GEOMORPHOLOGICAL FEATURES OF THE SKELETON COAST NATIONAL PARK

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SCIONA
BRIDGING THE KUNENE

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PREFACE

This document was prepared for SCIONA (*Co-designing conservation technologies for Iona - Skeleton Coast Transfrontier Conservation Area, Angola - Namibia*), a project funded by the European Union (EuropeAid/ 156423/DD/ACT/Multi; grant agreement FED/2017/394-802) and led by the Namibian University of Science and Technology (NUST), in association with the Instituto Superior de Ciências de Educação da Huíla (ISCED, the Higher Institute of Education Sciences of Huíla, Angola). The overall goal of the SCIONA project was to strengthen cross-border ecosystem management and wildlife protection in the Iona – Skeleton Coast Transfrontier Conservation Area (TFCA) through co-designing and implementing conservation monitoring technology with the park authorities and surrounding communities.

The document provides an overview of the main landforms and geomorphological features of the Skeleton Coast National Park (SCNP). The document is not meant as a textbook or scientific document, but to inform and assist users such as park managers.

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INTRODUCTION

The Skeleton Coast National Park (SCNP) stretches about 500 km along the Atlantic Coast of Namibia, from the Ugab River in the south to the Kunene River in the north. To the south, the park borders on the Dorob National Park, and in the north on Angola's Iona National Park, with which it forms the Iona-Skeleton Coast Transfrontier Conservation Area. On the east, the SCNP borders communal conservancies – Marienfluss, Orupembe, Sanitatas, Okondjombo, Puros, Sesfontein, Torra, Doro!nawas – and the Palmwag tourism concession area. The park's east-west width varies between 25 and 50 km and it covers $\pm$ 17,000 km^2. The elevation ranges from sea level to 703 m.

The area was classified as the *'Namib biogeographical province'* of the *'Africotropical realm'* by Udvardy (1975), the *'Namib Desert'* in the *'Deserts of Southern Africa ecoregion'* of the *'Afrotropical biogeographical realm'* by the WWF (2001; Olson & Dinerstein, 2002), and the *'(Namib) Desert biome'* by Irish (1994), Mendelsohn *et al.* (2002) and Shaw (n.d.).

The SCNP is bordered in the west by the Atlantic Ocean, and more specifically the Benguela Current Large Marine Ecosystem, stretching from Cape Agulhas in South Africa in the south (35^{o}S) to about Moçâmedes (former Namíbe) in Angola in the north (15^{o}S). The cold Benguela Current from the south meets the warmer, tropical Angola Current from the north at the seasonally shifting Angola-Benguela Front between 14 ^{o}S and 16^{o}S. During late summer and autumn, when southerly winds are weaker, the Benguela upwelling system also weakens and warmer water from the tropical, more saline Angola Current intrudes further south (Sakko, 1998).

Both Mendelsohn and el Obeid (2005), and the UNESCO/AETFAT/UNSO Vegetation Map of Africa (White, 1983) classified this region as the *'Namib Desert'* vegetation unit, while Giess (1971) classified the vegetation of the area between the Ugab and Huab Rivers as *'Central Namib'* and from the Huab to the Kunene as *'Northern Namib'.*

According to Robertson *et al.* (2012), the major landscapes are (i) seaward-sloping sandy or gravelly hard crusted pediment with rare inselbergs, (ii) sand dunes and sheets with rare inselbergs, (iii) rocky basaltic areas, (iv) broken terrain with thin soils, (v) coastal salt pans, (vi) incised valleys of ephemeral rivers, (vii) eroded igneous complexes and (viii) alluvial fans. The Kunene River is the only permanently flowing river.

The Ministry of Environment's Skeleton Coast National Park Management Plan (MET, 2013) lists the following coastal and terrestrial habitats: (i) sandy shores; (ii) rocky shores; (iii) littoral shelf above the high-water mark; (iv) Kunene mouth; (v) gravel plains used as Damara Tern breeding areas; (vi) coastal gravel plains; (vii) inland gravel plains and associated drainage lines; (viii) lichen fields; (ix) coastal rocky hills; (x) ephemeral rivers – Ugab, Huab, Koigab, Uniab, Hoanib, Hoarusib, Khumib, Sechomib, Nadas, Engo; (xii) sand dune belts; (xiii) small ephemeral fountains and springs – e.g. Uniab delta, Oasis, Auses, Hunkab, Okau, Ganias and Khumib springs; (xiv) vegetated dune hummocks; and (xv) salt pans (MET, 2013).

A coastal profile (Anonymous, 1999) of the Erongo Region further south identifies seven major habitats which are also found in the SCNP: (i) gravel plains; (ii) coastal hummocks; (iii) sand dunes; (iv) washes; (v) riverbeds; (vi) rocky ridges; and (vii) inselbergs.

GEOLOGICAL HISTORY OF NORTH-WESTERN NAMIBIA

The oldest north-west Namibian rocks (Epupa, Huab, Kunene Metamorphic Complexes) formed in mobile belts (long, narrow, tectonically active areas) between the ancient Angolan and Kaapvaal Cratons (large rafts of relatively stable, thicker crust, rooted deep in the upper mantle).

Cycles of rifting and assembly welded the mobile belts onto the cratons, accreting into the larger Congo and Kalahari Cratons. The original metavolcanic rocks underwent several phases of metamorphism during the Vaalian (< 2,100 Ma) and Mokolian (1,759-900 Ma) ages, with the Meso-proterozoic Kibaran Orogeny (1,400-1,100 Ma) culminating in the assembly of the supercontinent **Rodinia** (1,780-850 Ma; Upper Palaeo-proterozoic to Meso-proterozoic). A 400-million-year period of relative tectonic quiescence allowed extensive erosion of these mountains.

The evolution of the Neo-proterozoic **Damara Supergroup** got underway around 850 Ma, when Rodinia started breaking up through formation of two intra-continental rifts into which great thicknesses of sediments and some volcanic materials were deposited.

Rifting proceeded to spreading (starting 750 Ma). The northern rift deepened, with deposits of (possibly) up to 17,000 m thick sedimentary and volcanic rocks, but the continental crust did not rupture. The crust under the southern rift broke, with formation of the narrow (Red Sea-like) Khomas Ocean between the Congo and Kalahari Cratons, complete with oceanic crust, a mid-oceanic ridge and accumulations of turbidites. In the west, the Adamastor Ocean opened, separating the Rio de la Plata Craton from the Congo Craton. During rifting and spreading, the Otavi, Swakop, Witvlei and Zerrissene Groups were deposited. Two Snowball Earth events took place during the rifting and spreading: the Sturtian-equivalent (750 Ma, when initial rifting evolved into spreading) and Marinoan-equivalent (635 Ma) episodes, resulting in the Chuos and Ghaub Formations respectively. Sequences of tillites, cap carbonates and banded iron formations are seen as evidence of these extreme, equator-ward global glaciation events.

The Kaoko Belt formed a passive margin on the eastern side of the newly formed Adamastor Ocean. Around 650 Ma, on the west of the Adamastor Ocean, subduction, metamorphism and emplacement of granite formed the precursor of what would become the Coastal Terrane of the Western Kaoko Zone. Erosion of the cratons supplied sediments that accumulated in the oceans, possibly up to 10 km thick in the Khomas Ocean.

The spreading stopped (about 600 Ma) and reversed. West of the Adamastor Ocean, the São Francisco and Rio de la Plata plates amalgamated (630-620 Ma). The Adamastor and Khomas Oceans closed, followed by continental collision between the Rio de la Plata and Congo Cratons along the northwestern Kaoko Belt (595-550 Ma) and between the Congo and Kalahari Cratons along the central-Namibian Damara Belt (542 Ma). The Kalahari Craton subducted under the Congo Craton (until 480 Ma), which, in turn, subducted under the Rio de la Plata Craton. The Coastal Terrane (Kaoko accretionary prism) were added onto the rest of the Kaoko Zone.

The continental convergence and amalgamation resulted in the formation of Gondwana during the Damara Orogeny. Sediments from the ocean floor between the two cratons were folded, heated and pushed up, building the north-south-trending alpine-type Damara Mountain Belt (550 Ma), the deep root zone of which constitutes the intensely deformed granites and gneisses of today's Damara Supergroup. A foreland basin developed to its south, forming a warm, shallow shelf sea with carbonate reefs, into which erosion products of the high mountains accumulated as Nama Group sediments (starting 555 Ma). These dominate the present southern Namibian geology.

Magmatic bodies rose and intruded granites and pegmatites into the piled-up sediments of the Damara Belt (565-470 Ma). Erosional products from the Damara Mountains were deposited as the Mulden Group on the Otavi Group carbonates.

Today, the coastal (Kaoko) and intracontinental (Damara) arms of the Damara Orogeny underlie much of northwest and central Namibia. The Damara Orogeny is part of the much larger Pan-African orogenic cycle of opening and closing of several large oceans and collisions of a number of cratons that culminated in the formation of the **Gondwana**, the south-western part of the old supercontinent Pangaea.

Over the next 250 million years, erosion wore down the Damara Mountains as plate tectonics moved Gondwana slowly south. Extensive planation of the Gondwana surface took place before onset of the next development phase, represented by the **Karoo Supergroup**, which started with the Permo-Carboniferous glaciogenic Dwyka Group (320-270 Ma). When Gondwana straddled the South Pole, vast ice sheets (analogous to the present Greenland and Antarctic ice sheets) covered much of southern Africa for about 50 million years. The ice sheets ground down high mountain ranges, glaciers carved deeply into the Damaran bedrock (e.g. the valleys of the Kunene, Engo, Munutum, Klein Nadas, Sechomib, part of the Hoarusib) and left glacial debris (tillites) (300 Ma).

As Gondwana drifted northwards again, ice sheets and glaciers melted and deposited glacial, lacustrine and fluvial sediments in deep freshwater lakes, along rivers and deltas, in marine and coastal areas (270-200 Ma; Permo-Triassic). An aborted rifting phase cut deep ENE-WSW basins across central Namibia.

Further warming of the climate triggered extreme aridity and formation of an immense inner continental desert with vast sand seas. These were eventually petrified into aeolian sandstones of the Etjo Group (200-180 Ma; Upper Triassic to Lower Jurassic).

During the period 300-135 Ma, the Damara Mountain belt was weathered and eroded to a few remnants of smaller mountain chains and granite inselbergs, the land surface underwent peneplanation and glacial, fluvial and aeolian sediments were deposited in basins. The large inland Karoo basin of southern Africa accumulated more than 10,000 m of sediments and lavas.

Gondwana started **breaking apart** around 180 Ma (Lower Jurassic), with extensive volcanism over southern Africa. Around 135 Ma, deep-seated Proterozoic crustal weaknesses – that were reactivated during the Pan-African (Damara) metamorphism and again with initiation of the break-up of Gondwana – allowed violent fracturing of the crust and injection of pressurised magma into subsurface fractures. These formed NE-SW trending Giant Dyke Swarms which were exposed once softer country rock had been eroded away, for example the Henties Bay-Outjo Dyke Swarm.

Around 132 Ma (Lower Cretaceous), the rise of the Tristan mantle plume – a local heat source in the upper mantle (> 100 km deep) – and tectonic stresses caused crustal thinning. Deep crustal faults split open in kilometres-long fissures, from the Etendeka to deep into Brazil, from which an estimated 6,340 km^3 hot, fluid dark grey continental basalts erupted for 1-2 million years. Concurrently, quartz latites erupted from central shield volcanoes (like Messum, Cape Cross, Brandberg and others in South America) in hot, fluid, reddish to pale grey, silica-rich pyroclastic flows, travelling 650 – 700 km from the vents. These ignimbrites (Sarusas, Terrace Bay, Agate Mountain area) are amongst the largest ash flows in the global geological record, covering up to 170,000 km^2. The Sarusas quartz latite may have had a volume of up to 11,000 km^3. Horizontal layers (‘ traps’) of volcanic flows accumulated 1-2 km thick, to form the Paraná-Etendeka Large Igneous Province, which today covers about 78,000 km^2 in the Etendeka volcanic Plateau of Namibia and more than 1.5 million km^2 in the Paraná lava field of Uruguay, southern Brazil, eastern Paraguay and northern Argentina. Faulting parallel to the present coastline and gentle eastward tilting of the Etendeka lavas took place. The break-up of Gondwana proceeded from south to north over the period 139-128 Ma, the Atlantic Ocean intruded between South America and Africa, and the western coastline of Namibia was formed.

The Damaraland intrusive igneous complexes (plutons) formed more or less in the same period (137-125 Ma; Lower Cretaceous), with their magma crystallizing slowly several kilometres below the surface. These 32 ring complexes and diatremes (volcanic pipes) extend northeast from across central Namibia and include the Brandberg, Erongo, Spitzkoppe, Cape Cross, Doros, and Messum complexes. They are not associated with a mountain-building event, but with tensional stresses caused by the break-up of Gondwana and the Tristan hotspot.

The break-up caused isostatic up-warp of the Great Escarpment around the edge of southern Africa, forming the Kalahari Basin in the interior of the subcontinent.

