounts sorted is shown in figure 6.2a. For samples in the modern river, a similar method of sampling was used, with the horizontal bar surface, as opposed to a vertical face, being analysed. As with the MCS measurements in Section 5.3, the bar head was always targeted.

The precision of the compositional data is difficult to estimate because of the inherent variability of clast composition. The visual scheme developed for point counting of mineral grains was used to estimate the accuracy of the data (Fig.6.2b)(van der Plas & Tobi 1965). This method is based on twice the standard deviation and for 20-30 clasts, the percentage error is 14-16%, depending on the percentage composition of a particular clast type, and for 100 clasts, this error is 4-10% (Fig.6.2b). Thus, minor clast types are not statistically significant and are of little quantitative value. However, the continual appearance of certain minor clast types in one suite of deposits and absence in another, although not statistically significant, is significant and can be compared to the discovery of a fossil or rare sedimentary structure, which, although rare, are very significant. The interpretation of roundness data is difficult given the statistical uncertainties arising from the use of a semi-qualitative and subjective scale, thus care must be taken in the use of this roundness data.

A total of 78 samples were analysed (Fig.6.3). Of these, 71 were from the study reach (Fig.6.4a), 4 were from the Augrabies Falls area (Fig.6.4b) and 3 samples were from marine/aeolian re-worked Eocene-age deposits in the Bogenfels area (Fig.6.4c). Of the samples analysed in the study area, a total of 34 were from the Proto Suite, 29 were from the Meso Suite and 8 were from the modern river bars. An attempt was made to analyse the lower and upper gravel of every available terrace deposit on the Namibian side of the river, as well as all accessible deposits on the South African side so that longitudinal and vertical changes in the clast composition could be gauged within a particular terrace suite, and between the different suites.

In an attempt to link composition per size fraction to the relative proportions of clasts present in the gravel, to obtain an overall composition weight percent of the gravel, some of the clast assemblage samples analysed with the method described above were combined with accurate size distributions. Size distributions were done on six of the clast assemblage samples from the Sendelingsdrif and Daberas Proto deposits. Grain size distributions in coarse sediment require large samples, to ensure that the weight of the largest clast does not exceed more than 2% of the total weight of the sample (Kodama 1994a). Thus, to minimise error a large sample between 4 and 7 tons of gravel were hand screened through a series of sieves for the size distribution analysis. Clast assemblages, based on the weight percentage of each size class were calculated, the assumption being that the different clast types are of equal density.

6.4 Lithologies

In addition to the MCS lithologies described in Section 5.4, a number of exotic lithologies, derived from outside of the study reach, are found in the gravel clast assemblage in the finer gravel fractions. Due to the large contribution from the NMP lithologies, an exotic is defined as any clast which is derived upsteam of the Upington/Groblershoop area, upstream of the NMP.

6.4.1 Banded Ironstone

Commonly known as banded iron formation (BIF), these are fine-grained, highly siliceous and resistant iron-rich rocks that have a very distinctive appearance and are best developed in the Asbestos Hills and Koegas Subgroups of the Ghaap Group (previously the Griqualand West Sequence) of the Transvaal Supergroup (SACS 1980). There are also limited occurrences of BIF in the Chuniespoort Group of the main Transvaal basin, and in the ancient Kraaipan Group (>3.1 Ga) (SACS 1980). BIF is most commonly found as flat, elliptical pebbles (3-8 mm, 8-16 mm and 16-25 mm categories), although they do sometimes appear in the 25-40 mm and 40-90 mm categories, especially in the Augrabies Falls area. They are commonly brown coloured and banded [magnetite rhythmite *sensu* (Beukes & Dreyer 1986)], although unbanded, mono-coloured varieties are also present [lutites *sensu* (Beukes & Dreyer 1986)]. When brown, the shade of brown varies from light, honey coloured to dark brown, and sometimes almost black. Where banded, the banding is usually in the form of thin black magnetite layers or alternations of different shades of brown. Less common folded varieties are also present. Other very distinctive varieties of BIF are:

- Jasper (jaspilite), which is generally red and mono-coloured, although you sometimes find bands of red jasper in the common brown BIF.
-Riebeckite is a blue, mono-coloured variety of BIF in which the mineral riebeckite is abundant, although blue and brown mixtures do exist in unbanded varieties.
-Pearsons BIF is a dark purple coloured iron-rich variety.
-Metallic BIF is a less common variety with a metallic lustre.
-Tigers-eye (crocilodite asbestos) with a wavy, fibrous texture is present in minor amounts.

All varieties of BIF are highly resistant and denser than most other clasts. Examples of the various types of BIF can be found in Fig.6.5a.

6.4.2 Karoo Sediments/Hornfels

Flat, elliptical pebbles of shale and sandstone (less common) sourced from the Karoo basin are commonly found in the -40 mm size fractions. Although also outcropping in the study reach, clasts of Karoo shale and sandstone are prevalent along the entire length of the Orange/Vaal system and are thus considered with the exotic suite of clasts. The shale ranges from being relatively soft and fissile to a hard hornfels, a result of contact metamorphism by the numerous Karoo dolerites. These baked varieties have been referred to as lydianite (pyroxene hornfels) (McCarthy 1983). However, many of the softer varieties display a characteristic spotted hornfels texture (Fig. 6.5b) and it is suspected that most of the pebbles that are able to survive the journey to the study area have been subject to some degree of thermal metamorphism.

6.4.3 Karoo Basalt and Zeolite

Grey to red and purple-weathering basalts with abundant zeolite amygdales from the Jurassic Drakensburg Group (capping the Karoo sedimentary pile) are commonly present in the small pebble categories (+3 and +8 mm categories). White zeolite clasts, which have apparently been liberated from the basalt host rock are also found in close association with the basalt (Fig.6.5c). These clasts are common in the Orange River which drains the Drakensberg but are absent along the modern and older Vaal River deposits.

6.4.4 Agate and Chalcedony

Although agate is a banded variety of chalcedony (O'Donoghue 1987), they are considered together as there is considerable overlap between them and are hereafter referred to as "agates". Agates from Vaal River and Orange River sources can be readily distinguished. Vaal River agates (Fig.6.5d) tend to range in colour from pale yellow to a dark, resinous, almost black colour. Both clear chalcedony and banded agate varieties exist for all of these colours. There are also black and white and grey and white banded varieties, as well as some minor red and green varieties. Orange River agates (Fig.6.5e) tend to be clear or light grey to bluish grey in colour, and are sometimes tube-shaped, pipe amygdales.

The Orange River type agates are sourced from the Drakensberg Group basalts. The Vaal River type agates are sourced from both the Ventersdorp andesites and from an as yet, undiscovered source. Agates from the Ventersdorp volcanics are commonly the black and white, and grey and white banded varieties, although some of the resinous and red coloured types may also originate from here. A minor component of Drakensberg-type agates is also present in the Vaal River population. The yellow chalcedonies and agate found in the Vaal River deposits, as well as red carnelian and green chrysoprase are found in abundance in certain Lichtenburg gravel deposits. Although there is no direct evidence in terms of outcrop, they are thought to have been derived from the erosion of Karoo basalt that once covered that area (Armstrong *et al.* 1987).

Thus, when evaluating the agates in the study reach, the Ventersdorp volcanics and Lichtenberg-type agates are assigned a Vaal River source (Fig.6.5d), whereas the blue-grey and clear agates are asssigned an Orange River source (Fig.6.5e). Although this biases the population towards an Orange River source (as the blue-grey and clear varieties are also found in small amounts in Vaal River deposits), it is a useful approximation as the Venterdorp and Lichtenberg type are not present in the Orange River assemblages.

6.4.5 Chert

Chert is found mainly in the -40mm fractions. It is sourced mainly from the Transvaal Supergroup, although very limited amounts are also present in the Nama Group. Common colours include black, grey and red, although a wide range of colours is exhibited (Fig6.5f).

6.4.6 Epidosite

Epidosite pebbles are sourced locally from both the Vioolsdrif Suite (basement) and the Orange River Group volcanics. Due to the similarity in specific gravity of epidote (3.3-3.6) with diamond (3.50-3.53) (Sinkankas 1964), it is an important indicator mineral in diamond prospecting, and thus is considered as a separate clast type (Fig.6.5g).

6.4.7 Feldspar

The breakdown of the numerous granites and gneisses in the gravels yields abundant feldspar, which are predominantly found in the small grainsize fractions (Fig.6.5h).

6.4.8 Other

Clasts encountered too rarely to be assigned an individual category are metamorphic garnet (sourced from the NMP rocks), calcite, gossan, conglomerate and Makwassie porphyry (Fig.6.5) - a distinctive reddish/purple quartz porphyry that is abundant in the younger beaches.

6.5 Results

The list of clast assemblage samples (which appear in Figs 6.3 & 6.4), together with their position (latitude, longitude) can be found in Table 1, Appendix D. In addition to this, figures 1 to 78 in Appendix D details the clast assemblage results from each sample (percentage lithology per size fraction), average roundness estimate per lithology for each size fraction and pictures of the sample locality. For ease of comparison, the graphs appearing in Figs.6.7 to 6.43 have been reduced but should further detail be required then consult the original in figures 1-78 (Appendix D). The clast assemblages will first be discussed at each site to compare changes through time, after which the changes in clast assemblages along the channel will be discussed. Samples in the study reach (Fig.6.4a) will be discussed first, followed by samples along the Sperrgebiet coast (Fig.6.4c) and in the Augrabies Falls area (Fig.6.4b).

6.5.1 Clast Assemblages per Site

In the diagrams relating to the following section (Figs.6.7 - 6.37), at each site, the deposits are ordered in the sequence of deposition, from oldest to youngest (to be read left to right). Lower samples refer to those situated within 1.5 m of the bedrock contact, and Upper samples are those situated in the upper units of a deposit. Localities referred to are illustrated in figure 4.1, and specific sample locations are found in Figs.6.4a-c.

6.5.1.1 Noordoewer area

The Noordoewer terraces (Fig.6.6) are situated in the Nabas Basin on fairly incompetent Karoo shales, and as a result, the bedrock is flat and featureless. A sample from the Proto deposit (Sample 11), from the lower gravel (i.e. close to the basal bedrock contact, and one from the Meso deposit were analysed (Sample 13)(Fig.6.4a and 6.7).

The samples at this locality are broadly similar, and have decreasing proportions of Nama quartzite, and increasing proportions of the more labile lithologies towards the finer size fractions. In the coarse size fractions, +300, +200 and +90 mm, the Meso deposit has higher proportions of Nama quartzite, and less mafic volcanic clasts than the Proto deposit, whilst both have roughly equal amounts of granite and basement clasts. In the finer size fractions, the Meso sample has significantly higher proportions of granite and feldspar clasts than the Proto. The Proto has more Karoo sedimentary/hornfels clasts than the Meso, as well as more BIF. In addition, BIF clasts are present in the +40 mm size fraction in the Proto, but absent in the Meso sample. Riebeckite is present in both samples. Karoo basalt and agate clasts from both Vaal and Orange River sources are present in the Proto deposits, but are virtually absent in the Meso.

The average clast roundness of the two Proto samples are comparable, moving from the well rounded category in the large sizes down to the sub-rounded category in the fine size fractions. The Meso sample is more rounded in the coarse fraction, but less rounded in the +40 mm category and finer.

6.5.1.2 Seven Pillars area

Seven Pillars is situated in a rugged topographic area very close to the beginning of Reach and Link 2 (Fig.4.1 & Fig.6.8). In this area, two samples were analysed: 1 from the Meso terrace (Sample 14) and 1 from the modern river (Sample 68)(Fig.6.4a and 6.9).

Due to the lack of any adjacent bars in the modern river, the modern sample was taken a few hundred metres upstream of the first significant Rosyntjieberg quartzite input. As the Meso terrace crops out downstream of this input, the coarse size fraction cannot be compared directly with that of the modern river. The influence of the Rosyntjieberg quartzite is evident in the Meso samples, where it constitutes a large proportion of the clasts in the coarse size fraction, compared to Nama quartzite in the modern river. In the -40mm size fractions, the Meso contains more granite and vein quartz than the modern river. Both deposits have similar proportions of Karoo sedimentary clasts, although these clasts are present in the coarse size fractions of the Meso deposits. Karoo dolerite clasts are present in the modern sample where they are absent in the Meso sample. Further to this, a strong Upper Orange River component in terms of Karoo basalt, zeolite and Orange River type agate clasts are all present in the modern river but are absent in the Meso. Both samples have similar proportions of BIF clasts, although it occurs in the +25 mm fraction in the Meso sample, where that fraction is absent in the modern river. In the finer fractions, there are roughly equal amounts of BIF.

The average clast roundness of the two samples is similar, remaining fairly consistent until a sudden drop-off in the +3 mm category.

6.5.1.3 Grasdrif/Aussenkjer Area

The deposits situated at Grasdrif/Aussenkjer (Figs.4.1 & Fig.6.10) rest on Karoo shales and sandstones of the Nabas Basin in Geomorphic Reach 3. A total of six samples were analysed from this area: 2 from the Proto terrace (Samples 15 & 16), 2 from the

Intermediate terraces (Samples 17 & 18), 1 from the Meso (Sample 19) and 1 from the modern river (Sample 69)(Fig.6.4a and 6.11a).