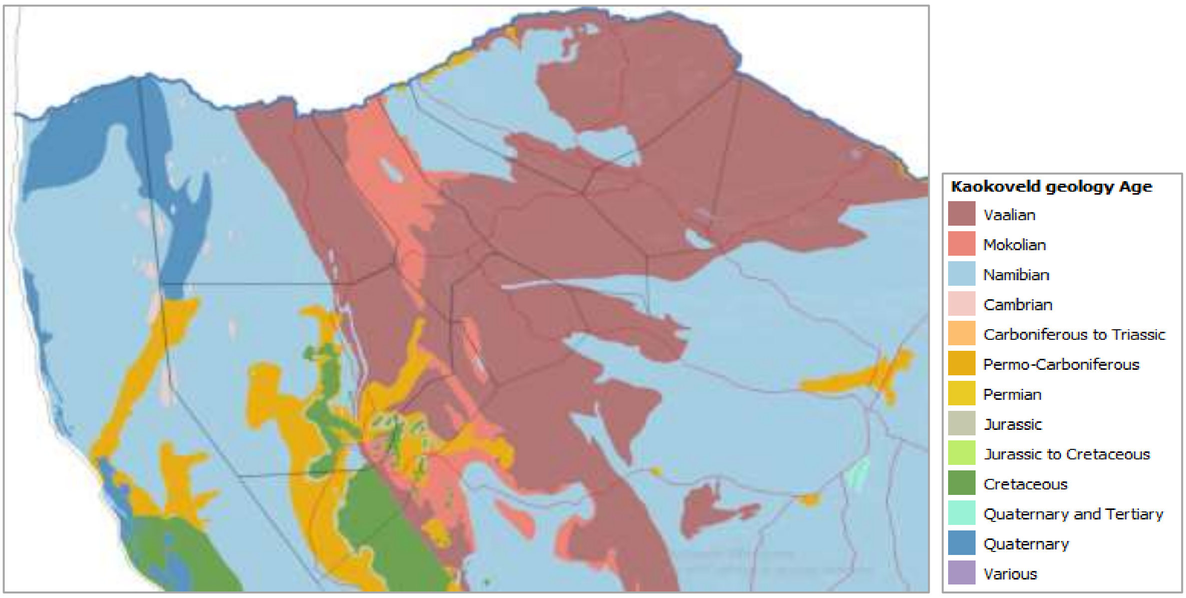
Erosion proceeded throughout the Cretaceous and Palaeogene and into the Miocene Epoch of the Neogene Period (135 – 65 Ma) and wore the landscape down to the Early Tertiary African Erosion Surface), which is lower at the coast (the 'Namib bevel') than in the interior due to the difference in base levels. The peneplanation removed great thicknesses of Karoo age rock and, in places, planed down to Precambrian basement complexes. The erosional debris were deposited on the coastal plain and offshore on the continental shelf.

The **Kalahari** and **Namib Groups** were deposited from 70 Ma. Isostatic uplift of southern Africa (70-60 Ma; Miocene) of up to one kilometre ended the African erosion phase and initiated the Post African I and II erosional phases, during which fluvial deposition took place.

The Namib Group includes marine, fluvial and aeolian deposits. Much of these were derived from Orange River and other river delta sediments, deposited on the continental shelf, transported north by longshore drift, placed onshore by wave action and transported inland by wind.

Arid conditions started around 43 Ma (Eocene epoch) and peaked by 15 Ma (Miocene epoch) with the development of the Antarctic Ice Sheet. Dune fields covered much of the coastal plain and is preserved as the Tsondab Sandstones.

More humid conditions followed up to about 10 Ma, during which gravels were deposited and river terraces formed. The onset of the Benguela Current and its upwelling system, and the return of aridity (5 Ma; Pliocene epoch) initiated formation of the present Namib Desert, as well as widespread calcretisation. Ice ages in the Quaternary (< 2.58 Ma) lowered sea levels and thus base levels for erosion. This caused intensification of erosion, further eastward retreat of the Escarpment and deeper incision of rivers into the coastal plain.



Geological age of north-western Namibia (derived from GSN data)

GEOMORPHOLOGICAL DEVELOPMENT

Desert landscape evolution is mainly about formation and extension of pediments, to end in pediplains – low-gradient and relatively featureless plains. Thus, the geomorphological development of the SCNP is described by Watson and Lemon (1985) as cyclic pediplanation of the series of land surfaces originally suggested by King (1962), namely Gondwana (Jurassic age), post-Gondwana (Cretaceous), African (Late Cretaceous – Early Tertiary), Post-African (Late Tertiary) and Congo / Zaire (Quaternary {Pleistocene}, Holocene and Recent). Each new cycle cut into features of its predecessor. A series of step-like erosional surfaces (platforms) formed, with the highest being the oldest. The current coastal platform represents the Early Tertiary African Surface slightly modified by minor fluvial deposition thereon and the younger erosion phases.

Pediplanation starts with vertical (epeirogenic[1]) uplift of the land, followed by erosion of the steep margins of the landmass through both sheet flooding and lateral erosion by streams. As mountain fronts erode, their cliff faces and the steep wash slopes below retreat in a process known as backwasting. At the same time, a relatively thin layer of material from eroding uplands is deposited, mainly by laminar sheet flow of water, over bedrock on the gentle toe-slopes. As a result, these pediments at the bases of receding escarpments, grow at the expense of the mountain fronts.

The Great Escarpment lies to the east of the Namib Platform and separates it from a mountainous hinterland at ± 900–1300 m height (Partridge and Maud, 2000). The Great Escarpment is a result of continental uplift and denudation following the break-up of Gondwana and opening of the South Atlantic during the early Cretaceous. Weathering-resistant rocks of the Precambrian Damara Supergroup of the Escarpment, for instance granites and gneisses, form outliers rising slightly above the coastal plain that had in some places been eroded to inselbergs. Pediments extend from mountain footslopes towards the coastal plain: gently sloping erosional landforms consisting of bedrock with a relatively thin covering of alluvial and / or colluvial material. The landform pattern of the coastal platform is largely determined by exposed bedrock aligned more or less parallel to the coastline.

The coastal plain's elevation varies from sea level to about 600 m, and the width between 30 and 50 km, with a very gentle slope of 1-2 °. Bedrock weathering is controlled by the geological structure and rock type. It is partly exposed and partly covered by dunes, sand sheets, sand streaks, river terrace deposits, thin sands and gravels.

The strip adjacent to the ocean is characterised by sand-, gravel- and pebble beaches, exposed bedrock headlands, northerly-extended sand spits and raised beach terraces. This strip shows the influence of Pleistocene and earlier sea level changes: six marine transgressions had been identified along the southern Namibian coast.

Intrusive granites form inselbergs, either as broad, low exfoliated domes or jointed castle koppies known as *tors*.

Geomorphic and soil-forming processes in arid environments differ substantially from those in more humid zones. Streamflow is irregular, intermittent, ephemeral and often endorheic – ending in inland depressions, viz. playas and salinas (salt pans). Chemical weathering is inhibited by absence of moisture, so physical weathering processes dominate. Unconfined sheet floods and mass-wasting are the primary erosional

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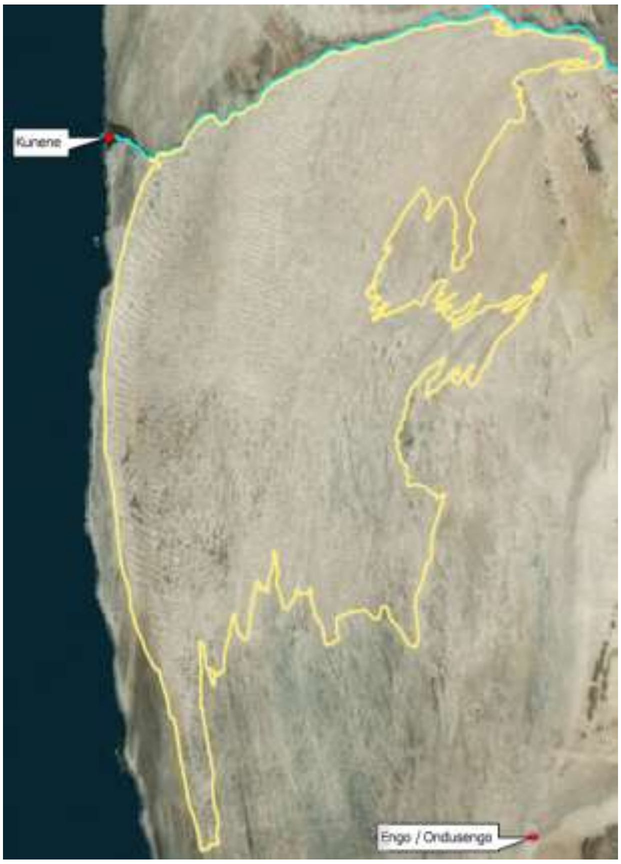
[1] Epeirogenic movements refers to slow uplift / upwarp of large, stable interior blocks of continents (cratons), causing little deformation of fracturing. It is in contrast to orogenic movements at plate margins caused by compressional or tensional forces that result in intense folding, thrusting, faulting and uplift of narrow belts

processes, more even than aeolian processes. Aeolian sand blasting of bedrock outcrops has been intense in the north.

SAND DUNES AND SAND SHEETS

The SCNP has two extensive dune fields, namely the Skeleton Coast Erg and the Kunene (or Northern Namib) Erg, as well as sandy plains and isolated dunes.



Kunene Dune Field, excluding adjacent sandy plains (BA/MC)

Skeleton Coast Dune Field (BA/MC)

SKELETON COAST ERG (DUNE FIELD)

The first isolated barchan dunes appear north of the Huab River. The Skeleton Coast Dune Field stretches from ± 20.36 °S, halfway between the Koigab and Uniab Rivers, to the Hoarusib River at ± 19 °S over 165 km, covering about 1,800 km^2. A 1-2 km wide band of barchans continue beyond the Hoarusib River Mouth for another 25 km to the Khumib (Garzanti *et al*., 2014). The western margin of the dune field is 2-5 km inland, parallel to the coast. The dune field's width varies between 6 and 22 km (Garzanti *et al*., 2014). It is interrupted by the Uniab and Hoanib Rivers and occasionally breached by large floods in some of the other ephemeral rivers. In the south, a deflation surface of exposed schists, granites and basalts are, in places, covered by a thin layer of wind-driven sand and isolated 3-10 m high barchan dunes. A larger sand supply further north allows the barchans to coalesce into compound transverse dunes of up to 50 m height and, in places, a longitudinal dune wall of up to 80 m height, with the slip face to the east. 3-10 m high barchan dunes march along the eastern, downwind margin of the dune field (Lancaster, 1982; Garzanti *et al*., 2014). NW-SE trending transverse dunes dominate south of the Hoanib. North of the Hoanib, their orientations swing more WNW-ESE and a few complex linear dunes appear.

NW-SE trending transverse dunes of the Skeleton Coast Dune Field (BA)

Complex linear dunes at the eastern edge of the Skeleton Coast Dune Field between the Hoanib and Hoarusib (NNW-SSE orientation) (BA)

KUNENE ERG

The Kunene (or Northern Namib) Dune Field, covering about 1,600 km^2, stretches from ± 17.86 ^{o}S, north of the Engo River, to the Kunene River at ± 17.16 ^{o}S, where the river abruptly stops the northward migration of dunes.

The erg is underlain by igneous and metamorphic rocks of the Swakop Group (570-900 Ma) that had been worn down by erosion over millions of years into a peneplain and subsequently covered by sand.

The most extensive dune type is transverse dunes in the centre and northwest of the dune field, with average ridge lengths of about 3 km. Their orientations change from N-S to more E-W as one progresses north-eastwards across the dune field.

Sand streaks and trains of barchans, with average lengths of 560 m and average width between horns of 193 m, occur in the southwest of the dune field. Closer to the coast, barchan dunes are oriented toward the north, while they point more northeast further inland.

Linear dunes, with average lengths of about 7 km, are found in the east and northeast of the dune field. Their shapes are more complex that those of the Kalahari Desert. They are to some extent overlain by transverse dunes, which indicates that they are older than the transverse dunes. Their orientation also swings around more E-W as one moves northeast across the dune field (Goudie, 2007).

Complex linear dunes south of the Kunene River, with their orientation swinging more E-W towards the northeast of the dune field (BA)

Transverse dunes overlying linear dunes (BA)

PROVENANCE OF DUNE SANDS

The sand in the Skeleton Coast and Kunene Ergs were originally deposited over millennia by the Orange River and, to a lesser extent, by the Swakop River on the broad continental shelf, laid bare and dried out during periods of sea level regression, and subsequently transported onto land by wind. This is an ongoing process. Studies by Garzanti *et al*. (2014) showed that sand derived from the Orange River is carried by strong swell-driven longshore currents more than 1,750 km to the northern Namibian and southern Angolan shores. This is the longest cell of littoral sand transport recorded on Earth.

Accumulations of garnet lend a reddish hue to dunes (AD)

The dark colours of magnetite and ilmenite accentuate ripples in sand (MC)

Long-distance sand contribution from the Orange River is dominant, but not exclusive. Studies of river, beach and dune petrography, mineralogy, geochemistry and geochronology indicate that more than 80 % of the Skeleton Coast's dune sands are supplied by the Orange River and most of the remainder by the Swakop River, namely sediments of eroded metamorphic and granitic rock of the Damara Orogen. The contributions from Etendeka lavas and other rivers draining the Damara Supergroup are minor (Garzanti *et al*., 2014; Garzanti *et al*., 2017). Skeleton Coast dunes have less volcanic rock fragments, less pyroxene, and more staurolite, garnet, tourmaline and amphibole than the Namib Erg (Garzanti *et al*., 2017).

Sand and heavy minerals are moved onshore by strong, sustained wave action. Further sorting by size and density takes place, mainly on the upper part of the eulittoral zone (high tide wave zone; swash zone) (Van Gosen, *et al*., 2014). Strong winds remobilise the fine sands and heavy minerals to form dunes landwards of beaches. Selective removal of larger, low-density grains by wind leaves behind smaller, denser grains and thus progressive heavy-mineral enrichment of beach and dune sands (Komar, 2007).

DUNE FORMATION

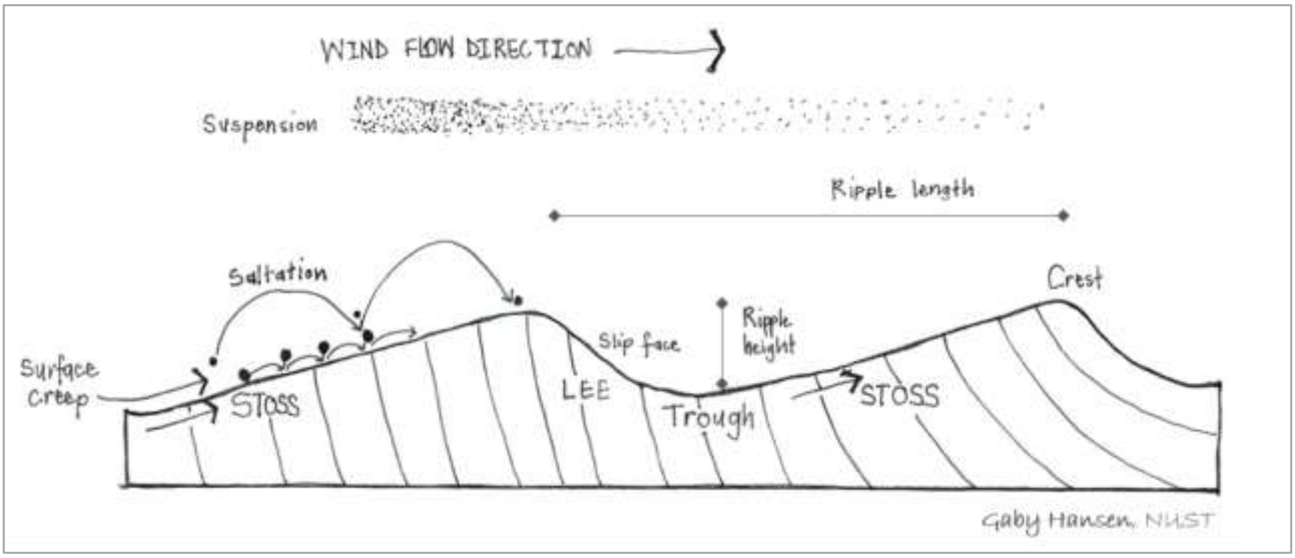
The wind speed and size, density and shape of minerals determine how transported material is winnowed and sorted. Aeolian (wind-borne) deposits are rather uniform in terms of grain size.

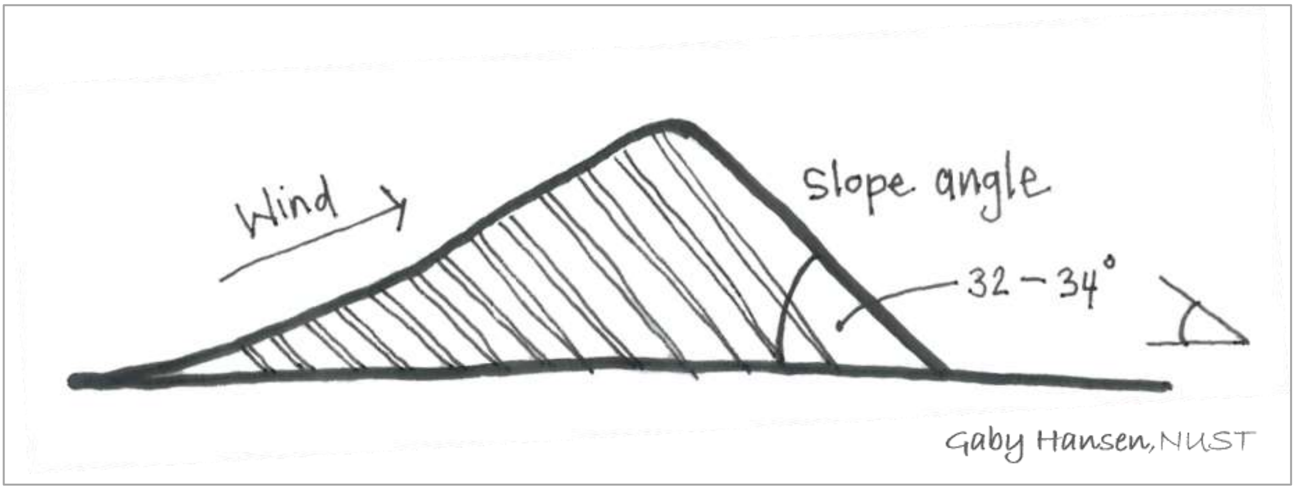
Wind transports weathered material in three ways: suspension, saltation and surface creep: The finest particles – silt and clay – are *suspended* when blown high into the air. Silt-sized dust remains aloft over long distances and fine, plate-shaped clay minerals and micas stay suspended even longer. Nutrients in dust from sandstorms over the Namib fertilise the Southern Atlantic (NASA, 2015; Dansie *et al*., 2018). Namib soils contain less silt and clay than the Sahara, so severe dust storms are less frequent and intense. The heaviest mineral particles – pebbles, coarse gravel, heavy minerals, coarse sand grains – are pushed and rolled along the surface in a process known as *surface creep*. Creep accounts for about 4% of a sand grain's movement. Fine gravel and sand are bounced along by a process known as *saltation*. Saltation accounts for about 96% of a sand grain's movement. It is initiated by small differences in air pressure caused by tiny irregularities of the surface – the Bernoulli effect. Every grain that lands knocks more grains into the air. As more sand grains become airborne on the stoss (upwind) side of the small irregularities and deposition happens on the lee side that is slightly more protected from the wind, self-organised ripples develop perpendicular to the wind direction. The average distance a grain bounces is the width between the crests of two sand ripples.

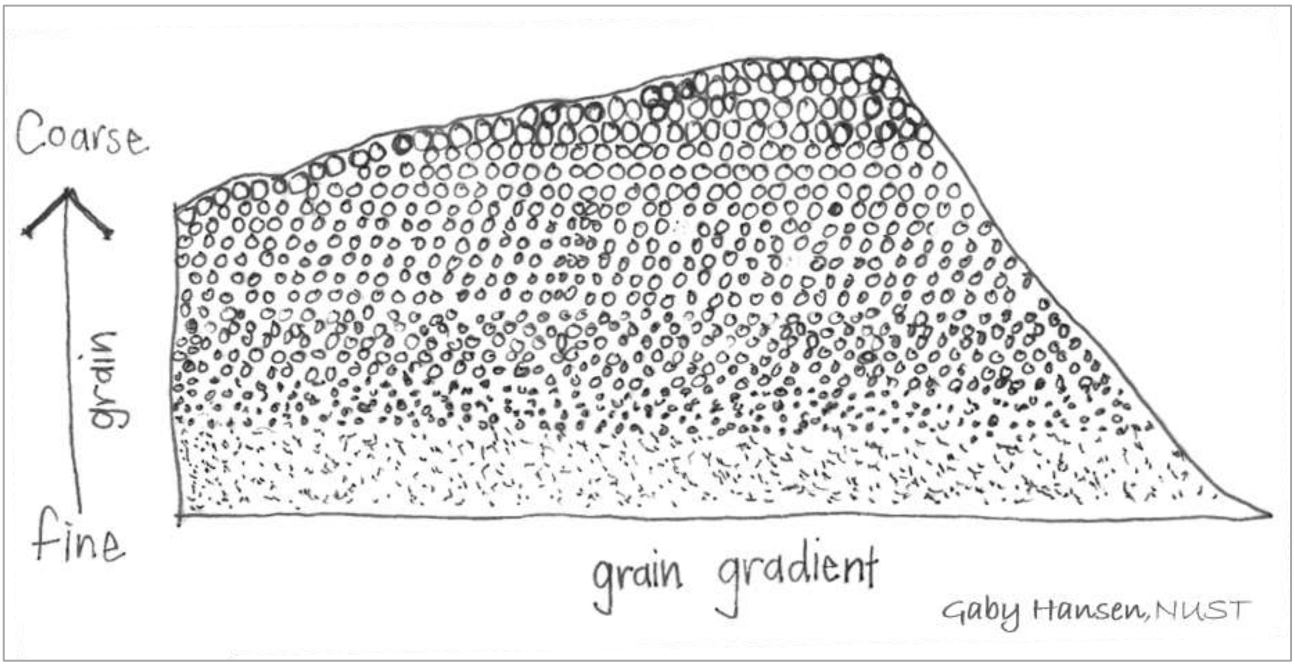
Dunes form in a similar process, on a larger scale. Transported sand settles as a sand patch in the lee (downwind side) of an obstacle (vegetation, rock, sand ripple) and grows in size. The transport capacity of wind decreases as it drives sand grains up the windward side of the growing dune, causing sand to be deposited before reaching the dune crest.

As the dune grows, the lee slope steepens until it exceeds the 'angle of repose', which is 32-34^{o} for dry sand. Gravity overwhelms the shear strength of the sand and it avalanches down the lee side, forming a slip face. Over time, the dune migrates downwind by deflation of the windward side and deposition on the lee side.

Smaller particles blown across the dune crest tend to travel farther than large ones. The result is that aeolian sand deposits show a coarsening-upward sequence.

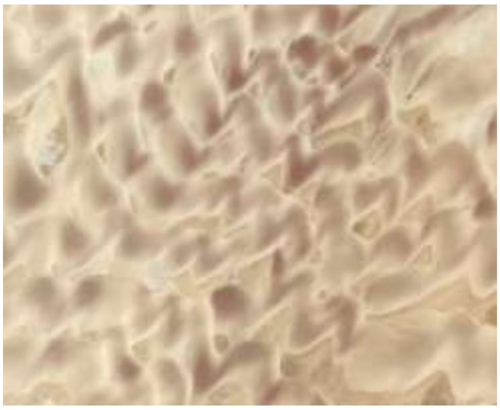






Dune formation (top); Angle of repose (middle); Grain gradient (bottom) (GH)

Larger, heavier particles tend to accumulate on the windward slopes, where minerals such as magnetite, ilmenite, basalt and garnet impart a dark grey-blackish and reddish sheen to dunes.



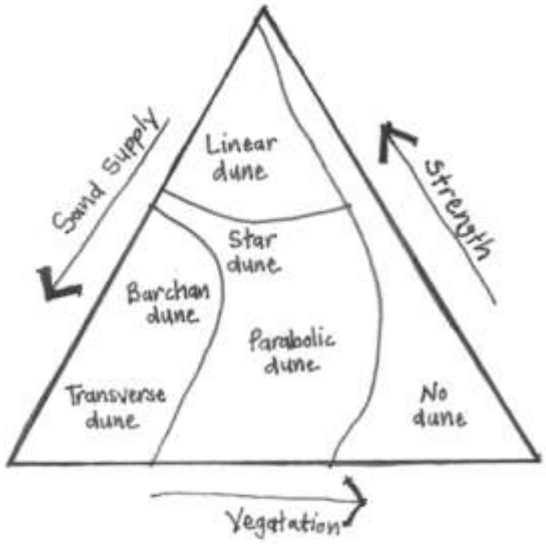
Dunes with garnet accumulation on the upwind side (BA)

Basalt grains accumulate on sand ripples and against obstacles (MC)

Windblown sand generally moves northward in the SCNP, but low bedforms are also deposited in riverbeds – such as those of the Hoanib and Hoarusib – up to 50 km east of the main dune fields by the eastward funnelling effect of these deeply incised river valleys (Garzanti *et al*., 2014).

DUNE TYPES

Dune type depends on wind strength and direction, sand supply, soil moisture and the presence or absence of vegetation.