All samples are dominated by Rosyntjieberg quartzite in their coarse fractions, the proportion of which decreases in the finer fractions. Once again, the basic clast proportions are similar. The proportion of NMP basement clasts increases significantly in the modern river sample. The proportions of granite and feldspar increase in the Intermediate, Meso and modern samples compared to the Proto samples, whilst the modern sample has an unusually large amount of vein quartz in the +3 mm fraction. The proportions of Karoo sedimentary clasts in the samples are similar, apart from the Grasdrif Intermediate and Meso samples, which have lower proportions of this clast type. As with the Seven Pillars example, Karoo dolerite is only found in abundance in the modern river samples. The proportions of BIF clasts in the samples are similar, apart from the Grasdrif Intermediate sample which has the highest proportion.

The average clast roundness is haphazard, with no clear pattern developing through time. Surprisingly, some of the samples have higher average roundness estimates in the +40, +25 and +16 mm categories than in the coarser size fractions, e.g. the Grasdrif Meso sample (Fig.19, Appendix D). The high average roundness estimates in the +40 and +25 mm categories are caused by relatively large amounts of well-rounded granite, basement, mafic volcanics and vein quartz clasts in the +40 mm category and well-rounded mafic volcanics and BIF clasts in the +25 mm category (evident in the close up photo in figure 17, Appendix D).

In addition to the above, the sand fractions from the lower samples of each age group were analysed in the 2-3 mm and 1-2 mm categories at this locality (Fig.6.11b). These fractions are dominated by vein quartz and granite/feldspar. Presumably, much of the vein quartz is derived from the breakdown of granites. The Proto Lower and Intermediate samples have similar proportions of vein quartz, the proportion of which decreases from these samples to the Meso and is lowest in the modern river sample. Concomitantly, the proportion of granite/feldspar is highest in the younger samples, with the most lithics being present in the modern river sample.

In contrast to the larger clasts, the average clast roundness of the +2mm and +1mm fractions in the modern river sample (Fig.6.11b) is the highest. This is mainly due to the well rounded mafic volcanic lithics present in this sample.

6.5.1.4 Gamkab River area

The Gamkab River is a major tributary that drains mostly low relief ground underlain by rocks of the Karoo basin, apart from the lowermost reach within 5 km of the Orange River where it is incised into the NMP basement, exposing the NMP/Nama Group unconformity. A Meso terrace is present on the downstream side of this tributary (Fig.6.12) and 2 samples were analysed from this site, a lower (Sample 20) and upper sample (Sample 21)(Fig.6.4a and 6.13). The Aussenkjer modern sample (Sample 69) is used as a comparison in Fig.6.13 as it is situated between the Grasdrif and Gamkab sites.

The Meso lower and upper samples have very similar clast compositions at this site. In comparison to the modern sample, there is a higher proportion of Karoo sedimentary clasts, and BIF clasts are well represented in coarser size fractions than in the modern sample. The average clast roundness per size fraction does not show any appreciable differences between the samples.

6.5.1.5 Block 6 Area

Block 6 is situated in Geomorphic Reach 4 and Link 3, the link in which the MCS population is dominated by felsic volcanics (Fig.6.4a and 6.14). Meso-Orange River terraces are present at this locality, which are terminated at the downstream end by a large tributary input terrace. A total of 3 samples were analysed in this area: 2 from the Meso deposit (upper and lower, Samples 22 & 23) and 1 from the modern river (Sample 70). In all 3 samples the +300 mm category is dominated by felsic volcanics, although to a lessor degree in the modern river sample.

A number of trends are evident at the Block 6 site (Fig.6.15). The proportion of mafic volcanic clasts is lower in the modern river sample, in contrast to the higher proportion of granite and feldspar, in comparison to the Meso samples. The amount of Karoo sedimentary clasts in both Meso samples is similar, whereas there is a significant decrease of this clast type in the modern sample. The proportions of BIF and riebeckite decreases from the Lower to the Upper Meso sample, and a further significant decrease of this clast type occurs in the modern river sample. As with the previous samples, Karoo dolerite is present only in the modern sample, although not in significant amounts.

In terms of average clast roundness, there is a significant drop in roundness in the modern river and all samples display better rounding in their coarse fractions compared to the smaller sizes.

6.5.1.6 Fish River Area

This area is situated approximately 5 km downstream of the Fish River confluence with the Orange River where a Meso terrace with an undulatory bedrock crops out on the right bank (Fig.6.16). Two samples were analysed from the lower (Sample 26) and upper Meso gravels (Sample 25)(Fig.6.4a and 6.17). In the coarse size fractions, the upper sample has higher proportions of Nama quartzite, but more basement in the finer size fractions, whilst the lower sample has significantly higher proportions of mafic volcanic clasts in the +40 to +8 mm size range. The proportion of BIF clasts decreases significantly in the upper sample, although it does coarsen, appearing in the +40 mm category. Other significant differences are the proportions of Karoo basalt, zeolite and Orange River type agate clasts, which all increase in the younger upper sample. There are no significant differences in the average clast roundness per size fraction at this site.

6.5.1.7 Boom River Area

The Boom River area (Fig.6.18) is situated at the start of Geomorphic Reach 6 and Link 4, downstream of the Boom River tributary confluence. The Boom River has supplied large amounts of Nama quartzite clasts to the Orange River since at least Proto times, although it may have been active since the early stages of valley incision. Evidence of this is a Proto-level terrace deposit which contains no exotic clasts characteristic of the Vaal-Orange drainage, but only locally derived clasts. Six samples were analysed at this locality: 2 from the Proto (lower and upper, Samples 28 & 29), 2 from the Meso (lower and upper, Samples 31 & 32), 1 from the modern (Sample 71) and 1 from the Proto Boom Local terrace (Sample 30)(Figs.6.4a and 6.19).

The clast assemblage in the Boom Proto Local terrace, containing only tributary-derived material, is dominated by Nama quartzites in all size fractions, with subordinate amounts of basement, granite and limestone. When mixed with Orange River gravel, the Nama quartzite maintains its dominance in the coarser fractions, but shows a reduction in a characteristic "stepped profile" towards a more diverse assemblage in the finer size fractions. In the Meso Upper sample, this profile is considerably flatter, probably indicating that a great deal of the material present in this terrace is derived from the Boom River tributary. Limestone clasts, derived locally from the Nama Group are more prominent in both the Proto and Meso upper samples and are also present in the modern sample. The proportion of mafic volcanic clasts, especially when comparing all the "Lower" samples, decreases significantly in the Meso and modern samples. The proportion of basement clasts shows the opposite trend and increases significantly in the Meso and modern river samples.

In the fine size fractions, the proportion of Karoo sedimentary clasts is high in the Proto samples, and decreases significantly in the Meso and modern samples. The proportions of BIF and riebeckite are significantly higher in the Meso lower sample. Although only present in small amounts, Karoo basalt and zeolites are only present in the Meso and modern samples.

In terms of the average clast roundness, the Boom Proto Local sample is the least rounded, as is to be expected of this locally derived input. Interestingly, both the Proto and Meso upper samples are almost as poorly rounded in the coarse fractions as the Local sample, this and their composition indicating much more local supply of clasts. In the lower samples, the roundness decreases in time from the Proto to modern river.

6.5.1.6 Lorelei East Area

Lorelei East (Fig.6.20) is situated at the confluence of the Zebrasfontein River, another Nama quartzite-bearing tributary from the north that is at least of Proto equivalent age. The Zebrasfontein valley hosts a Proto level terrace with locally derived clasts that was also sampled. A total of six samples were analysed at this locality: 3 from the Proto suite (Pre-Proto, lower Proto and upper Proto, Samples 6, 33 & 35), 2 from the Meso (lower and upper, Samples 36 & 37) and a Proto Zebrasfontein Local (Sample 34)(Figs.6.4a and 6.21). A Pre-Proto 2 age was assigned to the one sample based on its deep position in a bedrock scour feature, which is approximately 35m below the elevation of the Proto strath.

As in the case of the Proto Boom River local terrace, the Proto Zebrasfontein local terrace is also dominated by Nama quartzite in all size fractions, although to a slightly lesser extent than the Boom locality. This clast type is accompanied by subordinate amounts of Nama limestone, basement, vein quartz, mafic volcanic and a soft schistose metasedimentary rock which does not appear in any of the Orange River samples.

In terms of locally derived clasts, the samples display similar proportions, apart from basement clasts which have the highest proportion in the Meso upper sample, the youngest sample analysed at this site. The Pre-Proto and Proto samples are very similar, with both hosting large proportions of Karoo sedimentary clasts. However, the Pre-Proto sample incorporates significantly more agate clasts, notably the resinous Vaal River type, as well as chert. The proportion of Karoo sedimentary clasts decreases significantly in the Proto Upper sample and further in the Meso samples. Orange River agates are most prevalent in the Proto upper and Meso lower samples, whilst Karoo basalt is only present in the Meso samples.

Apart from the very poorly rounded Zebrasfontein Proto Local sample, there is very little difference in the average clast roundness of the different samples and they show a similar trend through the size fractions.

6.5.1.7 Lorelei West Area

Lorelei West is situated at the downstream end of Geomorphic Reach 6, where the Orange River valley begins to widen into the coastal plain (Fig.6.22). A total of six samples were analysed here: 1 from the Proto (Sample 38), 2 from the Intermediate (lower and upper, Samples 39 & 40), one from the Meso (Sample 41) and 1 from the modern river (Sample 72)(Fig.6.4a and 6.23).

As in the previous localities, the Proto sample has significantly more Karoo sedimentary clasts than the other samples at this site. There are significantly higher proportions of BIF clasts in the Meso and Intermediate than in the Proto and Modern samples. Karoo basalt and zeolite clasts, absent in the Proto sample, make an appearance in the Intermediate and Meso samples, but increase substantially in the modern sample. The Intermediate and Meso assemblages are very similar at this locality.

The average clast roundness increases with age, with those in the Proto being the best rounded and those in the modern being the most angular. There is considerable similarity in roundness between the Intermediate and Meso samples.

6.5.1.8 Sendelingsdrif Area

The Sendelingsdrif area is situated at the start of Geomorphic Reach 7, at the start of the coastal plain (Fig.6.24). A total of 6 samples were analysed at this site: 1 from the small Pre-Proto 1 remnant known locally as Snake (Sample 5), 4 from the main body of Proto gravel (Samples 42-45), and 1 from the modern river deposit (Sample 73; Figs. 6.4a and 6.25). Unfortunately, due to the lack of a suitable trench, no sample could be taken from the Meso deposit.

The Pre-Proto 1 remnant at Snake, the highest and undoubtedly the oldest terrace remnant preserved in the Lower Orange, yielded a surprising result. In comparison to the Proto samples (Samples 42-45, Fig.6.25) a significantly higher proportion of BIF (usually characteristic of the younger Meso deposits) and Vaal River agate clasts are present in this sample, but the proportion of Karoo sedimentary clasts is less. The samples from the Proto deposits have very similar clast proportions. The difference between the Proto deposits and the modern river is striking. The proportion of Karoo sedimentary clasts is considerably less than in the Proto. Limestone and zeolite clasts are restricted to the modern sample only and basement clasts are also more prominent. BIF is restricted to the two finest size fractions in this modern sample.

There is a decrease in the average clast roundness from the older deposits to the modern river (Fig6.25).

6.5.1.9 Daberas Area

The Daberas area is situated in the meandering reach of the modern river (Fig.6.26). A total of 7 samples were analysed from this deposit: 5 from the Proto (3 lower, 2 upper) and 2 from the Meso (lower and upper)(Figs.6.4a and 6.27).

The clast assemblages of the Proto lower deposits (Samples 47, 49, 51) are very similar. A notable difference between these deposits are the higher proportions of Gariep metasedimentary clasts in Sample 47, which can be explained by its proximity to an active tributary. Differences between the Proto lower and upper samples (Samples 48, 50) are minor, where the upper samples have lower proportions of Karoo sedimentary clasts and BIF clasts are present in +40 mm size fraction. The Meso samples (Samples 52 & 53) have significantly lower proportions of Karoo sedimentary clasts than the Proto samples, but higher proportions of BIF and riebeckite clasts. In the Meso upper deposit (Sample 53), Orange River type agate and Karoo basalt clasts make an appearance. All of the Proto samples have chert clasts present, whilst this clast type is virtually absent in the Meso samples. There are few clear differences in the average clast roundness at this site, with the Meso samples being are more poorly rounded in the finer fractions.

6.5.1.10 Auchas Major Area

Auchas Major has a very complete record of deposits ranging from Pre-Proto 2 to modern (Fig.6.28). A total of 5 samples were analysed: 1 from the Pre-Proto, 2 from the Proto, 1 from the Meso and 1 from the modern river (Figs.6.4a and 6.29a).

The AM11 Pre-Proto 2 sample (Sample 9) was taken at the base of a pre-Proto 2 scour, approximately 25m below strath level. As with the sample at Dreigat, it has abundant vein quartz and metasedimentary clasts, representative of the scour environment in which it was deposited. In comparison with the Proto deposits which are deposited above strath level, the Pre-Proto sample has higher proportions of Vaal River type agate and vein quartz clasts, but lower proportions of Karoo sedimentary clasts. The Proto lower and upper samples (Samples 54 and 55) have a very similar clast assemblage. In comparison to the Proto samples, the Meso sample (Sample 57) has lower proportions of Karoo sedimentary clasts, an almost complete absence of agate clasts, higher proportions of BIF clasts, as well as the appearance of riebeckite clasts. The gravel in the modern river deposit, at this locality, has fined considerably, as discussed in Section 5.7.2 and no clasts are present in the +300 mm category. Those present in the +200 mm category are all flat rafts of locally derived Gariep metasediments. In comparison to the older deposits, there are lower proportions of Karoo sedimentary, BIF and agate clasts. As at previous localities, limestone is present in the Modern River gravels but in none of the other samples and zeolites are in relative abundance in the +3 mm size fraction. At this locality, therefore, agate content decreases with a younging of the deposits whereas Karoo sedimentary clasts are most abundant in Proto times and BIF and riebeckite in Meso times.

In addition to the above, the sand fractions from the lower samples of each age group were analysed in the +2-3mm and +1-2mm categories (Fig.6.29b). Although the proportions of granite and feldspar in the +3mm fraction of the Pre-Proto sample is low, its +1 and +2mm fractions have the highest proportions, whereas the modern sample, which has the highest proportion of granite and feldspar in the +3mm fraction, has the lowest proportions in the finer categories. In addition, zeolite is present in the fine fractions of the modern sample and absent in older samples. The results from the +1 and +2 mm fractions from this site are in contrast to those from Grasdrif (Section 6.5.1.3), which displayed an increase in the proportion of granite and feldspar in the modern river sample.

The average clast roundness decreases with the age of the deposits, reaching a minimum in the modern river. In contrast, the sand fraction (1-3mm), as at the Grasdrif locality, is best rounded in the modern river.

6.5.1.11 Auchas Lower Area

Auchas Lower is a small Proto deposit flanked by a larger Meso deposit (Fig.6.30). A total of 3 samples were analysed: 2 from the Proto (lower and upper) and 1 from the Meso (lower) (Fig.6.4a and 6.31).

The Proto lower and upper samples (Samples 58 & 59) have similar clast assemblages, apart from the proportion of Karoo sedimentary clasts, which is higher in the upper sample. In comparison to the Proto samples, the Meso sample has higher proportions of basement and BIF clasts,but significantly lower proportions of Karoo sedimentary clasts. The agate clasts present in the Proto Meso samples are dominated by the Vaal and Orange River types respectively. Karoo basalt and zeolite make an appearance in the Meso sample.

The average clast roundness of the Proto samples is similar, but decreases in the Meso sample.

6.5.1.12 Arrisdrif Area

Arrisdrif is situated on the Grootderm metavolcanics, and consists of a Proto deposit, flanked by 2 levels of Meso terraces (Fig.6.32). A total of 5 samples were analysed at this site: 2 from the Proto (lower and upper), 2 from the Meso (both are lower samples from the Upper Meso terrace and Lower Meso terrace) and 1 from the modern river (Figs.6.4a and 6.33).

The Proto lower sample (Sample 61) is dominated by a combination of Nama quartzite and mafic volcanics, the latter derived from the Grootderm volcanics. This is due to the trench from which the sample was taken being situated close to a palaeo-tributary input point. The remainder of the samples are dominated by Nama quartzite clasts in the coarse fraction, apart from the modern river sample which is considerably finer grained, with a more diverse coarse assemblage. The Proto lower and upper samples (Samples 61 and 62) differ in that the proportions of BIF and basement clasts are higher, whereas the proportion of mafic volcanic clasts are lower in the upper sample. The Proto lower sample is the only sample at this site lacking both Karoo basalt and zeolite clasts. The Meso (Sample 64) and Proto upper sample, which at previous sites had differing proportions of BIF, are very similar at this site, the difference being that there are lower proportions of Karoo sedimentary clasts in the Meso sample. Agates change from being dominated by Vaal River types in the Proto samples, to Orange River types in the Meso and modern samples.

Apart from the poorly rounded +90mm fraction of the modern sample due to a local input of angular metasedimentary slabs, the average clast roundness of the clasts between samples does not differ much.

6.5.1.13 Oranjemund Area

The Oranjemund area is situated at the transition between the fluvial and marine deposits (Fig.6.34). A total of 4 samples were analysed from this locality: 1 is from a Meso fluvial terrace, 1 from a Pleistocene beach gravel, 1 from a suspected Eocene fluvial deposit, and 1 from a Pleistocene beach adjacent to the suspected Eocene deposit (Fig.6.4a and 6.35).

The Pleistocene beach deposit (Sample 66) has significantly higher proportion of quartzite and BIF clasts in comparison to the Oranjemund Meso deposit (Sample 65), but less basement, volcanic and Karoo sedimentary clasts. An upgrade in the amount of quartzite at the expense of the more labile clasts is what would be expected from the river to beach sample. As expected, the average clast roundness in the beach sample is significantly higher than the fluvial sample.

Interestingly, where most of the fluvial samples analysed during this study have higher values of roundness in the coarser fractions, the beach shows higher roundness values in the +40 mm small/medium cobble fraction. A similar pattern of clast roundness with clast size was obtained in beach gravels by Bluck (1969b).

The so-called Eocene MA1 sample is thought to be of this antiquity due to its similarity in clast assemblage to the dated Eocene marine deposits further north in the Sperrgebiet. It is inferred to be of fluvial origin, based on limited sedimentary structures of non-marine origin and the complete absence of marine shells. The clast assemblage difference between this deposit and the surrounding Pleistocene beaches are striking. Firstly, the coarse quartzite population is distinctly different and at a first glance appears to consist of only pure white quartzites. On closer inspection, however, once broken, some of the quartzite clasts do consist of dark grey or brown Nama quartzites, but they have a white "weathering rind" where the iron staining responsible for the dark colour is absent. None of these weathering rinds were observed in the younger deposits and they may also be indicating an old age for the deposit. Apart from these clasts, most of the clasts do consist of white quartzites which are sourced from the Rosyntjieberg quartzites and the upper units of the Nama quartzites.

Apart from the quartzites, this sample has a radically different clast assemblage to the younger samples, with significantly higher proportions of vein quartz, agate and chert clasts, but lower proportions of BIF. The agates are mostly the dark, resinous Vaal River type. A sample taken approximately 3 m away in the directly adjacent Pleistocene beach, yielded a different clast assemblage, dominated more in the coarse fractions by the dark, Nama quartzites, from lower in the Nama stratigraphy. In addition, proportions of NMP lithology clasts are higher, but scavenging from the older deposit has occurred as the vein quartz and Vaal River agate proportions are significantly higher than in the younger Sample 66.

6.5.1.14 Buntfeldschuh/Bogenfels Area

The Buntfeldschuh/Bogenfels area is situated approximately 150-180 km to the north of the Orange River mouth in what has been referred to as the "Trough Namib" (Beetz 1926; Stocken 1978) (Fig.6.4c). In this area, there are numerous high-level Eocene marine deposits, some of which have been dated by their well preserved assemblage of micro-and nanofossils to an Upper-Eocene age (Siesser & Salmon 1979). Three samples were analysed in this area (Fig.6.36). The Buntfeldschuh deposit, a 100 m thick succession of marine and aeolian sediments is exposed in an escarpment face. Although not dated directly, it has traditionally been correlated with the Eocene deposits on the similarity of the clast assemblage to those at the dated localities (Stocken 1978). The clast assemblage of the Bundfeldschuh (Sample 4) and the Bogenfels B10 (Sample 3) deposits is very similar. Both are dominated by Vaal River agate, (a large proportion of which is clear yellow chalcedony), vein quartz and chert clasts with minor amounts of quartzite, volcanics, basement and granite being present. The lithology represented in the "Other" category in both samples is green chrysoprase, which is only known to be present in the Lichtenburg deposits (Armstrong *et al.* 1987), and thus has a Vaal River source. Comparison of these samples with the Granitberg sample, also of Eocene age, reveals a similar clast assemblage, but the Granitberg sample (Sample 2) has higher proportions of BIF clasts, but lower proportions of Vaal River type agate clasts.

In terms of average clast roundness (Fig.6.36), the clasts in the Granitberg deposit are the most rounded whereas Buntfeldschuh and the Bogenfels B10 sample have very similar roundness.

6.5.1.15 Augrabies Falls Area

As outlined in Section 4.5.4, there are a number of alluvial deposits flanking Augrabies Falls (Fig.6.4b). Renosterkop is situated approximately 15km upstream of Augrabies Falls and hosts an upper and lower terrace that, on the basis of relative height above river level, have been tentatively correlated with the Proto and Meso deposits in the study reach (Ward *et al.* 2002). Downstream of Augrabies Falls is the Daberas deposit and the Stolzenfels kimberlite with its gravel infilling. Samples were taken from each of these localities (Fig.6.37). Although not one locality, this area will be treated as an entity.

The Renosterkop Proto deposit comprises gravel-filled potholes situated approximately 40-45 m above modern river level, that were mined for diamonds in the 1930's. One of the larger potholes was not mined to completion and was sampled. The Meso terrace consists of a more laterally extensive deposit of gravel-filled potholes on a fairly wide strath. The +90mm clast assemblage of both Renosterkop samples is dominated by NMP basement. The finer fractions of the Proto deposit (Sample 10) are dominated by BIF, with subordinate amounts of vein quartz, mafic volcanics (from the Brulpan Group and Ventersdorp Supergroup) and quartzite, probably from the Olifantshoek Supergroup. Orange River type agate is present in the +3mm fraction. The "Other" category in the +3mm fraction is represented by metamorphic garnet. The Meso deposit (Sample 73), also with a high proportion of BIF clasts has significantly less BIF than the Proto deposit, and has a more diverse clast assemblage including higher proportions of granite, NMP basement, vein quartz and Karoo sedimentary clasts. In addition, riebeckite and Karoo basalt clasts are present in the Meso, but absent in the Proto.

The Daberas locality comprises 3 small gravel-filled potholes approximately 50 km downstream of Augrabies Falls, and has been correlated with the Proto deposits on the basis of relative height above modern river level (ca. 100 m). Only gravel from the upstream pothole was analysed. In comparison to the Renosterkop Proto (Sample 10), the Daberas sample (Sample 77) has significantly more BIF, with the +25, 16 and 8 mm fractions composed entirely of BIF. Not only is the amount of BIF in this sample striking, but also the amount of nearly perfectly spherical BIF clasts (Fig.77, Appendix D), which is unusual, as most BIF occurs as flat, elliptical pebbles. In addition to their shape, the average clast roundness of the Daberas sample is far higher than the Renosterkop samples, although not in the +3mm fraction.

As outlined in Section 3.3.3.4, depending on the relative age of the Renosterkop and Daberas deposits, there must have been either a primitive Augrabies Falls or at least a very steep reach between the two deposits in Proto or Pre-Proto times. A turbulent reach, extending for this ca. 60 km would explain the better rounded and spherical BIF clasts, and relatively higher compositional maturity of the Daberas gravels in comparison to Renosterkop.

The Stolzenfels kimberlite is situated approximately 85 km downstream of Augrabies Falls and, as explained in Section 4.5.4 and Fig.4.19, is at higher level than any other deposit and on that basis is more than likely the oldest preserved inland deposit in this area. At the time of eruption (ca.60 Ma), this area had a cover of Karoo sediments as evidenced by the numerous Karoo xenoliths present in the pipe. The clast assemblage of the Stolzenfels deposit (Sample 78) clearly differs from other deposits in this area; although BIF clasts are present and well represented, the +25 and +16 mm categories are dominated by Karoo sedimentary and hornfels clasts, whereas the +8 and +3 mm categories are dominated by agate clasts, mostly of the Vaal River type. Chert is an important clast present in all size fractions, and vein quartz, feldspar, volcanic and quartzite clasts are minor components. The "Other" category is represented by clasts of rounded kimberlite and green crysoprase. This sample is the most poorly rounded of the deposits in this area (Fig.6.37). This assemblage is more similar to the Eocene samples found on the coast than to the Renosterkop or Daberas samples.

6.5.2 Summary - Clast Assemblages per Site

Proportions of clast types, as dealt with here, are difficult to analyse in that it is a "relative" dataset i.e. variations of a certain clast type, that are dependant on changes in other clast types.

Overall, the clast assemblages are evolving from a very exotic-rich, siliceous assemblage of Vaal River agate, vein quartz, chert and BIF clasts in the Eocene deposits, to being dominated by clasts which are locally-derived within the study area (Fig.6.38A-E). By Pre-Proto time, the Orange was well-incised into the Richtersveld, as evidenced by the locally-derived assemblage of the Pre-Proto deposits (Fig.6.38B). The proportion of Vaal River type agate is high in the Pre-Proto deposits in comparison with the younger deposits. The Proto deposits (Fig.6.38C) are generally characterised by a high proportion of Karoo sedimentary clasts, whereas the Meso deposits (Fig.6.38D) have high proportions of BIF clasts, accompanied by blue riebeckite. The modern deposits (Fig.6.38E) are characterised by low proportions of the exotic suite of clasts, namely BIF, Karoo sedimentary and agate clasts, although, along with the Meso deposits, generally contain Karoo basalt and zeolite clasts, both of which are generally absent in the older deposits. The proportion of Vaal River to Orange River type agate is highest in the older deposits and decreases in the younger deposits.

Although differences are not present at each site, average overall clast roundness per size fraction tends to decease from Eocene through to the modern deposits (Fig.6.38F), indicating that the Orange River is less able to deal with the sediment supplied to it in the younger deposits.

The presence of both limestone and Karoo dolerite clasts in the modern deposits and virtual absence of these clasts in the older deposits, as well as the higher proportions of metasedimentary clasts in the modern samples on the coastal plain, indicates that the competency of the modern river is less than that of the ancient river. Both of these clast types are fairly soft and easily eliminated in rivers (Plumley 1948; Sneed & Folk 1957; Abbott & Peterson 1978)) and their absence in the older deposits indicates higher rates of clast abrasion. McCarthy (1982) also noted an increase of limestone/dolomite in his younger samples and ascribed it to lack of preservation in the older deposits. However the fact that limestone is abundant and well preserved in the 2 local terraces of at least Proto age which were sampled (Boom and Zebrasfontein Local samples), indicates that it is able to be preserved in the deposits for this amount of time. Thus the presence of these more labile lithologies, together with the general lack of clast roundness, indicate a loss of river competancy in the modern river.