Relationship between dune types and wind strength, sand supply and presence of vegetation (GH)

Barchan dune (GH)

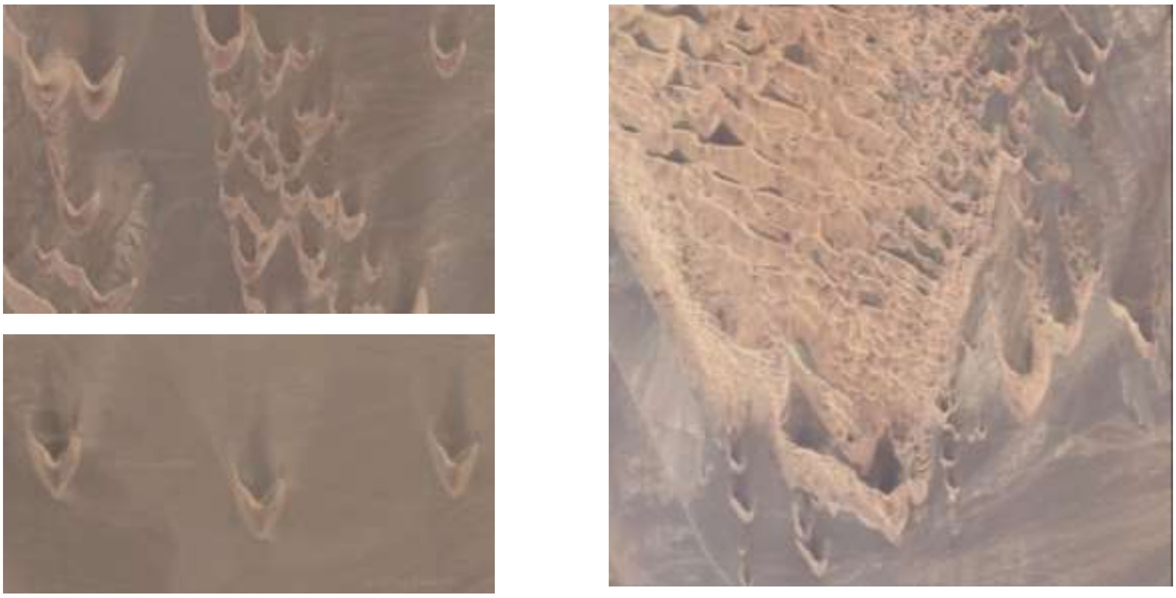
BARCHAN (crescent) DUNES

Highly mobile crescentic barchans develop where wind is uni- directional or narrowly bi-directional, sand supply low, precipitation very low and the ground surface fairly level. The sand advances at a rate that is roughly inversely proportion to the height of the crest, which means that the dune flanks move faster than the centre. Barchans thus move in the direction of the forward-facing horns that outstrip the bulk. The slip-face is on the forward-facing, concave side.

Short trains of barchans appear north of the Huab, Hoanib and Hoarusib River mouths and the lower Khumib, Sechomib, Nadas and Engo Rivers and peter out where there is an insufficient supply of sand. Barchan dunes coalesce into transverse dunes at the southern and eastern margins of both the Skeleton Coast and Kunene Ergs (Miller, 2008, p. 25-52).

Small groups of barchans north of the Huab (BA)

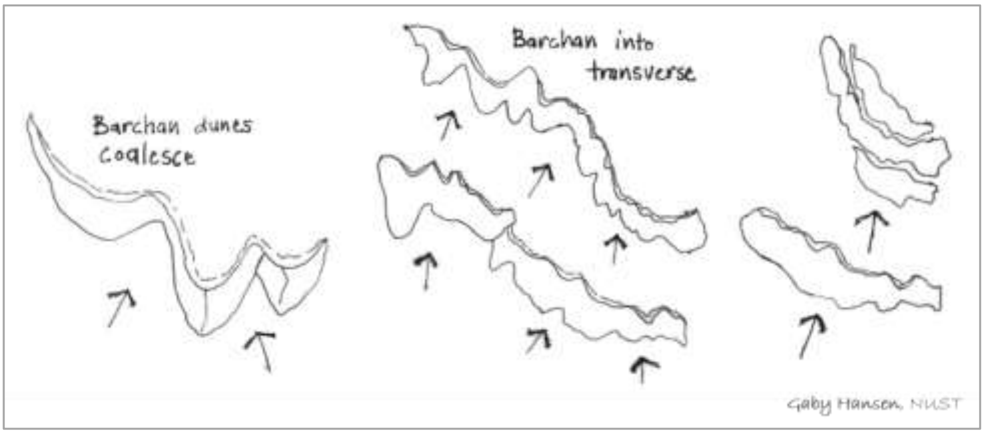
Side view of a barchan dune (MC)



North of the Koigab and Engo Rivers, individual barchans join up to form transverse dunes (BA)

TRANSVERSE DUNES

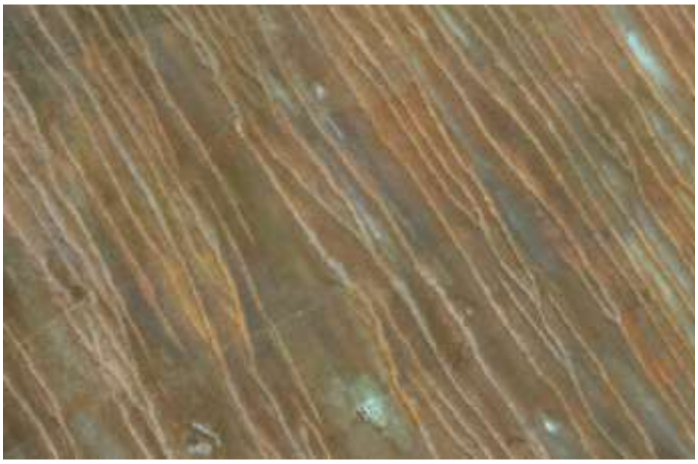
Where sand is more plentiful and wind uni-directional, barchans join up into transverse dunes. Their long axes are more or less perpendicular to the wind direction.



Transverse dunes are the most abundant dune type in the Skeleton Coast and Kunene Ergs. Their orientations are mainly NW-SE to WNW-ESE.

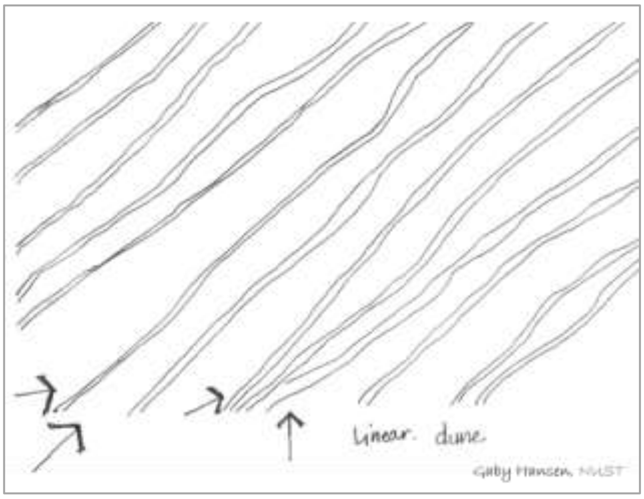
Transverse dunes (BA)

LINEAR (longitudinal; seif) DUNES



Linear dunes from the Kalahari (BA)

Linear dunes are straight or slightly sinuous symmetrical sand ridges. They form where sand is abundant and wind bi-directional, with one wind direction somewhat dominant. Wind pressure is more or less the same on both sides of the dune. Linear dunes are aligned lengthwise between the two wind directions. Inter-dune areas may be gravelly or sandy.



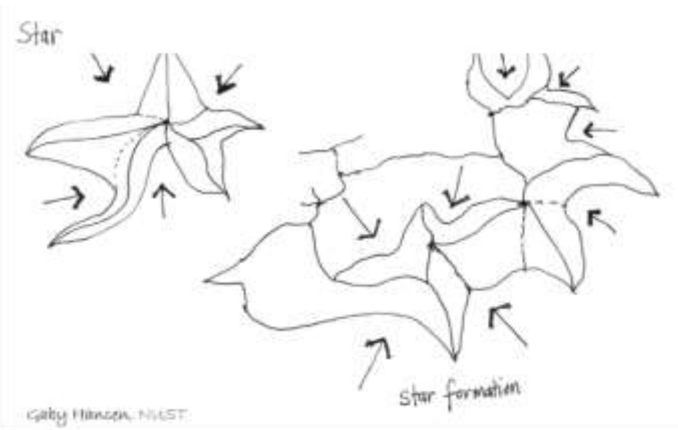
Linear dunes (GH)

The Skeleton Coast does not have simple linear dunes like those of the Kalahari, but rather linear dunes overlain by more recent transverse dunes.

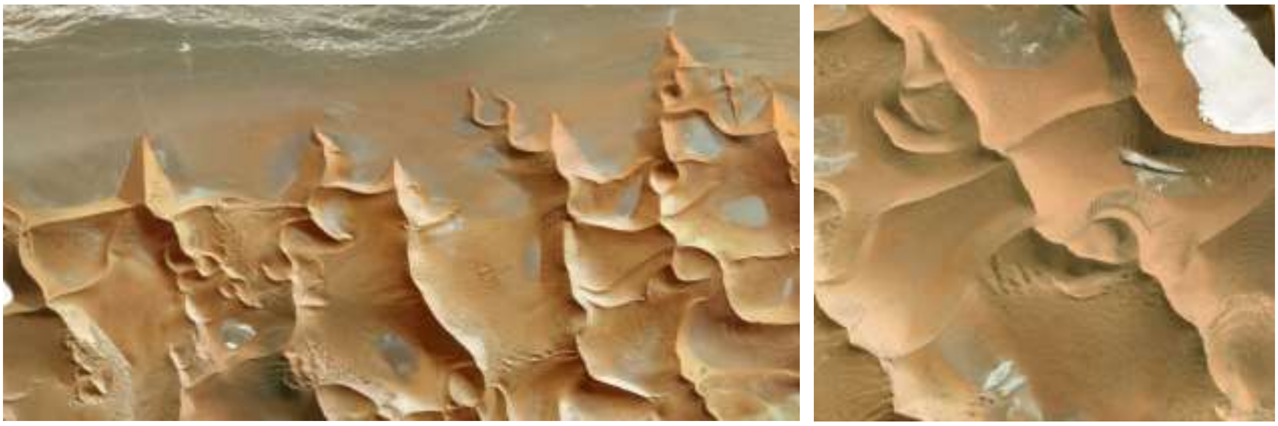
Complex linear dunes overlain by more recent transverse dunes, in the Kunene Erg (BA)

STAR (pyramidal; ghourd) DUNES

Star dunes form where wind is multi-directional – at least from 3 directions – and sand plentiful. They have sharp points and ridges, and at least 3 slip-faces. Splendid examples can be found in the Namib Erg around Sossusvlei. They have not been observed in the Skeleton Coast National Park.



Star dunes (GH)



Star dunes from Sossusvlei area (BA)

PARABOLIC DUNES

In parabolic dunes, the depositional lobe (the convex 'nose') on the downwind side leads forward movement, while the trailing 'arms' bracket a deflation basin. The slip-face is on the downwind, convex side of the parabola. Parabolic dunes form where dampness or vegetation suppress sand motion and slow down the arms, while the centre advances downwind (Goudie & Wells, 1995; Goudie, 2012; Goudie, 2011). They are not known to occur in the Skeleton Coast National Park.

A few poorly-defined parabolic dunes march across the Kuiseb Delta (BA)

STOSS AND SHADOW DUNES

Stoss dunes are low-angled sand ramps forming upwind of obstacles such as rock outcrops, while thicker shadow dunes, usually with well-developed crests, form in the lee of obstacles.

Shadow dune in the lee of a hummock (MC)

VEGETATED HUMMOCK DUNES

(Nebkhas / Nabhkah, Shrub-Coppice Dunes)

Plants (e.g. *Salsola*, *Zygophylum*, !Nara, *Arthraerua*) disrupt wind flow, causing deposition of sand on the lee side. Hummocks grow over time as plants grow upwards to escape the encroaching sand. In the fog zone, the exposed positions of these plants help with fog-harvesting. Plant material and windblown detritus accumulate, creating small ecosystems with characteristic desert wildlife such as insects and reptiles, some of which are endemic. Vegetated hummock dunes occur along the entire coastline in a long, intermittent chain parallel to the coast, on the coastal platform east of the littoral zone (above the high-water mark). This provides a discontinuous S-N migration route along the coast.

Birth of a hummock dune (MC)

Plant-covered hummock dune (DC)

Detritus accumulation (MC)

A !Nara hummock (AM/RB)

Salsola hummocks (BA)

Salsola hummocks (MC)

RIVERS

PERMANENT RIVER – The Kunene

The Namibian Skeleton Coast National Park and Angolan Iona National Park are separated by the only perennial river in the study area: the Kunene River or Rio Cunene, whose valley forms a linear oasis in the arid environment.

The Kunene is approximately 1,050 km long, with a catchment of 106,500 km^2, of which ± 92,400 km^2 lies in Angola and 14,216 km^2 in Namibia (Paterson, 2007; Midgely 1966; Morant, 1996; Greenwood, 1999; Strohbach, 2008). On average, about 5.5 km^3 water flows down the Kunene annually (Robertson *et al*., 2012) at a maximum discharge rate of about 1,000 m^3/s and transporting around 9 million tons of sediment (Garzanti *et al*., 2017).

Namibia has a hydropower scheme in the Kunene River near Ruacana and is considering the installation of another hydropower scheme at Epupa, further downstream. Such a hydropower scheme would require the flooding of a large part of the Kunene valley. The river is impounded at six places upstream of Ruacana.

The Kunene Dune Field's northward migration is halted abruptly by the permanently flowing Kunene River (BA).

KUNENE RIVER MOUTH

The Kunene River Mouth (KRM), also known as the Kunene Deltaic Complex, covers an area of ± 4,130 m^2. The river spreads out into braided channels between sand bars (some vegetated), with a periodically flooded lagoon and mudflats inland of a 2.5 km long linear sand spit on the southern bank and shorter one on the northern bank, that partially block access to the Atlantic Ocean (Paterson, 2007; Greenwood 1999).

The ± 3 km wide mouth of the permanent Kunene River is partially blocked by sandbars. These are cleared periodically by floods from good rains in the Angolan Highlands. The river spreads out in a small delta. The wetland is high in biodiversity (BA).