Many of the upper samples (samples taken well above the bedrock contact) contain a larger proportion of locally derived clasts, with poorly rounded coarse fractions. These data add weight to the argument that the aggradation of the terraces is due to increased input of material from the tributaries. Also, the fact that limestone is only present in the upper samples of the palaeoriver supports the view that abrasion is more important in a degrading river and is less effective in an aggrading river (Pettijohn 1957; Bradley *et al.* 1972). This is due to the fact that in a downcutting river, clasts are subject to collisions with other clasts for longer periods of time, compared with clasts in aggrading rivers, which are buried more quickly.

However, the overall average clast assemblages per suite (Fig.6.38) mask changes within each suite that occur along the channel, thus each suite needs to be analysed separately.

6.5.3 Clast Assemblage Changes Within each Suite

6.5.3.1 Eocene Suite

The deposits having a siliceous "Eocene-type" clast assemblage comprising mainly exotics (Fig.6.39) are found at both the Augrabies and coastal extremities of the river system, as well as 200 km farther north along the coast. Although the absolute age for some of the deposits is known, the relative age of these deposits is unknown.

As discussed in Section 6.5.1.15, the Stolzenfels deposit is clearly older than the remainder of the deposits in the Augrabies Falls area, by virtue of its height above modern river level and agate, chert-dominated clast assemblage. However, it is not known how the Stolzenfels sample correlates with the coastal Eocene deposits. The large amount of Karoo sedimentary clasts present in the Stolzenfels deposit (Fig.6.39A) could be interpreted in a number of ways: 1. An older age of the Stolzenfels deposit than the coastal deposits could be indicated by virtue of the fact that more Karoo rocks were present in the area at an early stage, thus incision into the landscape had not progressed much. 2. The relative lack of Karoo sedimentary clasts in the coastal Eocene deposits could be explained by abrasion and the destruction of these clasts in the 500 km journey to the coast, thus, a co-eval age of deposits could be indicated. 3. The large amount of Karoo sedimentary clasts present in the Stolzenfels deposit could represent the slug of Karoo that dominates the Proto deposits between the Boom River and Oranjemund, thus Stolzenfels could be younger than the coastal Eocene deposits, although this is unlikely given its relative height above modern river level.

In the Eocene MA1 sample (Fig.6.390B), coarse quartzite and basement clasts with origins in the study area must indicate that substantial incision into the Richtersveld had already occurred at this time. However, the relative lack of volcanics and NMP basement clasts in this sample indicates either that full incision had not yet occurred, or a more vigorous river with considerably more abrasion producing a quartzite, vein quartz, agate and chert rich assemblage was in existence. This may indicate a river in the initial stages of incision, in a steep gorge, before a system of lateral tributaries could incise and expand to choke up the river with locally-derived clasts. Another factor is that the breakdown of locally derived clasts would be enhanced by the more humid than present climate during the Eocene (Partridge 1993). However as the quartzites are mostly from the Rosyntjieberg Group and upper Nama Supergroup, this supports the case for a less advanced incision at this time. The exact provenance of the vein quartz in this sample is uncertain, and is probably either derived from the NMP between Augrabies Falls and Noordoewer, or from incision into the Gariep Supergroup.

The Buntfeldschuh (Fig.6.39C) and Bogenfels B10 sample (Fig.6.39D) would appear to be older than the Eocene MA1 sample by virtue of the larger amounts of agate and less Rosyntjieberg quartzite and vein quartz. The orogin of the Nama quartzite in these samples is uncertain as it could have been derived locally, by the scavenging of the Cretaceous Blaubok gravels (Stocken 1978) or could have been sourced from the Richtersveld. Rosyntjieberg quartzite is known to occur only in the Richtersveld, hence the lack of this quartzite could indicate less incision into the Richtersveld, hence that these samples are older. However, the abrasional effect on the clast assemblage that 150-180 km of transport up the coast has is not known. Plumley (1948) found chert to be the most resistant lithology to abrasion, more so than vein quartz and quartzite, thus, the difference in assemblages could be explained by breakdown of quartzite and vein quartz and the survival of agate, which probably has similar abrasional properties to chert.

The Granitberg sample (Fig.6.39E) is different to the other Eocene-type gravels in that it has a substantial volume of BIF. In this respect it is more compatible with the Pre-Proto 1 sample from Snake (Fig.5, Appendix D), which post-dates the incision of the Richtersveld, but also has higher concentrations of BIF than agate. Consequently, these two samples indicate a post-Eocene early slug of BIF.

As expected, the average clast average roundness increases from the fluvial samples to the marine/aeolian samples (Fig.6.39F).

6.5.3.2 Pre-Proto Deposits

The only example of a Pre-Proto 1 deposit is the Snake deposit (Fig.6.40A) at Sendelingsdrif. This key outcrop records the clast assemblage of the river at the earliest known time in the development of the Richtersveld landscape, the clasts of which are well represented in comparison to the exotic-dominated Eocene samples. The Snake deposit represents the transition, in the -40 mm clasts, from the Vaal River type agate dominated assemblage to a Karoo sedimentary dominated one. In addition, this deposit records an earlier slug of BIF clasts previously thought to have occurred only in Meso times (Fowler 1976; van Wyk & Pienaar 1986; Ward *et al.* 2002).

The younger Pre-Proto 2 deposits, Lorelei East (Fig.6.40B) and Auchas Major (Fig.6.40C), both of which are preserved in deep scour environments, have very different clast assemblages, with both having lower proportions of BIF than the Snake deposit. The Lorelei East deposit displays a very similar assemblage to the Proto deposits, with high proportions of Karoo sedimentary clasts, whereas the Auchas Major deposit has proportionately fewer Karoo sedimentary clasts, but more Vaal River type agate clasts. On the basis of these two clast types, with Vaal River type agate representing Eocene-type gravel and Karoo sedimentary clasts representing Proto-type gravel, the Auchas Major deposit is an older deposit than the Lorelei East deposit.

The average clast roundness of the Pre-Proto deposits is very similar (Fig.6.40D). The poorly rounded coarse fraction of the Lorelei Pre-Proto deposit can be explained by its location at a tributary input point and the more poorly rounded fine fractions of the Auchas AM11 sample can be explained by the higher proportions of more angular vein quartz and metasedimentary clasts present in this sample.

6.5.3.3 Proto Deposits

The Proto deposits are well represented in the Augrabies area and in the study reach. All of the Proto lower sample clast assemblages are plotted in sequence from upstream to downstream (Figs.6.41). The Renosterkop Proto deposit (Fig.6.41A) is almost completely dominated by BIF clasts. Comparison with the Daberas Upstream deposit (Fig.6.41B) indicates even more dominance by BIF, at the expense of other clast types. Thus during the time interval when these gravels were deposited, the clast assemblage passing the palaeo-Augrabies Falls was upgraded to a more mature, intensely rounded assemblage (Fig.6.37). By the time Noordoewer (Fig.6.41C) is reached, the BIF slug had been diluted to a large extent by the many tributary inputs of gravel en route. The presence of riebeckite in the Proto deposit at Noordoewer and the absence of it in the Augrabies Proto deposits is, however, problematic. Although the amounts of riebeckite involved are statistically insignificant, this absence in the upstream samples could have important implications. If the Proto deposits from these two areas are the same age, riebeckite would already have been present in the Augrabies deposits. The first flush of BIF that is present in both Eocene and Snake Pre-Proto 1 deposits have an assemblage of BIF that excludes riebeckite, whereas riebeckite is ubiquitous in the younger Meso deposits. Although a larger, more significant sample will be needed to prove this conclusively, it is suspected that the Renosterkop and Daberas upstream Proto deposits are older than the Proto deposits in the study reach, and possibly correlate with the Pre-Proto deposits. The riebeckite present in the Proto deposit at Noordoewer is the initial arrival of a younger slug of BIF that is very well represented in Meso times, which is accompanied by riebeckite. The Renosterkop Meso deposit, which contains riebeckite in its BIF population, has the potential to be correlated with either the Proto or Meso deposit at Noordoewer.

Although the Proto clast assemblage at Noordoewer represents a very dilute BIF population compared with the Augrabies Falls deposits, there is nonetheless a great deal of BIF present in the Noordoewer Proto deposit. In the study reach, the proportion of BIF declines downstream of Noordoewer. The proportion of Karoo sedimentary clasts increases to a maximum at the Boom River deposit (Fig.6.41E), whilst Vaal River type agate becomes more prominent from Sendelingsdrif downstream (Fig.6.41H-L). Based on these 3 clast type proportions, the Proto deposits between Noordoewer and Arrisdrif can be subdivided into 3 distinct reaches, an upstream, middle and downstream reach.

The upstream reach incorporates Noordoewer (C) and Grasdrif (D). These deposits are characterised by high proportions of BIF, lower proportions of Karoo sedimentary clasts and the presence of riebeckite and Karoo basalt. These two deposits have very similar assemblages to typical Meso deposits in the study reach. The middle reach, from the Boom (E) to Lorelei West (G), is dominated by Karoo sedimentary clasts but with very little BIF or agate. The downstream reach, from Sendelingsdrif (H) to Arrisdrif (L), has higher proportions of Vaal River agate and chert, but lower proportions of Karoo sedimentary clasts than the middle reach.

Thus, the longitudinal changes in the Proto deposits demonstrate a transition from a Meso-like assemblage in the upstream reaches, to a Vaal River agate and chert-rich suite in the downstream reaches, with the peak of the slug of Karoo sedimentary clasts being captured in the middle reach. These subdivisions in the clast assemblages co-incide with breaks in the Proto bedrock strath longitudinal profile (Fig.6.41M, Section 4.5.2). There is a substantial knickpoint in the profile between Grasdrif (D) and the Boom River (E), and there is a knickpoint between Lorelei West (G) and Sendelingsdrif (H). The downstream knickpoint between Auchas Lower (K) and Arrisdrif (L) does not show a significant change in the clast assemblage, although coarser clasts of BIF are present in the Arrisdrif deposit, but absent in the other coastal plain deposits.

Average clast roundness (Fig.6.41N) changes little downstream, with the obvious exceptions the Grasdrif and Boom River deposits. Both of these are explained by their proximity to the source of the Rosyntjieberg and Nama quartzites respectively. These deposits also have the least rounded clasts in the +3mm category (amongst the Proto deposits), which is probably due to the presence of more quartzite which would result in more intense abrasion. However, apart from these deposits, those most downstream deposits from Sendelingsdrif to Arrisdrif (colour-coded in blue) have the most poorly rounded fine fraction. Once again, this could be related to clast breakages yielding a greater fine fraction of angular material. The lack of a clear pattern in the remainder of the roundness data may indicate the complication of clast addition, as well as continual breakdown and rounding of clasts.

6.5.3.4 Meso Deposits

The Meso deposits are the best represented terrace suite in the study reach and particularly so in the incised Richtersveld reach where most of the Proto deposits have been removed by erosion or were never deposited. All of the Meso lower deposits are plotted in sequence from Renosterkop to Oranjemund in Figs.6.42A-Q.

As discussed in the previous section, the Renosterkop Meso deposit (Fig.6.42A) could be correlated with either the Proto or Meso deposits at Noordoewer. In common with the modern river MCS population, the coarse fractions of the Meso deposits change in response to the local tributary between Noordoewer and the Boom River locality. Downstream of the Boom River, Nama quartzite dominates to the mouth. The effects of abrasion are demonstrated from the Boom River to the mouth (Fig.6.42J-Q), where the amount of Nama quartzite increases to a maximum in the Oranjemund Meso deposit, at the expense of the more labile clasts.

From Noordoewer downstream, the amount of BIF generally increases to a maximum between Block 6 (Fig.6.42F) and Dabimub (Fig.6.42I), after which it decreases downstream. Karoo sedimentary clasts, are better represented in the deposits that are situated downstream from tributaries draining the Karoo-filled Nabas basin. Examples of this are the deposits of Seven Pillars (Fig.6.42C), downstream of the Sambok tributary, and those of Gamkab (Fig.6.42E), downstream of the Gamkab tributary, both of which drain the Nabas basin. The Auchas Major (Fig.6.42N) Meso deposit has an unusually large amount of Karoo sedimentary clasts for its position downstream, and it is suspected that a substantial proportion of this was scavenged from the adjacent Proto gravel, which outcrops very close to the Meso trench that was sampled. To test this, another trench will need to be excavated and sampled well away from the Proto deposit.

Most of the Meso deposits contain riebeckite, Karoo basalt, zeolite and Orange River type agates in their assemblage and, unlike the Proto deposits, there are far fewer longitudinal (downstream) changes in the deposits. This may be the result of the long time gap between the deposition of the Proto and Meso deposits. If the Meso signature of clasts was already present in the upstream reach of the Proto deposits by 17.5-19 Ma, then it is quite concievable that they would have been fully distributed throughout the study reach by the time the Meso was deposited at 2-5 Ma. However, the deposit furthest downstream, the Oranjemund Meso, has an anomalous clast assemblage for a terrace of this age in that the Orange River signature is completely absent, and only Vaal River agates are present. The only characteristic of a Meso clast assemblage is the presence of riebeckite. Thus, two possibilities exist for this deposit: either it is older than the Meso, or this is the first indication that the Meso is "ageing in a downstream direction", as the Proto does.

As with the Proto deposits, no downstream pattern appears to be present in the average clast roundness data (Fig.6.42R).