Kunene River Mouth (BA)

Kunene River Mouth: tip of the sandspit with the Angolan bank visible in the background (MC, January 2020)

Atlantic Ocean on the left, KRM lagoon on the right (MC, January 2020)

Kunene River Mouth: channels, mudflats, vegetated islands (MC, January 2020)

A study by Simmons *et al*. (1993) described the lagoon as being 2.36 km long and 1.60 km wide, and its water up to 10°C warmer than the sea. The tidal range is around 1.4 m. Tidal influence – discernible up to 4 km upstream (NACOMA, 2009) – is mainly by backing up the river water during high tide, when the lagoon is flooded up to 70 cm deep. At low tide, only 10 % (at low river flow) to 50 % (at high river flow) of the lagoon is under water, exposing sand and mudflats (Simmons *et al*., 1993).

The KRM lacks estuarine benthic fauna, marine and estuarine plankton, and marine fish species (Carter & Bickerton 1996; Morant & Carter 1996), which indicates that it acts as a river mouth rather than an estuary, according to the classification of Whitfield (2001). The system is fluvially dominated, with very little evidence of intrusion of seawater at low-flow periods (Carter 1996), provided a permanent minimum flow of ± 20 m^3/s is maintained (NAMANG, 1997). Simmons *et al*. (1993) found that the salinity of water just inside the mouth during peak flow (April) was about 10 times as much as 4 km upstream, but still predominantly fresh. Low flow resulted in a fourfold increase in salinity at that same place, but still only a 10th of that of seawater.

Satellite imagery reveals a 100 km^2 plume of warm, nutrient-rich river water extending NNW into the Atlantic Ocean at the time of peak flow. (Simmons *et al*., 1993). As the freshwater plume from the river mixes with seawater, it creates estuarine conditions in the coastal waters just north of the river mouth. The northward-moving longshore current ensures that fresh river water has very little significant influence on the marine environment south of the mouth. A flood pulse was seen to result in a sediment-laden plume of river water spreading about 650 m south of the mouth, but was cleared by the longshore current within two days (Coetzee, 2020, personal observation).

Sediment plume south of the Kunene River Mouth, with clear seawater visible in the background (MC, January 2020)

Sediment-laden Kunene River floodwater (right) meeting seawater (left) at the tip of the sandspit (MC, January 2020)

Sediment deposited within the mouth is mainly of aeolian origin, from dunes encroaching from the south along the lower stretches of the river, as well as sand blown inland from beaches south of the river mouth. The mouth is never completely closed (Robertson *et al*., 2012). Simmons *et al*. (1993) found that the opening varied between 30 and 80 m at low and high flows, respectively, and mentioned that the opening was more than a kilometre wide in 1975. They also discovered a 275 m northward and 150 m westward migration of the southern sand bar between 1975 and 1992. This points to the highly dynamic nature of sediment deposition and river mouth morphology.

The elongated triangle of the coastal plain between the Kunene Dune Field, Kunene River Mouth and Bosluis Bay is mainly exposed granitic bedrock covered in places by windblown sand, and alluvial gravel deposits, palaeo-beaches and a series of very large Salsola hummocks shaped into linear features by the constant wind.

Linear ridges of large *Salsola* hummocks (MC; BA)

EPHEMERAL RIVERS

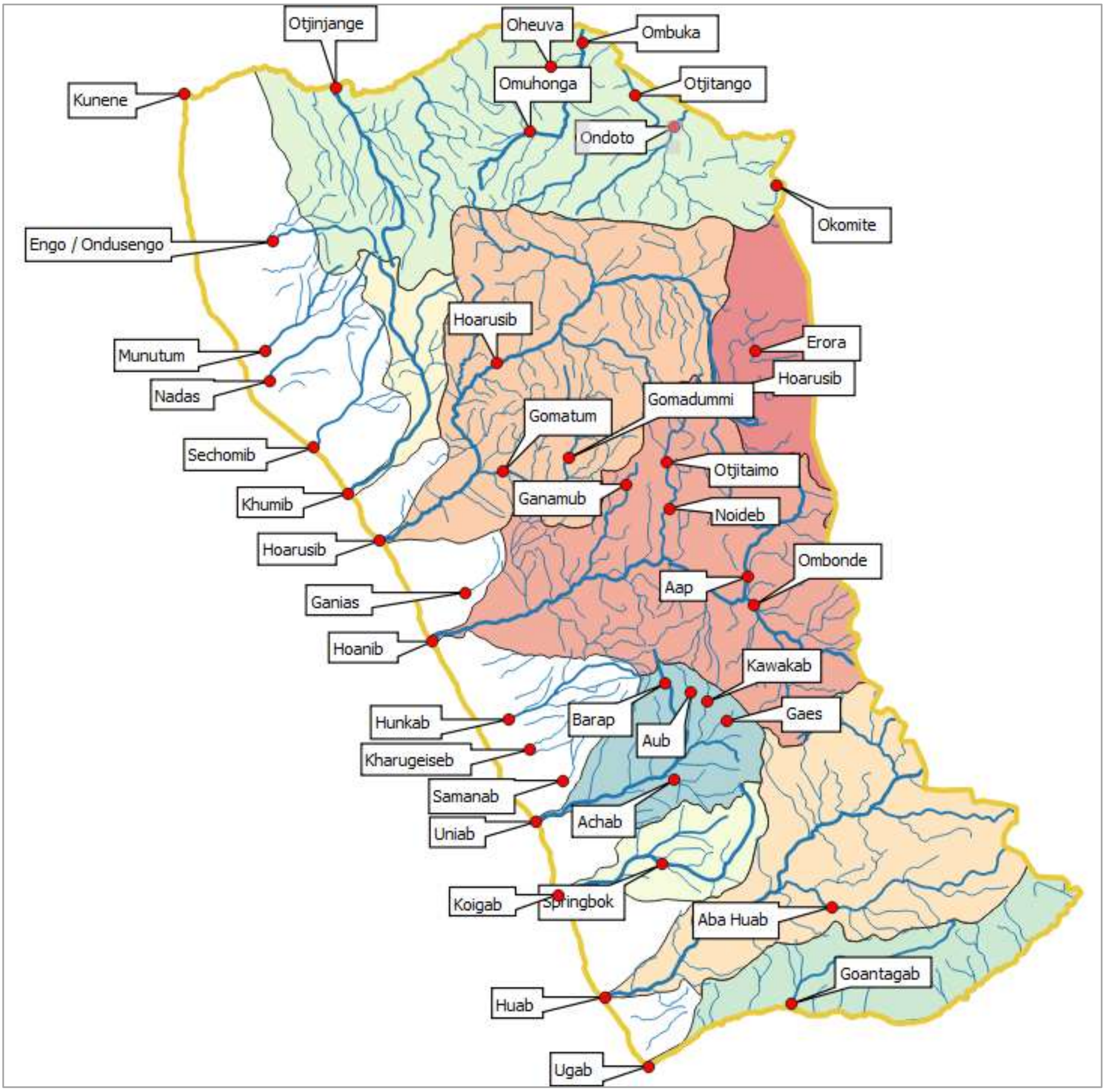
The westward-flowing ephemeral rivers of the SCNP originate in the highlands beyond the escarpment. So-called 'Benguela Niño' events – when the South Atlantic is warmer than normal and the Inter-Tropical Convergence Zone progresses further south – brings higher than normal rainfalls to the Kaoko highlands which result in high-magnitude flash floods that last several days (Jacobson *et al*., 1995; Krapf *et al*.,2003; Stollenhofen *et al*., 2014), for example in 1982 and 1995. These floods, usually during February to April, bring water, sediments and nutrients to the lower reaches of the ephemeral rivers. Where ephemeral rivers are barred by dunes, most flows terminate in flat inter-dune playas or active fan systems, forming short-lived wetlands. Only exception floods in those rivers with large catchments can break through the dunes, such as the Hunkab in 1995 (Ward and Swart, 1997) and Hoanib (Stanistreet and Stollhofen, 2002).

The ephemeral (seasonal / intermittent) river courses are of vital importance for supporting biodiversity in the hyper-arid environment. Seasonal flows sustain riparian vegetation and recharge relatively shallow groundwater in alluvial aquifers, which occasionally surfaces as springs and seeps, thereby forming 'linear oases'. These provide habitats and movement corridors for both resident and migrating wildlife, from where they undertake grazing forays into the surrounding desert. Biodiversity is higher along the river courses than in the surrounding desert, even supporting large herbivores such as desert-adapted elephant, black rhinoceros and giraffe. The river courses also provide dispersal routes for plants.

Named rivers, from north to south are the Engo, Munutum, Nadas, Sechomib, Khumib, Hoarusib, Hoanib, Hunkab, Kharugeiseb, Uniab, Koigab, Huab and Ugab. They do not all reach the sea.

The ephemeral rivers of the SCNP provide good examples of the range of interactions between fluvial and aeolian processes. Where fluvial processes dominate, for example the permanent Kunene River and frequently flowing Hoarusib, rivers flush out the sand and interrupt the northward migration of dunes. The sidewalls of ephemeral river channels are frequently draped by sand ramps, which are eroded by floods. These rivers support riparian vegetation and wetlands develop at their mouths.

The Ugab, Huab and Hoanib drain the Damara Orogen, while the Koigab and Uniab drain the Etendeka volcanic province (Garzanti *et al*., 2017).



Ephemeral rivers and the major catchments of the Kunene Region. Only the Kunene flows permanently. (MC)

Rivers of the western Kunene Region on a satellite image backdrop (BA, MC)

ENGO

(alternative names found in literature: ONDUSENGO, ENSENGO, ONDONDODJENGO)

The Engo is a shallow, indistinct drainage line that disappears well before reaching the coast. Part of its course is along a palaeo-glacial valley. The catchment size is 1,010 km^2 (Strohbach, 2008)

Engo (MC)

Engo (AD)

MUNUTUM

The Munutum has a spring and small swamp at Okau, off Cape Fria. The catchment size Is 672 km^2 (Strohbach, 2008).

Munutum spring (left AD; right VdC)

NADAS

The catchment size is 670 km^2 (Strohbach, 2008),

SECHOMIB

(alternative spelling SECHUMIB)

The catchment size is 1 534 km^2 (Strohbach, 2008).

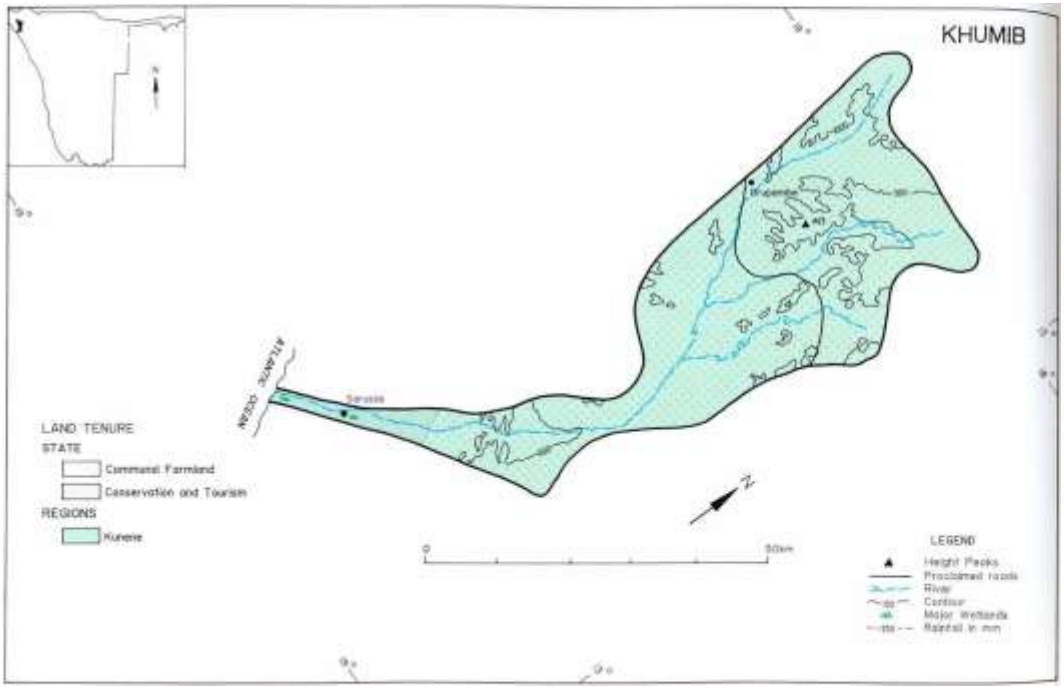
The Sechomib, Nadas and Munutum Rivers widen into small alluvial fans where their valleys leave the hills. (Miller, 2008, p. 25-36, 25-37, 25-38).

KHUMIB

The Khumib has the strongest and most regular flow of rivers north of the Hoarusib and occasionally reaches the sea. It has several springs. The catchment area is 2,308 km^2 and river length is 80 km (Jacobson, Jacobson & Seely, 1995; Strohbach, 2008).

Khumib River valley, as seen from Hartmann's Beacon (MC)

A spring in the Khumib River (MC)



(JJS)

The Khumib is an example of an anastomosing river, with channels repeatedly branching and rejoining around permanent, vegetated alluvial islands (Swart & Marais, 2009). Anastomosing rivers typically have very low gradients and thus low energy, so they carry only the finest sediments.


Braided channels and permanently vegetated alluvial islands of the Khumib (BA)

The Khumib interrupts the northward march of those barchans that manage to cross the mouth of the Hoarusib. The dune field reforms further north of the Khumib, starting out as separate barchan dunes, coalescing into transverse and eventually linear dunes.

The Khumib River's palaeo-delta is similar to that of the Uniab. It has bedded gravels and a coastal cliff, up to 30 m high, that decreases in height to the south and north and is broken by the current river channel (Miller, 2008, p. 25-44).

Khumib River Mouth with fluvial deposits of the palaeo-delta (MC)

Bedded gravels of the palaeo-delta (MC)

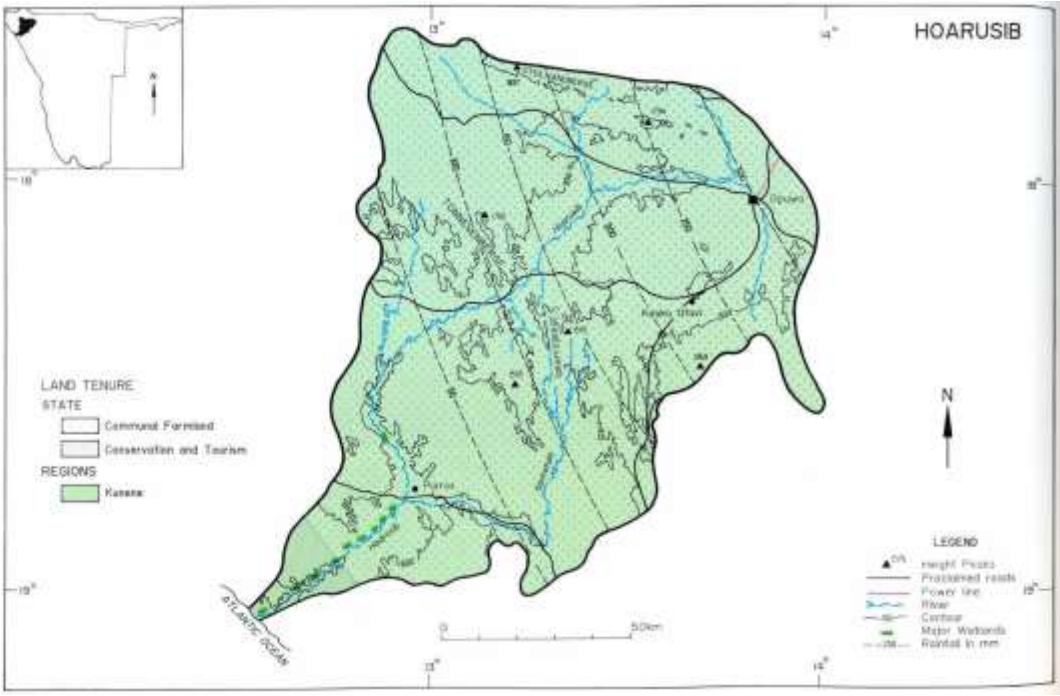
Palaeo-delta of the Khumib River (BA)

HOARUSIB

With its relatively large catchment, the Hoarusib flows regularly and reaches the sea nearly every year. It has several large wetlands. The Hoarusib interrupts the northward migration of the dune field, just as the Kuiseb does for the Namib Sand Sea. A 1-2 km wide band of dunes crosses the river mouth where the river valley is shallower and floods less frequent. In 1995, the river deposited a large volume of sediment in a delta in the sea after more than 200 mm of torrential rain fell in 2 days in its catchment. The delta lasted only a few days under the onslaught of waves (Swart & Marais, 2009; Goudie & Viles, 2015). The catchment area is 15,237 km^2 and river length is 300 km. (Jacobson, Jacobson & Seely, 1995; Strohbach, 2008).

Lower reaches of the Hoarusib River (BA)

The Hoarusib in flood, near the mouth (JP)



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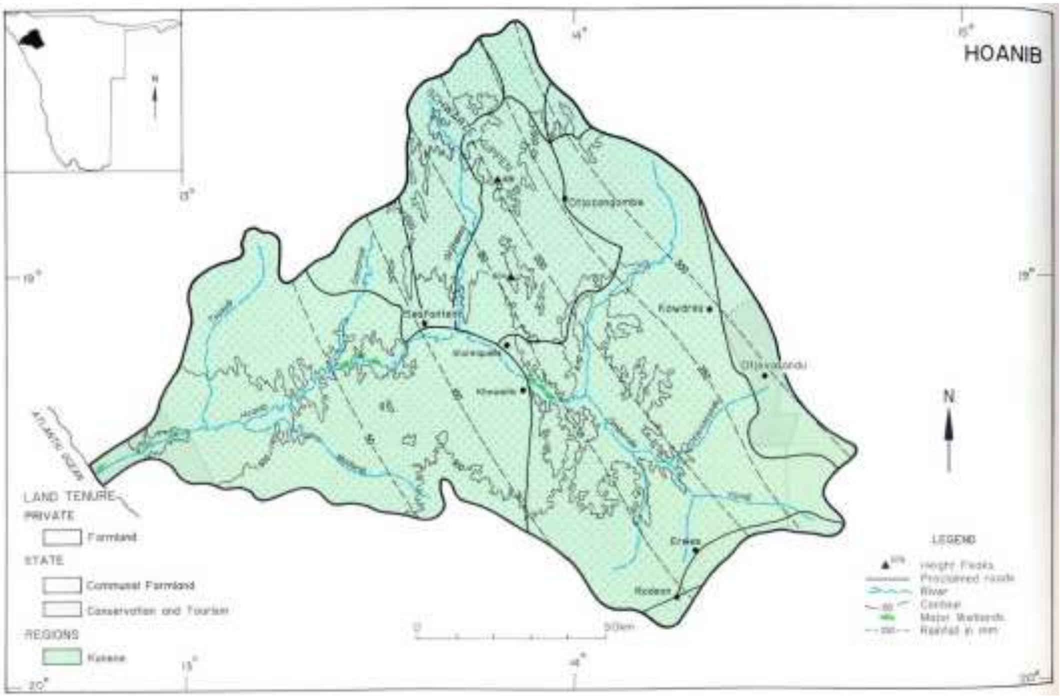
The Hoanib and Hoarusib Rivers have remnants of dissected palaeo-delta gravels that are at least 8 m thick where these are cut by the current river channels. Terrace gravels occur north and south of the river mouths at 80 m and at 20-40 m above sea level. (Miller, 2008, p. 25-43, 25-45)

The Hoarusib is an example of a braided river, repeatedly branching into channels around sand bars and islands – which are inundated during large floods, but mostly retain their positions – and rejoining further on (Swart & Marais, 2009). Braided rivers are typical of areas of low gradient, high sediment load and seasonal floods.


Braided channels of the Hoarusib River (BA)

HOANIB

The larger catchments and closer proximity to the Inter-Tropical Convergence Zone allows for larger, more frequent and longer floods in the Hoanib and Hoarusib. The Hoanib has several springs. The catchment area is 15,761 km^2 and river length 270 km (Jacobson, Jacobson & Seely, 1995; Strohbach, 2008).



(JJS)

The Hoanib has a silt-rich alluvial flood plain that starts spreading out from Uhima, 15 km upstream, to reach its maximum width of 4 km where it encounters the dune belt. (Miller, 2008, p. 25-43, 25-45). As the Hoanib River Mouth is blocked by dunes, flood waters spread out in a permanent, brackish marshy wetland east of the dunes, parts of which are known as Oasis, Auses, Hoaswater and Soutvlei (NACOMA, 2009). Exceptionally large floods allow a build-up of water until the first dune is overtopped at its lowest point. Water spreads out laterally in interdune areas behind the breached dune, and the process repeats itself until the water supply stops or the river breaks through to the ocean. Successive floods have an easier path through the dunes. It only takes a few weeks after a flood for low bedforms to reform from fluvial sediments in the riverbed (Stanistreet & Stollhofen, 2002; Swart & Marais, 2009; Goudie & Viles, 2015).

The lower Hoanib, showing the wetland where the river is interrupted by the dunes (BA)

Brackish wetland at the Hoanib Mouth (MC)

HUNKAB

The Kharugeiseb, Hunkab and other ephemeral rivers with relatively small catchments are blocked by the dune belt, spreading out in terminal playas behind the dunes. In 1995, exceptional rains in the Hunkab catchment allowed the river to break through 15 km of dunes. Another flood in 2000 cleared out some sand that had started to encroach again (Robertson *et al*., 2012; Swart & Marais, 2009; Goudie & Viles, 2015). The catchment area is 593 km^2 (Strohbach, 2008).

Hunkab River, blocked by dunes (BA)

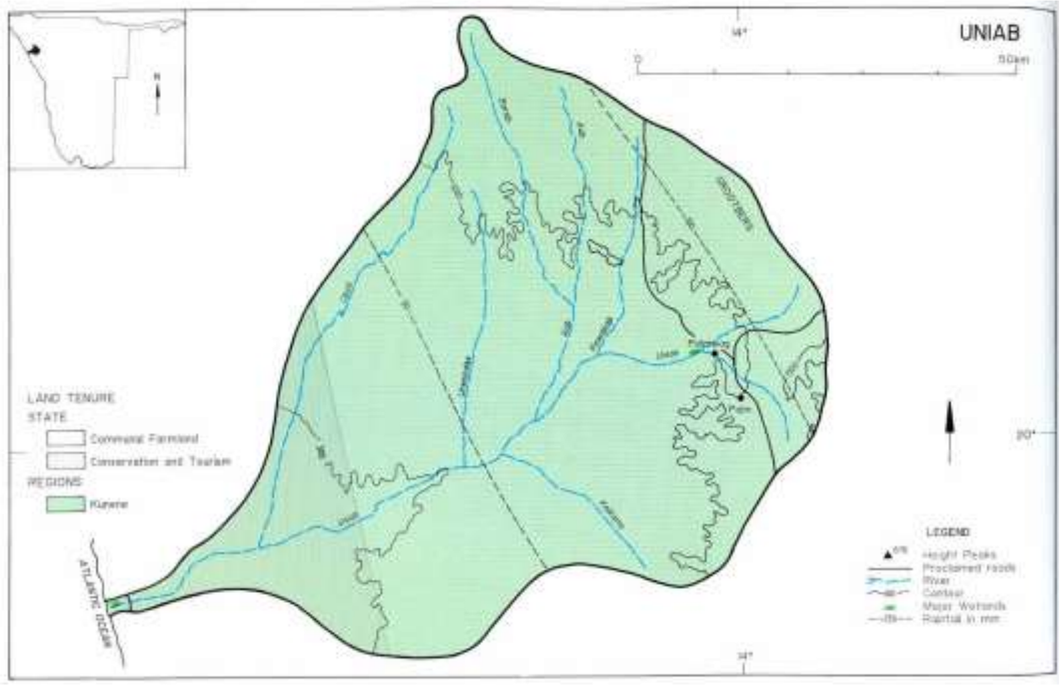
Temporary delta – Hunkab – 1995 flood (PJ/R)

The Hunkab River palaeo-delta shows high-angle cross-bedded aeolianite, fluvial gravels and low-angle cross-bedded fluvially-reworked aeolian sands. It is probably of Holocene age. (Miller, 2008, p. 25-45).

UNIAB

The Uniab and Koigab flow infrequently and for short periods only, as only about 2% of their relatively small catchments receive more than 100 mm annually. The Uniab drains the red volcanic rocks of the Grootberg area, flows through rocky desert, traverses a narrow strip of low dunes, passes through a spectacular canyon and over a waterfall (normally just a trickle) at the coast.

The catchment area is 3,961 km^2 and river length is 110 km (Jacobson, Jacobson & Seely, 1995; Strohbach, 2008).



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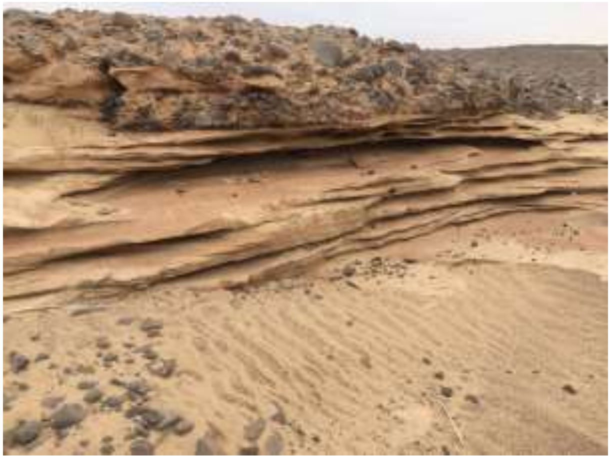
The Uniab manages to break through the ~ 7 km wide southern tip of the Skeleton Coast Dune Field, with its dunes of less than 20 m height, during strong floods, such as in 1982. There is mention in literature of flash floods carrying boulders of up to 5 m diameter (Krapf *et al*., 2005; Svendsen *et al*., 2003, Garzanti *et al*., 2014; Robertson *et al*., 2012).

The Uniab River has a well-developed, raised palaeo-delta, spreading out in 9 distinct channels (1 with active surface and 4 with active subsurface flow) from ± 6 km upstream to a terminal width of ± 12 km at a 30 m high coastal cliff (Robertson *et al*., 2012). At its highest point the delta is ± 40 m above sea level, thinning to the north and south. At the coastal cliff and along recent channels cut into the delta sediments, one can see semi-consolidated fluvially-deposited brown gravelly sand, gravels and whitish lithified fluvial sands that are interspersed with reddish deposits of aeolian origin, which may have been reworked by fluvial action. The surface layers of unconsolidated gravels are probably equivalents of the Gobabeb Formation gravels of late Pleistocene age (ca. 44-20?) (Miller, 2008, pp.25-43, 25-44, 25-45).