6.5.3.5 Modern River Deposits

The modern river deposits (Fig.6.43A-H) have a coarse clast assemblage change in the downstream direction as each sedimentary Link is encountered. In the Stormberg Modern sample (C) which is situated at the beginning of Link 3, the felsic volcanics only dominate the very coarse fraction (+300mm), after which a combination of quartzite and basement comes to the fore in the coarse to mid-size ranges, followed by granite and feldspar dominance in the fine sizes. The basement slug is most prominent in the modern river samples and can be clearly seen to increase towards the geomorphic Reach 6 samples (Fig.6.43C, D & E), after which it fines and declines into the Reach 7 samples (Fig.6.43F & G).

BIF proportions peak in the upstream samples (Fig.6.43A & B), after which they remain at fairly low concentrations. Proportions of Karoo sedimentary clasts also peak in the upstream samples proximal to the Nabas basin, after which they decline in the downstream direction. All of the modern samples have a strong Orange River signature in the form of Orange River type agates, Karoo basalt and zeolite. As with the Proto and Meso deposits, there are no easily discernable downstream trends in the average clast roundness (Fig.6.43H), apart from the poorly rounded coarse fractions of the downstream-most samples, due to the concentration of poorly rounded metasedimentary clasts in these samples. In addition to this, the Stormberg sample at the beginning of Link 3 is on average less well rounded than the other samples, which can be explained by its position in a reach with the highest amount of tributary bars per river km (Table 3.1).

6.5.3.6 Study area Summary

The data contained in figures 6.41-6.43 for the study area is most effectively summarised into an upstream, mid and downstream reach, corresponding with Noordoewer to Boom River, Boom to Sendelingsdrif and Sendelingsdrif to the mouth respectively. The "lower" samples in each of these reaches were combined, and the averages plotted (Fig.6.44). This combination makes the effective sample larger and more significant. The downstream changes in the Proto clast assemblage are evident (Fig.6.44A-C), with BIF clasts dominating the upstream reach, Karoo sedimentary clasts the mid reach and Vaal River agate clasts becoming more prominent in the downstream reach. The Meso deposits (Fig.6.44D-F) show less significant longitudinal changes and the modern deposits (Fig.6.44G-I) display decreases in Karoo sedimentary and BIF clasts in the mid and downstream reaches.

Comparing the deposits per reach, the upstream reach (Fig.6.44A,D,G) there are no significant differences between the Proto, Meso and modern deposits, apart from the presence of dolerite in the modern samples. In the mid and downstream reaches, there are significant differences between the 3 deposits, with the Proto being dominated by Karoo sedimentary clasts, Meso by BIF clasts and the modern containing fewer exotic clasts generally than the older deposits, apart from zeolite. In addition the modern river contains higher proportions of the more labile limestone and metasedimentary clasts in these reaches.

6.5.3.7 Downstream changes in maturity and roundness per suite

Due to the large additions of clasts from tributary input, as well as the continual breakdown of clasts by abrasion, it is difficult to analyse the downstream changes within the local suite of clasts for each terrace suite. To simplify matters, a maturity index, comprising the ratio of quartzite and vein quartz, to the softer, more labile locally-derived clasts was calculated for each sample per size fraction (Fig.6.45). A high value indicates a mature assemblage.

There is a general tendency in all grain sizes towards a more mature population downstream (Fig.6.45). However, the scale of the quartz/labile ratio increases with clast size, with values of 5 to 30 being common in the coarse size fractions (>90 mm), and values less than 2 being common in the finer grain sizes (<40 mm). This difference in scale is reflecting the general abundance of quartzite in the coarse grain sizes, as well as the loss of the more labile clast types into the finer grain sizes as distance along the channel proceeds. The ratio generally increases after the Boom River locality, where large amounts of Nama quartzite are added to the system. Comparisons between the gravel suites reveals a fairly variable dataset, but in the +90 mm fraction (Fig.6.45), linear trends fitted to the data are similar for the Proto and Meso, but much flatter for the modern. Although this coefficient of determination (r^2) is not significant, the +90 mm data indicates that the modern assemblage is more mature in the upstream reaches, but less mature in the downstream reaches, as noted in the MCS data (Section 5.6).

Due to the lack of any clear patterns in the overall roundness data, the roundness of only Nama quartzite vs distance downstream for each size fraction was plotted for the reach between the Boom River and the mouth (Fig.6.46). The coarse fractions display increases in roundness downstream. The rate of rounding is initially high in the first 10-odd km, as evidenced in the Proto and Meso suites for the +200 and +90 mm fractions (Fig.6.46). After this initial increase, the roundness stays constant or increases at a slower rate downstream, as evidenced in the MCS dataset (Section 5) and reported by earlier workers (Plumley 1948; Sneed & Folk 1957). The roundness of the fine fractions decreases downstream in the Proto and Meso suites, but increases in the modern suite. This may be reflecting the relative lack of coarse clasts in these downstream deposits in the modern system, thus there is not as much abrasion or clast breakage occurring in comparison with the older deposits. Similar plots were done with other lithologies, but the results showed no clear patterns, probably reflecting the more rapid rounding and breakdown of these clasts in comparison with the more resistant Nama quartzite clasts.

6.5.4 Summary - Clast Assemblage Changes Within each Suite

Between Noordoewer and Oranjemund, the gravel assemblage becomes more mature downstream with respect to quartzite and vein quartz content in the Proto, Meso and modern suites, although to a lessor extent in the modern system (Fig.6.45). There are no clear downstream patterns in the overall clast roundness dataset, which is complicated by the continual input and breakdown of clasts. When analysing only the Nama quartzite population downstream of the Boom River, the coarse fractions of the Proto and Meso suites increase in roundness downstream but the fine fractions decrease, as opposed to the fine fractions of the modern river which increase in roundness downstream. This is most likely due to the higher amounts of clast breakage in the more vigorous older rivers which host coarser framework clasts than the finer modern gravel, and may also be reflecting the difference between the older bedrock-based channel and the modern sediment-based channel.

The changes occurring in the exotic (clasts derived from upstream of the study area) lithologies in each suite of gravel are summarised schematically in figure 6.47. The Eocene suite is dominated by clasts of Vaal River agate, Karoo sedimentary, BIF and vein quartz at the Stolzenfels deposit. Relative to this deposit, the coastal Eocene deposits have higher proportions of Vaal River agate and vein quartz, with slightly less BIF and a great deal less Karoo sedimentary clasts. The age of the Stolzenfels deposit relative to those at the coast is unknown, so reasons for the difference in clast assemblages could be age-related or process-related changes along the channel, as discussed in Section 6.5.3.1.

The Pre-Proto 1 suite of gravels, represented by the Snake deposit in the Richtersveld, is correlated with the Renosterkop and Daberas Proto deposits in the Augrabies Falls area, as well as the Granitberg deposit on the coast, based on the high BIF content and relative lack of riebeckite. This Pre-Proto time marks the transition from a Vaal River type agate dominated assemblage to one with roughly equal amounts of BIF, Karoo sedimentary and Vaal River type agate clasts. The Snake deposit is a key locality as it records the first slug of BIF, as well as the start of the slug of Karoo sedimentary clasts. The Pre-Proto 2 deposits are small, locally preserved scours that display a mixture of Pre-Proto 1 and Proto clast assemblages.

The Proto deposits display an array of changes along the channel in the study area (Fig.6.44 & 6.47), reflecting the changes in clast assemblages through time. The upstream reach has a very similar assemblage to the Meso deposits, and the downstream deposits become more Vaal River type agate-rich, indicative of the older suite of deposits. This downstream change in the Proto clast assemblage is either representing the downstream advance of the Meso suite of clasts through the study reach, or that the Proto deposits in the upstream reach at Noordoewer and Grasdrif area cannot be correlated with the downstream deposits and are younger deposits. The former interpretation is preferred.

The Meso deposits have very consistent clast assemblages along the channel, and all are characterised by large amounts of BIF clasts, accompanied by riebeckite, as well as the Orange River signature of Karoo basalt and zeolite clasts. The slug of BIF clasts, which, in Proto times had the highest proportion at Noordoewer, had moved further downstream and reached a maximum in the Fish River area (Fig.6.47).

The exotic suite of gravel is very poorly represented in the modern river gravels, apart from Karoo basalt and zeolite clasts, which increases in proportion in comparison to the Meso deposits.

6.5.5 Detailed BIF Analysis

6.5.5.1 BIF Types

Apart from the amount of BIF clasts present in each sample, a more detailed investigation was undertaken on the BIF assemblage. This was done to investigate if the proportions of BIF types changed through time. The BIF population was subdivided on the basis of colour, patterns, or combinations of both. The principal colour groups being red (Jasper), dark purple (Pearsons), honey-coloured, mid-brown, dark brown, black, blue (Riebeckite), tiger's eye and metallic. Patterns include mono-coloured, random mixes of colours, rhythmites (banded), folded rhythmites and brecciated. Different combinations of colours and patterns result in a total of 48 types of BIF (Fig.6.48). The overall BIF population is dominated by the magnetite rhythmites, followed by honey-coloured rhythmites, a black/brown mixture, mono-coloured mid-brown lutites and riebeckite (Fig.6.48). This total is biased toward the BIF population present in the Meso deposis, which were the most abundantly sampled. Subdividing the BIF population per age of deposit:

6.5.5.2 BIF Types per age

The breakdown of BIF types present in the Eocene samples (Fig.6.49a) reveals an assemblage dominated by jasper and mono-coloured honey and mid-brown varieties. The Pre-Proto deposits (Fig.6.49b) are dominated by jasper, Pearsons, and mono coloured varieties of BIF. Both the Eocene and Pre-Proto deposits have significantly different suites of BIF clast types to the younger Proto, Meso and modern deposits (Figs.6.49c-e), which have significantly higher proportions of magnetite rhythmite and riebeckite, but lower proportions of jasper types. Within the Proto, the BIF clast types show a downstream transition from a Meso-like assemblage in the upstream deposits (Fig.6.50a), to an assemblage which contain more Pearsons (purple), jasper and mono coloured varieties in the downstream deposits (Fig.6.50 b & c) - with the upstream, mid and downstream reaches defined as for figure 6.44. This partitioning in the BIF population thus confirms the longitudinal changes that occur in the Proto deposits in the other clast types as discussed in Section 6.5.3.3. The Meso deposits do not show a significant downstream change in their BIF assemblage (Fig.6.50d-f). The modern deposits show a downstream transition from a rhythmite-dominated (Meso assemblage) to a mono-coloured assemblage

(Proto assemblage) (Fig.6.50g-h). This, together with the relative lack of BIF, in the modern deposits leads to the suspicion that a significant proportion of the BIF clasts present in the modern river are derived from the erosion of the terraces flanking the modern channel.

6.5.5.3 BIF types Summary

The proportions of BIF types change from an assemblage with red, purple and mono-coloured brown varieties to one with more rhythmites, riebeckite and black/brown mixtures. Reasons for this change in BIF types through time could either be due to broad changes through the vertical stratigraphy of the BIF sequence itself, spatial changes in BIF types across the Transvaal basin, or a combination of both. If the changes in BIF types through time are due to vertical changes in the BIF stratigraphy, then this would reflect incision by the Orange River into the BIF sequence. However, spatial changes across the BIF basin would reflect change in provenance through time as the drainage evolved.

A complication with attempting to determine the provenance of the BIF clast types is that the BIF sequences contain highly variable sedimentological signatures, right down to the mm scale and that a variety of BIF types can be encountered in the space of a few metres (H.Tsikos pers comm. 2004). However, broad changes in the BIF stratigraphy do exist. The Kuruman iron formation, at the base contains more rhythmites, whilst the Griquatown iron formation, in the upper part of the stratigraphy contains the more massive lutites (Fig.6.51)(Beukes 1986). Thus the increase in the amount of rhythmites in Meso gravels can be explained by incision of the river into this sequence.

The riebeckites form due to Na-metasomatism in the southernmost part of the outcrop area (Griquatown, Prieska) (Gutzmer, J. Pers Comm, 2004), thus the increase in the amount of riebeckite clasts in Meso times may be related to more tributary input from this area, or that the area was covered by Karoo basin rocks in Eocene times.

Jaspilite is a common constituent of the Kraaipan Group banded iron formation, which crops out in the Vaal River catchment area (SACS 1980). Pebbles of jasper also occur in the Lichtenburg gravel deposits (Armstrong *et al.* 1987) and thus, as with the yellow chalcedony, jasper clasts can be considered a Vaal River type clast, which could explain its association with jasper in the older deposits.

Although very briefly dealt with here, a more detailed study on the BIF clasts and their provenance could be highly beneficial to analysing changes in the upstream areas of the Orange and Vaal River drainage basin.

6.5.6 Clast assemblage linked to grain size distribution

In an attempt to obtain an clast assemblage, based on the weight percentage of the size fractions, clast assemblages at the Sendelingsdrif and Daberas deposits were combined with accurate size distributions done on the gravel. The results indicate that Nama quartzite is the dominant lithology in gravels from Sendelingsdrif and Daberas (Fig.6.52a-f), constituting between 40-60% of the bulk composition. An exception is the Daberas sample 4 lower (6.52d), which has only 30% Nama quartzite, but an increased amount of local metasedimentary clasts. This can be explained by this sample's proximity to a tributary input point supplying large amounts of metasedimentary clasts.

This method of analysing the clast assemblage is not as effective as that used in the remainder of this study, as most of the detail from the less significant (in terms of weight) finer size fractions is lost. In terms of calculating the bulk assemblage changes in the gravel, for accurate comparisons of assemblage maturity, it could be a very effective tool, but the large size of sample required for the grain size distributions make it logistically difficult limiting its use. Unfortunately the size distributions obtained in figure 6.52 could not be applied to all of the deposits with confidence, as it is expected that (as with the MCS) the bulk size distribution of the gravel will change from upstream to downstream and its effect on the overall composition calculated is unknown. Although impractical, this combination of size distribution with clast assemblages is probably the most accurate method of determining the clast assemblage.

6.6 Overall Summary and Discussion

The survival of less competent clast lithologies (dolerite, limestone and metasedimentary) in the modern deposits, as well as the relative lack of clast roundness, indicates that the river competancy has decreased relative to the older deposits, and has become less able to deal with the sediment supplied to it.

The method of analysis used in this study was successful in characterising the clast assemblage from each terrace suite, as defined using bedrock strath height and the overall deposit geometry. Linking the clast proportions per size fraction with the size distribution of the gravel to obtain an accurate bulk composition of the gravel, has potential to provide important information. In the samples that were analysed using this method, almost 50% of the gravel is composed of Nama quartzite, which makes up a considerably smaller proportion of the clast types entering the river. This method could be powerful for analysing downstream maturity changes, which were not suitably addressed in this study. In addition, the impact of abrasion and selective sorting on the clast assemblages could not be addressed due to the immense complications introduced by the finer size fractions.

The overall clast assemblage in the study reach changes from being dominated by siliceous clasts, mostly derived from the Middle Orange and Vaal River reaches inland, to locally derived lithologies, reflecting the ongoing incision of the Orange River into the southern African hinterland, as well as the aridification of the sub-continent and change from a Vaal to an Orange River dominated assemblage (Fig.6.53).

Vaal River type agate clasts, derived from both the Ventersdorp andesites and an unknown source of lavas in the Upper Vaal River basin, probably from an eroded variety of Karoo basalt (Armstrong *et al.* 1987), dominates the Eocene deposits, and gradually diminishes through time in the younger deposits. This agate-rich population in the older deposits is explained by aggressive erosion of Karoo and Ventersdorp volcanics in the Vaal River catchment. Vein quartz clasts, abundant in the Eocene deposits, also decrease in proportion in the younger deposits. BIF clasts, present as two discrete slugs are initially accompanied by jasper clasts in the 1st slug, which occurs in post-Eocene/Pre Proto 1 times and then diminishes in amount in the Proto deposits. The 2nd slug, accompanied by riebeckite clasts becomes prominent in Intermediate and Meso times, diminishing in modern times. Karoo sedimentary clasts, most of which have been subject to some degree of contact metamorphism from the numerous Karoo intrusive rocks become prominent in Proto times, after which they diminish in amount in the Meso and modern samples. Karoo basalt, zeolite and Orange River type agate clasts, all indicative of an Orange River source become prominent in Meso times, and increase in amount in the modern deposits (Fig.6.53).

By the time the first Eocene samples were deposited, substantial incision into the hinterland had already occurred, to account for the presence of abundant +25 and +40mm exotic clasts which have been sourced from ca. 2000 km upstream. Furthermore, the presence of Rosyntjieberg and Nama quartzite large cobbles/small boulders in the sample at the mouth of the Orange River, which has been correlated with the Eocene age deposits on the basis of its exotic clasts, indicate that considerable incision into the study reach had occurred by this time. The relative lack of volcanics or NMP basement clasts in these deposits is consistent both with a young, canyon-like incision without a well-developed tributary network, as well as an aggressive, abrasion dominated system aided by intense weathering. The exact provenance of the vein quartz clasts present in the Eocene deposits is unknown, but is thought to be from either the NMP or Gariep Supergroup, both of which host numerous quartz veins. The presence of so much vein quartz, without the accompanying NMP lithologies lends weight to an abrasion-dominated system which has battered out the softer volcanic, granite and gneissic clasts. Sklar & Dietrich (2001) in their tumbling experiments noted that lithologies were abraded 3 times faster when tumbled with quartzite as opposed to against each other. Parallels can be drawn with the study of Abbott & Peterson (1978) who attributed the mature composition of a conglomerate in California to an abrasion dominated system aided by a humid climate. However, the presence of only Rosyntjieberg quartzite and upper Nama quartzite, and absence of lower Nama quartzite clasts lends weight to the incomplete incision argument.

By Pre-Proto 1 time, incision of the Richtersveld was essentially complete and a well developed tributary network would have been supplying locally derived gravel to the Orange River. Karoo sedimentary clasts begin to become prominent in Pre-Proto 1 times, and peak in Proto times and it is not known whether this trend is related to the Orange River or the Karoo sedimentary clasts themselves. The lack of Karoo sedimentary clasts in the Eocene deposits is puzzling as there should have been more Karoo outcropping in the drainage basin in earlier times. Thus, the lack of Karoo could be indicating that few were able to survive in an aggressive, abrasion-dominated river, and only when the river was flooded by a variety of clasts types, were they able to survive. An alternative is that the level of erosion in the main Karoo basin had proceeded down to the level of the dolerite sills (Veevers *et al.* 1994), producing large volumes of hornfels baked to varying degrees. Only when the metamorphosed varieties of Karoo shales were supplied to the river, were they tough enough to survive fluvial transport. Most of the Karoo shales encountered in this study did show signs of baking/alteration, and many were rounded to some degree, which does indicate that they are more competent than the typical fissile shales usually encountered in outcrop. The decrease in amount of Karoo sedimentary clasts in the Intermediate, Meso and modern deposits is consistent with a shrinking Karoo basin with time.

The two slugs of BIF clasts, distinguished by the presence of riebeckite in the younger slug, appear to be confirming incision into the BIF source rocks of the Transvaal basin, with the younger slug having a higher proportion of rhythmites to lutites, the inverse of the stratigraphy from which they are derived. The two pulses of BIF could be representing tectonic activity in the Asbestos Hills area on the edge of the Craton, supplying varying amounts of BIF to the river at different times, or the exhumation of Transvaal Supergroup rocks beneath Karoo basin cover. The later exhumation of riebeckite-bearing BIF sequences in the Prieska area would account for the presence of riebeckite in the 2nd slug of BIF, which is most prevalent in Meso times.

Orange River type agate clasts are present in limited amounts in the Eocene deposits, although, this together with the lack of Karoo basalt and zeolite clasts in the older deposits is puzzling, as the Drakensberg would have been an elevated area, being actively eroded at this stage. This may be indicating climatic differences in Eocene times, or that the Orange River had not networked into a major river in these regions, and that the Vaal River was the true trunk stream. Once again, abrasion may be excluding the basalt and zeolite clasts from the older deposits. The relative increase in the Orange River derived suite of clasts in the younger deposits reflects the demise of the Vaal River as a source of clasts by Meso times. This trend is probably due to the aridification of southern Africa, with the Drakensberg area being the primary supplier of both water and clasts to the area.

6.7 Synopsis

- The modern river is less competant, and less able to deal with the clasts supplied to it, as evidenced by the relative lack of clast roundness, downstream rounding of Nama quartzite, gravel maturity and survival of softer lithologies.

- Methods of clast assemblage and roundness analysis per size fraction used in this study could successfully differentiate between the various deposits that were defined using strath height and deposit geometry.

- The Eocene is characterised by high proportions of siliceous clasts with origins in the hinterland, namely, Vaal River type agate, chert, BIF (accompanied by red jasper) and vein quartz. This assemblage is explained by both a less advanced stage of incision and higher amounts of abrasion, with both being likely.

- Pre-Proto 1 deposits are dominated by locally-derived clast varieties, but proportions of Vaal River type agate are still high. BIF and Karoo sedimentary clast proportions increase from the Eocene deposits.

- Pre-Proto 2 deposits contain a mixture of clast assemblages between Pre-Proto 1 and Proto, and vary according to their local position within the study reach.

- Proto deposits, contain lower proportions of Vaal River type agate and BIF than the older deposits, but higher proportions of Karoo sedimentary clasts. Proto assemblages vary within the study area, either signifying changing longitudinal assemblages, or younger gravels preserved on Proto straths in the upstream reaches.

- Intermediate and Meso deposits have lower proportions of Karoo sedimentary and Vaal River type agate clasts than the Proto, but higher proportions of BIF, Orange River type agate, Karoo basalt and zeolite.

- Modern deposits have less exotics generally, apart from Orange River type agate, Karoo basalt and zeolite, proportions of which are similar to, or higher than in the Meso deposits.

- Two slugs of BIF are evident: the first, which is most prominent in the Pre-Proto 1 deposit has high proportions of red, purple and mono-coloured varieties. The second, which is most prominent in the Meso deposits has high proportions of riebeckite and rhythmite varieties, explained by incision into the BIF source rocks.

- Augrabies Falls upper terraces are tentatively correlated with the Pre-Proto deposits in the study area, on the basis of their BIF population. However, larger samples will need to be analysed to prove this conclusively.

- Clast assemblage evolution is consistent with stripping off inland stratigraphy, as well as a switch to a purely Orange River source, following regional aridification of the middle and western parts of southern Africa by Plio-Pleistocene time.

7 Distribution of Diamonds in the Lower Orange River

7.1 Introduction

The Orange River has played an integral part in producing the most spectacular gem diamond placer yet discovered – in which diamonds from the Kaapvaal Craton have been transported to the coast to be concentrated by fluvial, marine and aeolian processes. The region has produced in excess of 100 million carats (cts) of gem quality diamonds, the bulk of them having been recovered from the Namibian coastal strip between Oranjemund and Lüderitz (Ward *et al.* 2002). Since the discovery of diamonds at Kolmanskop near Lüderitz in Namibia in 1908, there has been much speculation as to their source, with theories ranging from a submarine origin to an interior-derived source (Lotz 1909; Cornell 1920; Wagner & Merensky 1928; Williams 1932). Cornell (1920) recognised the significance of the exotic suite of clasts in the gravel as being the same as that present in diamond diggings along the Vaal River, and prospected at many localities within the study reach, albeit unsuccessfully, as did Reuning in 1927 and Caplan in the 1930's. However, it was only in the 1960's that the first discoveries were made: at Reuning near Sendelingsdrif by geologist R.Baxter-Brown in 1966, in gravels a foot deeper than Caplan had dug (Wilson 1972b;a). The reason for their delayed discovery is the inherently low grade of the Orange River deposits, in contrast to the marine and aeolian placers on the coast which host higher concentrations of diamonds. It is now fairly well accepted that the Orange River diamonds are derived from the numerous kimberlite intrusions that occur on the Kaapvaal craton (de Wit 1999)(Fig.7.1), although it is not always agreed as to their mode of transport, with the Dwyka glaciers being an alternative to fluvial transport (Moore & Moore 2004).

From their initial date of discovery in 1966, prospecting for diamonds commenced on both sides of the river, and exploitation commenced in the 1970's, and has continued unabated until the present. Most mining has focussed on the Proto terrace deposits which yield higher grades than the Meso terrace deposits. Diamonds are extremely rare and are measured in carats per 100 tons (cpht) of gravel, where 1 carat (ct) weighs 0.2g. Although rare, diamonds are essentially just another clast-type present in the gravels, and information relating to their presence and concentration in deposits furnishes valuable information on the system, from both a geomorphological and sedimentological perspective.

7.2 Previous Work

Although studies of the fluvial controls on metalliferous placers are numerous (Smith & Minter 1980; Slingerland 1984; Kuhnle & Southard 1990; Day & Fletcher 1991; Levson & Blyth 2001; Carling *et al.* 2004), those relating specifically to diamonds are more scarce (Sutherland 1982; Allan & Frostick 1999).

In addition, very little relating to controls has been published from the Lower Orange River area and much of the prospecting information is contained in company reports. Fowler (1976), detailed the extensive prospecting work done at the Daberas deposit on the north (Namibian) bank whilst Pienaar (1977) and Steyn (1982) reported on occurrences on the south bank. Most of the reporting is very site specific and a lot of the sampling done treated too little gravel for accurate grades to be determined, or obtain a clear picture of the controls on diamond concentration. This seemingly unpredictable distribution led to the belief that "a diamond is like a pig – it lies where it wants to"(van Wyk & Pienaar 1986). Van Wyk and Pienaar (1986) were the first to document the overall characteristics of the gravel and diamonds in the area, synthesising data from mining and prospecting from the south bank in an excellent paper which linked higher concentrations of diamonds with the most poorly sorted gravel in areas of uneven bedrock in the oldest terraces. Ward *et al.*(1993) recognised the significance of oversize clasts and its role in the concentration of diamonds.

Jacob *et al.*(1999), using results obtained from controlled production bulk samples (>5000 tons) from Auchas Mine and the Daberas prospect, as well as selected historical data from Oenas, Reuning and Baken Mines, linked the distribution of diamonds to depositional environments found in the Proto-Orange River system, with analogues in the modern river system. A summary of these environments and controls on diamond concentration in the Lower Orange River follows.

7.3 Controls on diamond distribution

7.3.1 Sedimentary Setting: Control on Diamond Grade

Diamond differs from the metallic placers in that in comparison, diamond is a relatively light "heavy mineral", with a specific gravity (SG) of 3.50-3.53 compared with magnetite (5.18), cassiterite (6.8-7.1) and gold (19.3)(Sinkankas 1964). In addition, the size of diamond targeted is much coarser than that of the metallic placers. Diamonds are targeted from the grit and small pebble fractions (with a +2 or +3mm bottom cutoff being used on the Orange River), in comparison to gold which is recovered most commonly from the medium sand to very fine silt fractions (0.055mm to 0.5mm) (Wang & Poling 1983; Robb & Robb 1998). Thus, prospecting for diamonds targets areas where grit and pebbles are concentrated, avoiding areas where there is a buildup of sand. These coarser size fractions are better concentrated in areas of higher energy, and are accompanied by oversized clasts, an increase in the average clast size and packing density of the gravel. On the Lower Orange River, these settings are generally found in areas of steeper gradients and irregular bedrock (Jacob *et al.* 1999). This is in contrast to metallogenic placers which show higher concentrations in gentler gradient settings (Day & Fletcher 1991).