Semi-consolidated deposits in the lower Uniab River (MC)

Uniab River Canyon and waterfall near the mouth (CDK)

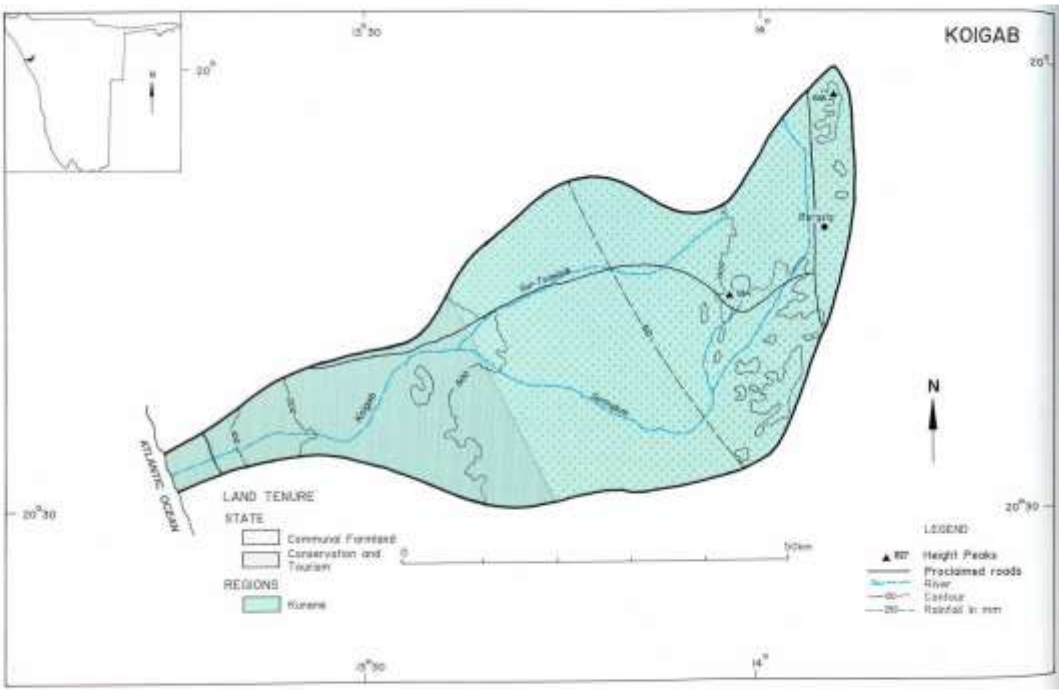
One of the five recently cut channels flows from time to time, while the other four have active underground seepage that manifests itself in numerous brackish pools and associated vegetation.


One of the vegetated channels of the Uniab close to the coast (MC)


A spring in the lower Uniab (MC)

KOIGAB

The Koigab originates in the Etendeka Plateau. It is a braided river that It flows through a 24 m deep canyon, the Koigab Poort, about 20 km from its mouth, and has several springs and seeps. It is not impeded by dunes. Isolated barchans move northwards across the fan deflation surface towards the Skeleton Coast dune field. The catchment area is 2,321 km^2 and river length 130 km (Jacobson, Jacobson & Seely, 1995; Strohbach, 2008). A detailed study of the river and particularly its alluvial fan had been done by Krapf (2003).The fan is about 15 km long from apex to toe, and 23 km across its maximum lateral extent (Krapf, 2003).



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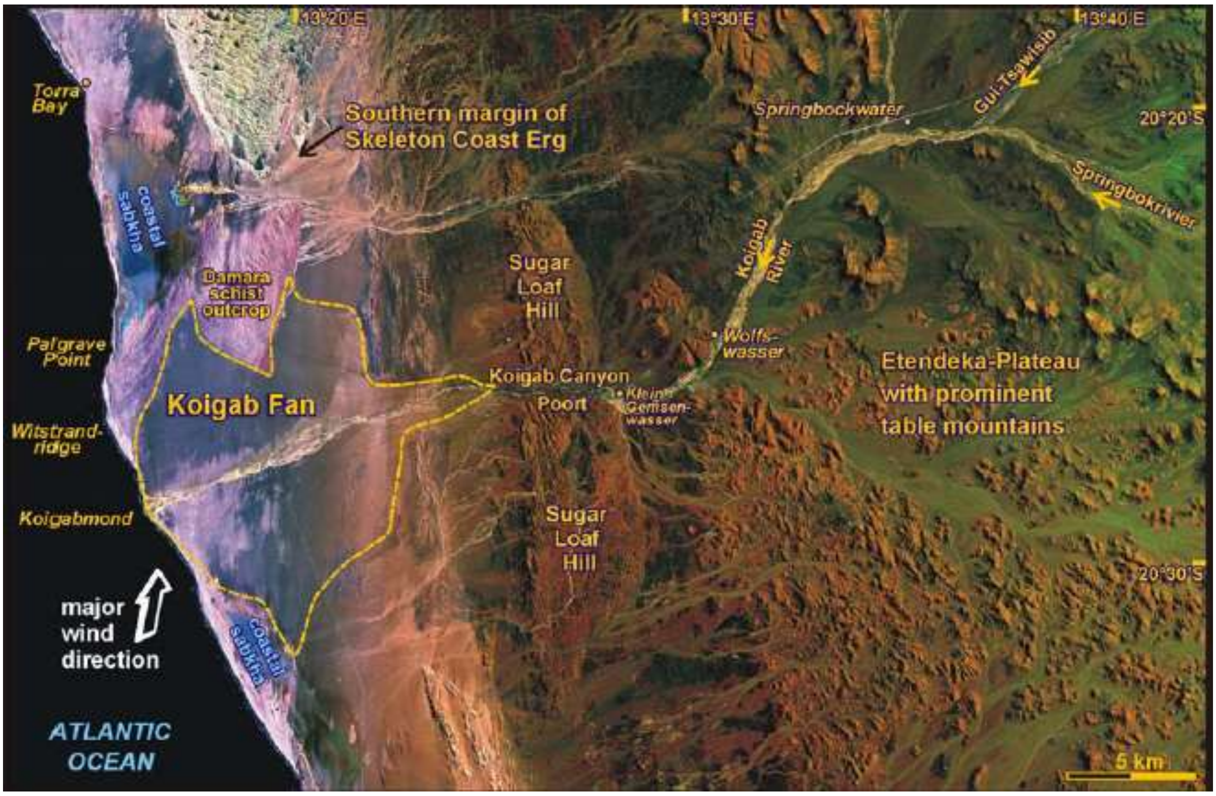


The Koigab Alluvial Fan, reproduced from Krapf (2003). Landsat TM-5 image 171-074, 7-4-1 R-G-B, 31/03/1995)

	Koigab	Uniab	Hunkab	Hoanib	Hoarusib
Catchment area (km3)	2,321	3,961	593	15,761	15,237
River length (km)	130	110	65	270	300
Elevation range (m)	0-1,571	0-1,625	0-1,220	0- 1,821	0-1,964
Precipitation range (mm/a)	0-100	0-125	0-50	0-325	0-325
River gradient over last 20 km (%)	0.98	1.18	1.68	0.93	0.55
Catchment area (%) with < 100 mm/a rain	2	2.3	-	72	40
Persistent riverbed through erg	y	y	n	n	y

Comparison of some ephemeral rivers of the SCNP.

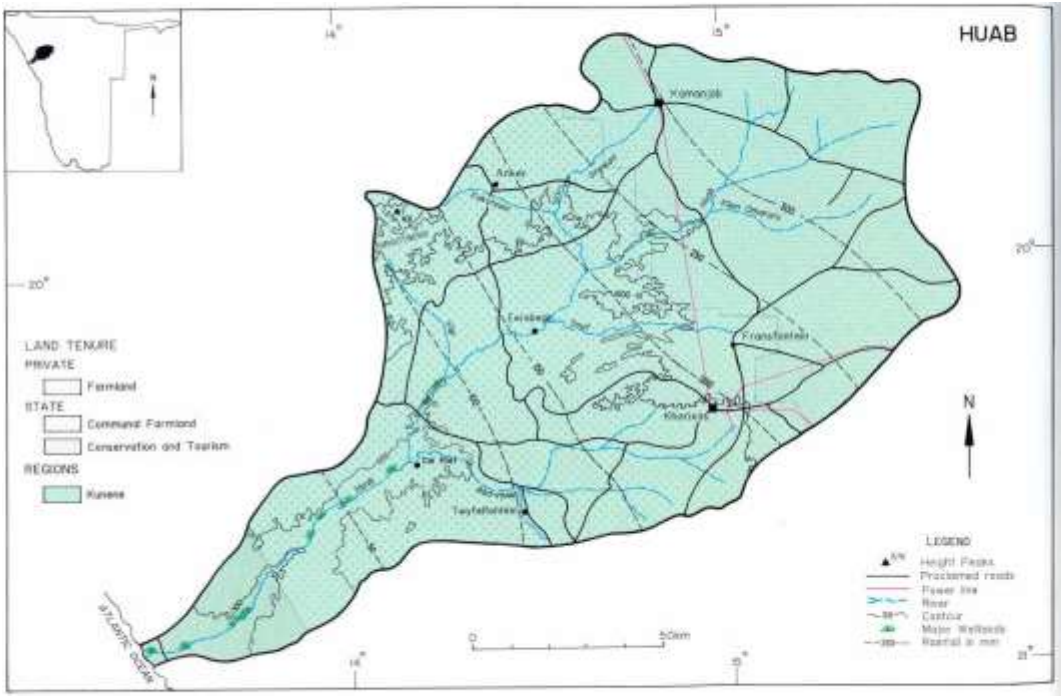
[Reproduced from Krapf (2003, p. 168), based on original data from Jacobson *et al*. (1995), Lancaster (1982) and Miller (1988), with some adjustments following Strohbach (2008)]

HUAB

The Huab River drains the Kamanjab basement and flows along the southern edge of the Etendeka plateau (Garzanti *et al*., 2014). The catchment area is 16,466 km^2 and river length 300 km (Jacobson, Jacobson & Seely, 1995; Strohbach, 2008).

Lagoon at the mouth of the Huab River (BA, MC)

The river ends in a salty 'lagoon' that is separated from the ocean by a sand bar. It supports populations of seabirds such as flamingos and cormorants, but no vegetation.



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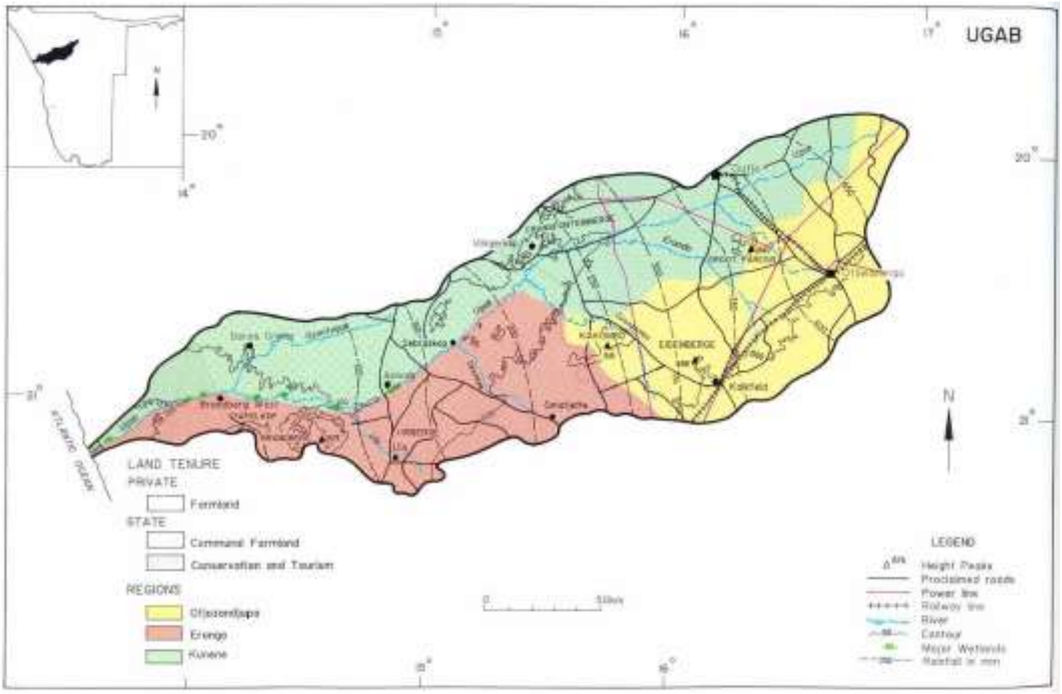
UGAB

The Ugab forms the southern border of the SCNP and skirts the northern flank of the Brandberg. It is the longest of the SCNP ephemeral river, with a catchment area of 29,355 km^2 and river length of 450 km (Jacobson, Jacobson & Seely, 1995; Strohbach, 2008).

Gravel terraces are present about 40 m above the current level of riverbed, where the river cut deeper into the former floodplain when sea levels dropped from 12 Ma.