Small diamonds are easily transported by a river capable of imbricating clasts up to 3 m in diameter, thus the problem becomes one of diamond retention. Diamonds, present in the passing population of clasts slot into the gravel pore space of gravel accumulations. The retention of hydraulically sorted diamonds requires them to reside in a host gravel which is not readily subject to reworking. Bedrock irregularities not only provide for fixed sites of variable turbulence but also generate relatively fixed gravels around them, in which concentrated gravels are retained in the gravel pore space and over a long-enough period to allow successive concentrations to be added to the gravels. Thus, a fixed site of turbulence generates a fixed trapsite and these are invariably linked to the bedrock contact where the highest concentrations of diamonds are found. Fixed trapsites are generally associated with the downcutting, degradational phase and early filling of the valley (Jacob *et al.* 1999).

Examples of fixed trapsites include scours, push bars and bedrock highs, summarised in figure 7.2 and described in more detail by Jacob *et al.* (1999). Something not fully appreciated by the authors at that time was the influence of lateral tributary drainage on the positions of scours, as described in Section 3.3.4. In addition, the Orange River obtains most of its oversize clasts from tributaries, which provide further sites of relatively fixed turbulence. Gravel fabrics associated with the higher grade fixed trapsites tend to comprise a large cobble to small boulder framework which is in-filled by second to fourth generations of in-filling made up of small cobble through pebble to grit-sized material. These are termed compound fabrics and are associated with good diamond recoveries (Jacob et al, 1999). These correspond to the "poorly sorted" gravels of van Wyk and Pienaar (1986), and although they are poorly sorted from a sedimentological analysis perspective, they are well organised, stable fabrics that have built up over a period of time and the result of intensive sorting processes (Bluck 1979;1982). The grades and average stone (diamond) size increases dramatically in these fixed bedrock trapsite settings. Increased concentrations of gold from similar high energy settings are reported from Canada (Levson & Blyth 2001).

The lower energy areas of the river, situated off or above the active channel return considerably lower grades. These areas are associated with the aggradational settings, where the river flows in its own alluvium in the absence of bedrock, where there are less permanent sites of turbulence and gravel accumulation. In addition, the scale of turbulence in the system is considerably less. Clasts remain in fixed positions for a relatively longer period in some sites than in others. These are referred to as mobile sedimentary trapsites (Jacob *et al.* 1999) and include the riffle/platform (subaqueous plinth onto which a bar accretes, separate from the channel deposits), and the bar head and mid bar regions (Fig.7.2). Bar heads are recognised by horizontal or gently, upstream-dipping sheets of gravel with a framework rich in coarse discs and filled with well-packed gravel typically with a compound fabric. These sheets alternate with coarse sand and sometimes with finer gravel. It is well documented that bar heads are preferential sites of heavy mineral accumulation (Slingerland 1984; Day & Fletcher 1991; Carling *et al.* 2004). Bar tails, being sites of active scour and subsequent re-deposition during and after floods (Bluck 1971), as well as inner bank environments, where large volumes of fines are deposited are very poor sites of diamond and heavy mineral concentration as they are replaced frequently on flood events (Jacob *et al.* 1999; Carling *et al.* 2004). Bar tails are recognised by thin gravel foresets, often with more spherical shaped clasts, and large thicknesses of coarse, pebbly sand, which is cross-stratified (Fig.7.2).

Due to the importance of time in the slow accumulation of gravel bodies that are able to extract from the passing population of diamonds, short term studies over one flood cycle of heavy mineral deposition (Hattingh & Rust 1993) are thought to be relatively ineffective in addressing the controls on heavy mineral concentration.

7.3.2 Sedimentary setting: Controls on diamond size

For the Proto deposits within the study area, as with the gravel clast size, the average diamond size per site decreases in a downstream direction from 2.7 carats per stone (cts/stn) at Grasdrif to 0.7 to 1.0 cts/stn at Arrisdrif and Swartwater (Fig.7.3). Within this overall downstream decrease, there are large variations within each site, as evidenced by the ranges in values at some sites on figure 7.3. On the Lower Orange River, grade is linked to the average diamond (stone) size. This is illustrated at the Auchas Major locality (Fig.7.4), where fixed trapsites with a high grade, invariably have a high average (diamond) stone size, whereas mobile trapsites deliver lower grades and average stone sizes. This is thought to be linked to the energy and turbulence levels present in these fixed trapsites, which also host larger clast sizes. The size frequency distributions from these two populations reveal that the bar sample has less large stones, and more small stones, in contrast to the scour sample which has more large stones and less small stones (Fig.7.4). Both samples have almost equal proportions of medium sized stones (4.52mm which weighs ca. 0.9ct).

The diamond grade and size recovered from different areas in a section of the Auchas loop is illustrated in figures 7.5 & 7.6. The higher grade areas are situated at the base of the bedrock channels and scours, whereas the flatter and elevated areas of bedrock host lower grades and stone sizes (Fig.7.5), and the upper units of the deposit host the lowest grades and stone sizes (Fig.7.6).

7.3.3 Diamond concentration through time

In addition to the local concentration, an additional parameter in explaining the distribution of diamonds in the deposits is a change in the diamond supply through time. In common with the other exotic clast types which have moved through the study reach in large slugs, as discussed in Section 6.6, the passage of the "slug" of diamonds can be followed. Since the discovery of Pre-Proto 2 deposits underlying Meso gravels in the Xarries North area on the south bank, an appreciation of the potentially large time gap between the Pre-Proto 2 and Proto aggradational deposits has developed, and subsequently, most of the basal Proto deposits preserved in deep scours have been re-classified to a Pre-Proto 2 age. The almost order-of-magnitude higher grades returned from the Pre-Proto deposits preserved in the scour environments is consistent with a slug of diamonds, the peak of which passed through the area during this time. By Early-Middle Miocene times, there were less diamonds in the river declining further into Intermediate, Meso and modern times.

Although minimal mining has occurred in the Meso deposits, areas of fixed bedrock trapsites return grades of 2-5 cpht (Barker 1994), an order of magnitude lower than equivalent sites in the Proto Suite. Meso bar gravel deposits return very low grades from barren and 0.05 to 0.3 cpht (Pienaar 1977). The fact that very little mining takes place in the Meso deposits confirms their very low grades. No reliable data from the modern river is available, although it would appear that, from diver-assisted operations in favourable trapsites, grades are very low (van der Westhuizen, pers. commun., 2002).

It is very difficult to make grade comparisons further back in time due to a paucity of deposits and a lack of age control. Unfortunately, the small remnant of Pre-Proto 1 gravel at Snake is too small to yield a reliable grade or stone size. The Eocene deposit at MA1 yielded a grade of 1.2 cpht and an average stone size of 0.53 cts/stn but as the sample was only 320 tons of gravel, it is considered to be too small to be of significance. Samples generally need to be a minimum of 2000 tons and preferably larger to represent the true grade of the deposit accurately (At Auchas, the benchmark of 5000 tons appeared to represent a satisfactory sample size). Also, being a small remnant preserved in a topographic depression in the bedrock, the environment of deposition within the fluvial environment is difficult to interpret, making it more difficult to compare with younger upstream deposits. If the diamond population in this deposit is representative of the grade of the Eocene at this position, then diamonds were more abundant in Eocene times. However, it could have originally been situated at the base of an incised scour, the lip of which has been subsequently planed down by marine transgressions and regressions. If this was the case, the grade would be poor for the scour environment. Grades in the Eocene deposits in the Buntfeldschuh and Bogenfels areas are generally high, however, the marine and aeolian settings in which they occur (Stocken 1978) would be expected to upgrade a fluvial deposit from the same-age system. Thus, comparing different depositional environments through time could be extremely tenuous and will not be attempted here.

7.3.4 Diamond size through time

Although comprisons of grade are difficult to achieve, diamond size comparisons can be done with the Eocene deposits in the Bundfeldschuh/Bogenfels area. The average stone size returned from Eocene deposits is 0.2 cts/stn, whereas an increase to 0.5 cts/stn occurs in the younger deposits (Millad 2004). Further north in the Luderitz area, the Eocene gravel in the Buschtal area delivers an average stone size of 0.1 cts/stn (Hallam 1964), whereas younger deposits perform consistently at 0.35 cts/stn[1] . Although the average stone size of 0.53 cts/stn obtained from the Eocene MA1 sample near the mouth of the Orange River is tenuous, as it is only based on a total of 7 stones, the surrounding Pleistocene beaches in this area increase to stone sizes of between 0.8 to 2 cts/stn

[1] http://www.diamondfields.com

(Spaggiari *et al.*, in press). Thus, the stone size increases from the Eocene to the Pleistocene deposits.

Further upstream, the high stone sizes which are linked to the high grades in the Pre-Proto and Proto deposits (Fig.7.4a) appear to contradict the trend of decreasing stone size with age. However, equivalent settings in the Meso deposits appear to deliver higher average stone sizes (Pienaar 1977; Barker 1994) and recent prospecting results from Grasdrif confirm this.

Thus, as with the BIF population, the diamond population coarsens through time from the Eocene through to the Meso. Within each suite, the decrease in average stone size upwards in each terrace can be explained by a decrease in energy from the fixed bedrock trapsites to the more mobile sedimentary trapsites.

7.4 Discussion and Summary

Diamond concentration in the Lower Orange River is consistent with a declining and coarsening supply of diamonds through time (Fig.7.7). Although it could be argued that this is due to progressively larger amounts of dilution by local clasts, the order of magnitude decrease in diamond grades through time suggest a slug of diamonds, decreasing in amount through time, the coarse tail of which comes through last of all. Within this overall control, sedimentary settings and fixed and mobile trapsites determine the amount of diamonds able to be extracted from the passing population.

It is difficult to determine when the maximum of the pulse of diamonds reached the study area. It would appear from the dataset (Fig.7.7) that this maximum occurred in Pre-Proto 2 times. However, diamonds in transit need to be trapped in gravel, and in the early stages of incision, the river was not able to retain much gravel and would have been a predominantly flushing system. Only when incision had advanced to the stage that the tributaries were delivering abundant amounts of coarse clasts to the Orange, was some gravel retained, in the scour and push bar settings of the Pre-Proto 2 deposits. However, the amount of diamonds present in the passing population may have already been in decline during this time period and this may have been the first opportunity that the river had to retain diamonds. Thus, the peak in the diamond slug could have been in Eocene or between Eocene and the deposition of the Pre-Proto 1 deposits.

Expanding on this idea, the slug of diamonds may well mirror the slug of Vaal River agate (Fig.6.53), which peaks in the Eocene and gradually declines through the deposits. Thus the erosional event which must have downwasted large thicknesses of volcanics to yield the large abundance of Vaal River type agates, must have also denuded many of the kimberlites on the craton, yielding many diamonds which, together with the agates and other resistates, were transported to the coast by an incising Vaal River in Early Tertiary times.

If the diamonds are linked to the presence and amounts of Vaal River agates, then it should be evident in the grade of the Proto deposits which displays distinct downstream changes in clast assemblage between Noordoewer and the mouth, from a Meso-like assemblage in the upstream reach, to higher proportions of Vaal River type agates in the downstream reach (Section 6.5.4). From a process point of view, the upstream localities should return higher grades as the energy and clast size increases upstream. However, this is not the case: in the upstream deposits, Grasdrif Proto and Meso deposits have similar very low grades, the difference being that the Meso hosts a larger average stone size. Thus, at this locality the diamond data confirms the clast assemblage data. In the middle reach, dominated by Karoo sedimentary clasts, Maclekop and Lorelei West reportly yielded fairly low grades for the Proto deposits (Pienaar, pers comm., 2001). In the downstream reach, grades are variable, but appear to increase in the lowermost Proto deposits (above strath level) to a maximum at Koeskop and Swartwater. Thus, it would appear that the clast assemblage match the diamond data, and therefore time exerts a significant control on diamond distribution in the study reach.

Thus, at each locality, the distribution of diamonds present is principally controlled by the age of the gravel, and secondly by the sedimentary setting, which has a large influence on the average stone size.

7.5 Synopsis

- Diamond concentration in the Lower Orange River deposits generally declines through time and diamond size coarsens.

- Within this overall time control, sedimentary settings provide secondary control on extracting and retaining the passing population of diamonds, with areas of the river providing fixed sites of turbulence hosting the highest grades and coarsest diamonds as opposed to the more mobile sites of turbulence.

- Diamond concentration mirrors the demise of the Vaal River suite of clasts, thus clast assemblage analysis, as performed in this study could become a reliable grade indicator.

8 The erosional and Cainozoic depositional history of the Lower Orange River

8.1 Discussion

The Orange River is unusual for a large river, as indicated by its longitudinal profile which steepens in the lower reaches as it flows through a through a high relief landscape. In addition, the Orange derives the majority of its water from its upstream reaches in the Drakensberg Mountains. This has resulted in the unusual situation of a large river in an arid landscape. The ungraded profile of the Lower Orange River, together with its major knickpoint at Augrabies Falls, indicates that it has incised in response to an uplift event. Projection of the graded Upper Orange River profile indicates up to 1 km of uplift has occurred, which is in agreement with estimates of uplift obtained from studies in the offshore Orange basin (Dingle *et al.* 1983; Brown *et al.* 1995; Aizawa *et al.* 2000). In addition, the Lower Orange River in the study area is ca. 700 m below the level of the surrounding African Surface peneplain, thus 600-1000 m of incision has occurred. The timing of this uplift event is controversial, with the fission track data indicating large amounts of uplift in Early Cretaceous times (Brown *et al.* 1990), although a lot of the denudation occurring at this time would be due to the lower base level that breakup would have provided, without invoking large amounts of uplift.

Partridge & Maud (1987) demonstrated two relatively late uplift events: in the Early Miocene and Plio-Pleistocene. However, they use the Proto Orange River terraces as evidence for Miocene uplift, which is incorrect as these are aggradational deposits: the incision associated with these terraces must have occurred prior to this time. Aizawa *et al.* (2000) recognise an Early-Tertiary uplift event in the offshore Orange basin, and this event is most likely to have caused the up to 1 km incision of the Orange River. The expression of this uplift is a change in character from a sand-carrying Orange River, building a large delta offshore near the present position of its mouth (Brown *et al.* 1995) in the Late Cretaceous to a gravel bearing river in Eocene times, bringing gravel clasts up to 60 mm in length from a source 1000-2000 km upstream (Ward & Bluck 1997). At this time, incision into the Richtersveld had begun, evidenced by the presence of small boulder sized locally-derived quartzite clasts in an Eocene deposit at the Orange River mouth. The presence of these quartzites, derived from the Rosyntjieberg and upper Nama units, with a relative lack of more labile NMP clasts is either indicating extremely high rates of clast abrasion, or incomplete incision into the Richtersveld, with incomplete incision being more likely due to the lack of quartzite from the lower Nama units, which is common in younger deposits.

There is very little depositional record for the main part of the incision, which proceeded until the mid-Oligocene sea level low, recognised globally (Vail *et al.* 1977). Deposits preserved in the later stages of this long phase of incision are limited to gravels preserved in the Stolzenfels kimberlite near Augrabies Falls, and two isolated occurrences preserved on bedrock straths ca. 80 m above the modern river level (Pre-Proto 1 deposits). In addition, a number of deposits are preserved at the peak of the incision, at the base of scours (Pre-Proto 2 deposits), situated up to 50 m below bedrock strath level. Thus, ca. 90 % of the incision occurred before any record was deposited. The clast assemblage in the Pre-Proto deposits evolved to an assemblage dominated by locally-derived clasts, but with significant proportions of BIF and Vaal River type agate clasts. A long period of aggradation followed the mid-Oligocene lowstand to build the Proto deposits which are dated to 19-17.5 Ma (Pickford *et al.* 1996b). There is evidence for both sea level rise and climatic fluctuations to account for this aggradation, although the natural expansion that would occur in the tributary network, following an incision event, could also account for this aggradation. The abundance of more poorly rounded locally derived clast types in the upper units of deposits adds more weight to the argument that aggradation is primarily due to an increase of material supplied by the tributaries.

Late Miocene/Early Pliocene incision straightened the Orange River channel by cutting out many of the meander loops that characterised its inherited course. Following this, several shorter-term cycles of aggradation and incision ensued in the Plio-Pleistocene, recorded by the Meso suite of deposits. Development of the terraces during this time is consistent with the river incising into a slow, isostatically uplifted margin, with bedrock straths being cut during base level rise, and incision in times of base level fall, and aggradation as result of tributary activation following the uplift event or wetter climatic phases. Clast assemblages evolved from a Karoo sedimentary and Vaal River dominated suite of exotics in the Proto, to a BIF and Orange River dominated suite of clasts, reflecting the stripping out of the Karoo basin, as well as the progressive aridification of the central parts of southern Africa.

The expression of the incision of the Orange River is a very dissected landscape. It is difficult to distinguish between pre-existing tributaries, and tributaries which would have originated at the Orange River and cut their way back into the landscape following the initial incision event. The largest tributary in the study reach, the Fish River undoubtedly existed prior to incision as evidenced by its incised meanders near its confluence with the Orange River. The expression of the Orange River incision is best preserved in the Fish River Canyon, a ca. 500-600 m deep canyon which displays the various stages of incision perfectly in a downstream direction.

For the remainder of the tributaries the drainage basin area increases towards the coast, as does the Orange River valley width, both being consistent with incision being generated from the coast. In the upstream reaches, softer lithologies host larger tributary catchments. Knickpoints are more prevalent on the coastal plain reach tributaries than the upstream reaches, either indicating response to recent Orange River incision, loss of the incision signal in the upstream reaches or fossilisation of knickpoints generated by much older Orange River incisions, resulting from the hyperarid conditions that have prevailed in this area since Plio-Pleistocene times (Ward *et al.* 1983).

Although the main phase of incision occurred between the Early Cainozoic and Mid-Oligocene, the Orange River is not graded through the Richtersveld reach and the landscape has still not fully adjusted to incision, as evidenced by the persistence of steeper gradients in Proto, Meso and Modern times in this reach, as opposed to the coastal plain reach which has a low, constant gradient. In the Richterveld reach, the modern Orange River gradient is highest in the more resistant volcanic and granitic rocks, especially when flowing across strike. However, most of the sites of turbulence (rapids) and knickpoints are situated at tributary input points, where the gravel delivered by the tributary confines the channel and inhibits bedrock incision. Deep scours are located in areas where the channel is confined, namely in inner channels developed in resistant bedrock, on meander bends and downstream of tributary input bars. In the modern river, considerably less scouring occurs in the coastal plain reach. However, deep scours are common in the Pre-Proto bedrock surface along the coastal plain reach, indicating that this reach was a significantly more confined and higher energy reach than at present. In addition, tributaries that are presently carrying sand were delivering coarse gravel to the Proto-Orange River channel. Thus it would appear that the coastal plain reach has matured relative to the Richtersveld reach since Miocene time.

There are not many large, incised rivers flowing through arid landscapes with which to compare the Orange River. One such river is the Colorado in the western USA, which derives its water from the Rocky Mountains and flows across the semi-arid Colorado plateau, which, together with its tributaries has incised the landscape to form a spectacular array of canyons, the largest of which is the Grand Canyon. Incision of the Colorado River has occurred fairly recently, ca. 6-1.2 Ma (Hamblin 1994). So although comparable, these two rivers provide examples of incisions in very different states of advance. The association between tributary input points and rapids is stong in the newly incised Colorado River and its tributaries. The fact that this association is still present in the relatively old Orange River incision could be indicating the importance of tributary input in limiting the rate of landscape evolution. The world-wide dataset of incision rates in modern rivers (Wohl 1999) indicates that the Orange could have completed its entire incision within less than a million years. These data, however, apply only to the channel, and it must take considerably longer to remove the remainder of the landscape, via this channel and its system of tributaries. The rate of lowering of the trunk stream determines the rate of lowering of the tributaries. If the trunk stream is not incising because it cannot remove all of the material supplied by the tributaries, lowering of the tributaries can not occur, only flattening of the tributary gradients until they supply less material to the trunk stream. At some point the trunk stream is able to incise again, and will continue to do so until the tributaries start supplying too much material to cope with and the cycle should continue until the tributary gradients have flattened out sufficiently not to yield any more gravel, as is the case in the present coastal plain reach of the study area. This cycle of incision governed by tributary inputs could have implications for landscape evolution models.

Tributaries are also important in the study area in that they control the grain size present in the Orange River bars and rate of downstream fining in the maximum clast size (MCS) population. In the modern river within the study reach, there are 4 major sedimentary links (discreet downstream fining reaches which are hardly disturbed by tributary inputs). The position of the start of a new link being controlled by significant additions of durable clast types, namely Rosyntjieberg and Nama quartzite, and felsic volcanics. Although these resistant clast types are a minority of the incoming tributary-supplied clast population, they dominate the link into which they are introduced, a consequence of abrasion as opposed to selective sorting. The material present within each link also controls the channel morphology to some extent, with a complete lack of bedrock straths being evident in links dominated by quartzite clasts. The rate of downstream fining in the Proto and Meso deposits is significantly lower than in the modern river, which is either indicating greater volumes of tributary input in the older systems, a more competant older river or a coastline some distance offshore. The composition of the MCS population indicates a greater volume of tributary input, as well as a more competant river in the older deposits.

The more labile clast lithologies breakdown quicker than the resistant clast types, and dominate the fine size fractions. Clast roundness generally decreases with grain size, reflecting continual clast break down. Clast maturity increases in a downstream direction, more evident in the coarser size fractions which continually lose clasts to the finer size fractions. The downstream increase in clast maturity is also more evident in the Proto and Meso deposits as opposed to the modern deposits. On the basis of more poorly rounded clasts and presence of higher proportions of labile lithologies, modern deposits appear to have been laid down in a less competant river than the older river, unable to deal with the sediment supplied to it.

The general assumptions made about increases of clast roundness with distance travelled do not hold true in this study, probably due to the continual introduction and breakdown of clasts into smaller size fractions. Downstream changes in clast roundness of the resistant Nama quartzite population reveals that the coarse size fractions increase in roundness downstream for all deposits. However, in the smaller size fractions the clast roundness decreases downstream in the Proto and Meso deposits, but increases in the modern deposits. This reflects the general lack of energy and large clasts in the modern river on the coastal plain in comparison with the coarser older deposits.

Arguably the most important consequence of tributary input has been the supply of coarse gravel necessary for the trapping of diamonds from the passing population. Concentrations of diamonds are higher in areas of the river where coarse gravel has been trapped and retained for long periods of time. However, diamond concentration in these favourable settings decreases through time, with the peak occurring in the Pre-Proto 2 deposits, inferred to be of mid-Oligocene age. However, it is not surprising that this peak coincides with the change from an exotic-rich to a more locally-dominated assemblage in Pre-Proto times, notably with the addition of coarse gravel to the river. Thus, it is suspected the actual peak in the diamond slug occurred prior to Pre-Proto 2 times, but were unable to be retained in the incising, post-Eocene River. The drop in diamond concentration through time can be linked to the lack of Vaal River type agate clasts in the younger deposits and switch to an Orange River provenance, a catchment area hosting relatively fewer diamond-bearing kimberlites than the Vaal River catchment.

8.2 Future Research

- Follow up work needs to be carried out on the sand fraction from the samples collected in this study, as well as extending the modern river sampling both upstream, as well as up the coast into the Namib sand sea.

- U-Pb SHRIMP dating of zircons separated from samples collected during this study, currently in progress, for accurate provenance information.

- A more detailed geomorphological study on the Fish River Canyon, which appears to have retained a more complete record of the incision than the Orange River.

- Burial cosmogenic analyses, reportedly able to date up to 5 Ma reliably, to provide some age resolution on the Meso deposits.

- Investigate rates of size decline in individual lithologies more thoroughly in both the modern system as well as in the terrace deposits to obtain more conclusive clast abrasion data.

- Testing the concept and developing landscape evolution numerical models which incorporate the idea of a cycle of incision governed by tributary inputs.

8.3 Conclusions

1. The Orange River incised the landscape between 600-1000 m, following a Late Cretaceous/Early Tertiary incision event after which it changed from a sand-to gravel bearing river. Height below the planated African surface, longitudinal profile projection and offshore data constrain the magnitude of incision. The majority of the incision (ca. 90%) proceeded leaving behind no record within the Orange River valley.

2. In the series of terraces flanking the modern river in the study area, two suites can be distinguished by their river course, bedrock strath level, overall geometry and clast assemblage.

3. The older Proto suite of terraces are the higher lying terraces, with a sinuous river course markedly different to the modern river. This suite comprises the Pre-Proto and Proto deposits. The Pre-Proto deposits are small remnants fortuitously preserved close

to the end of the Eocene to Mid-Oligocene phase of incision. The Early-Middle Miocene Proto deposits are aggradational deposits, built in response to a combination of base level rise or increased supply of material from tributaries, due either to climatic fluctuations or the natural expansion of the tributary network following trunk stream incision. Clast assemblage and downstream fining data indicates tributary supply to be the more important variable.

4. The younger Meso suite of terraces are the lower lying deposits with a course very similar to the modern river and comprises the Intermediate, upper and lower Meso deposits. The Intermediate deposits are only present in specific reaches of the river. Upper and lower Meso deposits are generally thin, laterally extensive deposits that represent short phases of incision and aggradation during the Plio-Plestocene.

5. Comparison of longitudinal bedrock profiles, rates of downstream fining and clast type proportions and roundness indicate that the modern river is a considerably less competant river than in Proto and Meso times. In addition, in Proto times, the coastal plain reach was a more confined, high energy river flanked by a higher-relief landscape. Although relatively incompetant in comparison with the older river, high rates of clast abrasion are indicated in the modern system by the change in coarse clast assemblage from a quartzite poor population entering the river, to a quartzite rich population in the river.

6. Clast assemblages change from siliceous, exotic-dominated assemblages in the Eocene deposits, to locally-dominated assemblages by Pre-Proto time. The changes in proportions of exotic clast types gradually reflect the stripping out of the Karoo basin, and the demise of the Vaal River derived suite of clasts in favour of an Upper Orange River suite as result of aridification of the central and western parts of southern Africa.

7. The expression of the Orange River incision is a dissected landscape, where most tributaries appear to have developed in response to the original, and continuing Orange River incision. The persistance of the ungraded profile for so long after the initial large incision event, points to either recent tectonics, which cannot be proved, or the lag time involved in landscape adjustment following incision into a plateau. Tributary input points, accounting for most of the knick points, limits the rate of bedrock incision in the channel and consequently, landscape evolution.

8. The highest grades and coarsest diamonds occur in the sedimentary settings that are able to extract and retain the passing population of diamonds effectively, notably in fixed bedrock trapsites. However, diamond concentration in these settings decreases in the younger deposits, as diamond size increases, consistent with a declining and coarsening supply of diamonds in the river. Diamond input appears to be linked to the Vaal River suite of clasts pointing to that catchment as a more prominent source than the Upper Orange.

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