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RIVER VALLEYS OF GLACIAL ORIGIN

Parts of the Kunene, Engo and Munutum river valleys [and the Gomatum, Ombonde, Aba-Huab and upper catchment of the Hoarusib, all outside the SCNP] are of glacial origin. These date from Permo-Carboniferous Dwyka glaciation around 300-280 Ma and occur at the base of the Karoo Supergroup. At that time, Namibia was in the interior of Gondwana, close to the South Pole. It is postulated that these glacial valleys were formed by outflow glaciers on the margins of a large icesheet, somewhat akin to the present-day Greenland Icesheet. The evidence of glacial origin is in the typical U-shapes of valleys, striations in valley walls and floors, remnants of glacial sediment deposits and presence of erratics (randomly dropped large rocks). (Schneider, 2004; Robertson *et al.*, 2012; Jacobson, Jacobson & Seely, 1995).

Location of lava flows, glacial deposits and glacial valleys (Jacobson, Jacobson & Seely, 1995)

WASHES AND GULLIES

[Arroyos]

Rare, brief desert rainfalls generate sheet flow floods that concentrate in fast-moving but short-lived streams that cut drainage channels in the gravel plains – from indistinct, broad, shallow washes to deeper gullies with steep sides and gravel-strewn flat bottoms. These dry watercourses in arid regions, which only flow briefly after rains, are known as *washes* or *arroyos*. Being in the lower, water-receiving position of a landscape and with coarse substrates that readily absorb and retain water, washes support more perennial species than the surrounding plains. This can include shrubs and even trees that provide habitats for small animals and nesting sites for birds.

Shallow wash (MC)

Wash with rocky bluff (MC)

ALLUVIAL FANS

Alluvial fans are accumulations of alluvial material deposited by streams where they exit the mountains onto plains or valley floors. At the exit point, erosion debris accumulate in a semi-circular cone that has a steep gradient and a pattern of unconfined channels that shift over the depositional body. This is caused by the change in gradient that lets the flow lose speed. The stream cannot carry the sediment load further and deposits it, quickly blocking the channel. The stream sweeps left or right around the obstacle, and over time breaks up into multiple braided channels that migrate back and forth, depositing alluvial sediments in a broad fan-shaped sediment cone. Sediments tend to be coarser at the proximal end of the fan (close to the fan head or apex) than at the distal end (the fan toe) (Driessen *et al*., 2001).

Fans formed by adjacent canyons along a mountain front can join to form a continuous fan apron, called a bajada in arid regions, or a piedmont in more humid regions.

In the mist belt, pedogenic gypsum crusts tend to stabilise alluvial fans, with a higher degree of induration corresponding to lower activity of that part of the fan (Eckardt and Spiro, 1998).

Alluvial fans (BA)

SPRINGS

Springs and seeps found in the normally dry streambeds of ephemeral rivers play a disproportionately large role as oases in the surrounding hyper-arid landscape. They occur where impermeable underlying geological formations, such as dykes, refract groundwater upwards (Watson & Lemon, 1985). The upwelling waters can be either fresh or brackish. Leggett (1997) reported that all the springs of the Hoanib up to 85 km from the coast are brackish. The following is an incomplete list of springs and seeps in the SCNP.

NAME	LAT (S)	LONG (E)	NAME	LAT (S)	LONG (E)
Auses	-19.39926	12.89117	Sarusas	-18.82462	12.49886
Die Oase	-19.45234	12.82287	Soutvlei	-19.47663	12.50975
Ganias	-19.24243	12.92238	Lower Uniab (several)	-20.18841	13.15288
Hoaswater	-19.46000	12.77826	Unjab	-20.12230	13.34860
Lower Khumib (several)	-18.86436	12.44386	Wolfswasser	-20.43232	13.50969
Munutum	-18.31594	12.08624	Klein Gemsenwasser	-20.44633	13.49376
Okau (in Munutum)	-18.30668	12.08896	Ugib Canyon	-20.44526	13.46810
Quicksand	-18.65888	12.30658	Wildwater	-20.33705	13.60090

(MAWF, Hydrogeological Map of Namibia; coordinates corrected by MC, using Bing Aerial and Google Satellite Images)

Okau Spring in Munutum River (AD)

One of several springs in the lower Uniab (MC)

PANS

INLAND, TERMINAL PANS OF RIVERS

(playas)

Many westward-flowing ephemeral rivers are blocked by dune fields, so flood waters pause and spread out in short-lived, shallow wetlands. Unless floods are energetic and voluminous enough to breach the dunes, water infiltrates and evaporates to leave dried lake beds known as *pans or playas* (Shaw & Thoma,s 1997).

The Hoanib River spreads out in an elongated, silt-dominated flood plain where it meets the dune belt. When in flood, it reaches the coast along a sand-encroached channel. (BA)

Evaporation concentrates any salts that were dissolved in the water. The order of crystallisation is $CaCO_3$ and $MgCO_3$, that precipitate as calcite, aragonite or dolomite, followed by gypsum ($CaSO_4.2H_2O$) and eventually halite (NaCl) and other highly soluble salts. When the temporary lake / pond dries up, the mud at the bottom shrinks and cracks, and precipitated salts form a crust on top of the playa floor and in cracks in the soil surface. A playa with high salt content is known as a *salt pan* or *salina*.

The Kharu-geiseb, Hunkab and Engo Rivers end in playas east of the dunes. Their floods seldom have the required volume and energy to breach the dunes. In 1995, the Hunkab broke through 15 km of dunes, overtopping and eroding each dune in turn. Once the water reached the sea and water levels dropped in interdune valleys, lobes of gravel were deposited in these embayments.

Hunkab alluvial fan and terminal playa (BA)

Kharugeiseb (left) and Samanab (right) alluvial fans and terminal playas

Engo alluvial fan and terminal playa

Inland playas can also be located at the ends of active or palaeo-washes, or where water temporarily ponds after sporadic, localised rainfall, or where groundwater is very shallow. Capillary rise of groundwater brings dissolved solids to the surface, where evaporation concentrates them into halite deposits.

Salt pan with Puffic Solonchak (MC)

Polygonal features on surface (MC)

Petrosalic Solonchak (MC)

COASTAL SALT PANS

(sabkhas, salt flats, tidal flats, salinas)

Coastal pans and tidal flats with brown surfaces, flat, level topography, and usually elongated shapes occur parallel to the shore. One origin of coastal salt pans is when sand spits grow northwards from headlands across bays and eventually form lagoons. The lagoons silt up from material washed and blown in, forming layers of fine sand, medium sand and clayey sand. Pans also form when beach ridges build up between low-lying coastal areas and the ocean. As most pan surfaces are close to sea level, seepage of seawater through the sand bars and inundation during spring tides and storms replenish the brine. Some pans show the effect of tides.

Salt pan with seawater seepage, showing tidal effects (MC)

Coastal salt pan (MC)

Salts precipitate when they are sufficiently concentrated by evaporation. Salt precipitation proceeds from the surface downward, so a hard salt crust may cover concentrated brine and sulphurous, dark mud.

The soils of salt pans are mostly Solonchaks (high soluble salt content), though some pans, or parts of them, may be dominated by high exchangeable sodium content and be classified as Solonetz. They also contain significant amounts of hydromorphic surface gypsum.

Saltpan (former lagoon at Angra Fria) with Solonchak and possibly Solonetz (DC)

Salt precipitation at the edge of a salt pan (MC)

Typical layering of sediments in salt pan (MC)


Polygonal pattern caused by salt heave (MC)

Salt pans are relatively low in biodiversity, though some provide breeding sites for Damara terns. They contribute to the SCNP's 'sense of place' and have visual appeal. Some pans (e.g. at Cape Fria) are thought to hide old shipwrecks. Vehicle tracks can persist for years on pans.

GRAVEL PLAINS

Large expanses of the coastal platform consist of broad, almost flat (i.e. low relief) and almost level (i.e. low slope) pediplains with shallow washes and drainage lines, low ridges of resistant rock and some inselbergs. Slopes, even at very low angles, channel water to the local lowest points of a landscape and, thus, have a large influence on vegetation development. Tiny relief changes, even of just a few centimetres, are significant for control of rare surface runoff (Robinson, 1976).

COASTAL GRAVEL PLAINS

A strip of approximately 10 km wide along the coast is a mosaic of salt flats, gravel plains and gravel-covered foothills with biological soil crusts dominated by lichens, and extensive surfaces of exposed schists, granites and basalts (Lalley, 2005).


Coastal gravel plains (MC)

Coastal gravel plains are fragile systems, with frequent fogs providing moisture to lichen fields and some plants (MET, 2013). Biological crusts of lichens, cyanobacteria (and to a lesser extent algae and mosses) entrap mineral grains and stabilise the surface of gravel plains, especially those in the fog belt closer to the coast. When damaged, for example by off-road driving, the bio-protection is lost, and recovery is extremely slow.

Biological crusts (MC)

Vehicle tracks destroy the puffy Gypsisol structure, push stones into the soil and flatten microtopograhy, thus reducing growth surfaces for soils, in addition to damaging the lichens themselves. By destabilising the gravel pavement, soil structural properties are disturbed and susceptibility to wind erosion enhanced (Daneel, 1992; Eckardt & White, 1997)

INLAND GRAVEL PLAINS

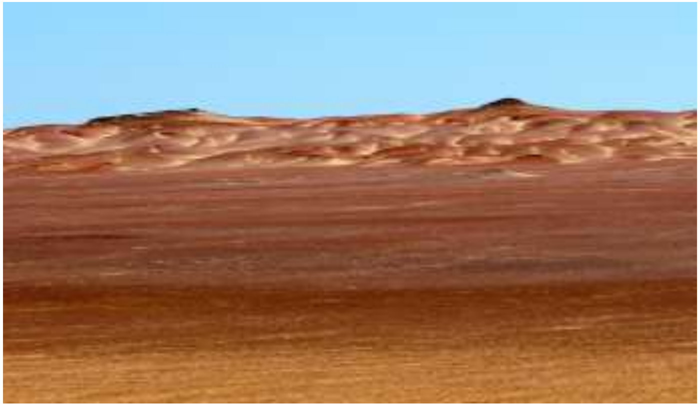
Vast, visually striking, multicoloured plains are found inland of the coastal dunes. These are only occasionally subjected to fog. After rains, they transform into annual grasslands that support plains game.

Inland gravel plains (AD)

These plains have a variety of soils, such as Gypsisols, Calcisols, Regosols and Cambisols – many with aridic or yermic (desert pavement and/or vesicular layer) properties.

Heavy mineral particles are abundant, with garnet adding a pinkish to dark reddish hue and ilmenite, magnetite and fine basaltic granules adding black hues.

Inland gravel plain with ephemeral grasses (MC)



Inland gravel plains (AD)

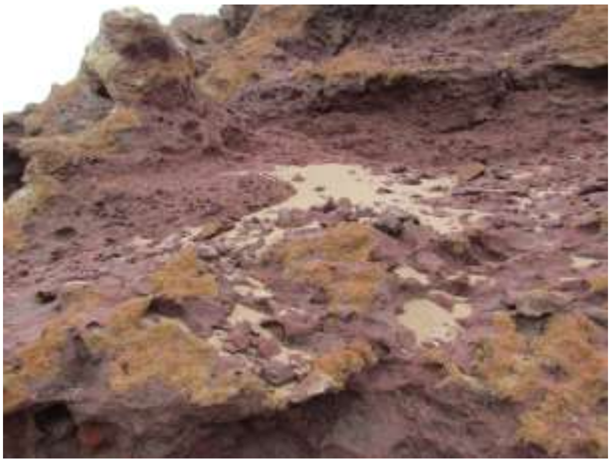
ROCKY AREAS AND INSELBERGS

Following the break-up of Gondwana, magma intruded into pre-existing rock. The most weathering-resistant intrusive rocks withstood the millions of years of subsequent erosion in the form of exposed expanses of rock (see 'Aeolian Deflation and Abrasion'), inselbergs and dykes. In places, the bedrock has a scattering of pebbles with no associated sand, which had long ago been removed by the relentless wind.

INSELBERGS

Inselbergs (isolated mountains) are important habitats within the arid environment, as they support greater numbers and more diverse biota than the surrounding plains. They intercept fog and may receive more orographic rainfall. Some SNCP inselbergs are horizontally-layered remnants – in the form of mesas and buttes – of Etendeka basalts, while intrusive granites appear as jointed koppies and exfoliated domes (Robertson *et al*., 2012; Anonymous, 1999). Differences in rock hardness allow formation of hollows and cracks where colluvial and alluvial material collect. Runoff from rock surfaces provide more mesic conditions in these sandy/gravelly pockets, providing favourable conditions for plant growth. Rock crevices, hollows and ledges provide shelter for reptiles, insects and birds. Inselberg biota is strongly influenced by size, lithology and distance from the coast (in terms of fog and temperature conditions). The foot slopes of inselbergs typically have slightly deeper and finer-grained soil with somewhat higher water-holding capacity than the surrounding plains and they support more species-rich plant communities and more biomass.

Fog-supported plants in a soil pocket (MC)



Lichens flourishing on the seaward side of Agate Mountain (MC)

Inselberg covered in lichen and dry grass (MC)

Deeper, finer-textured materials at the foot of a hill or mountain support more vegetation (MC)

Agate Mountain is a prominent, almost circular hill of 1.5 km diameter near Cape Fria. It is a carbonatite, composed of igneous rocks with > 50 % calcium, magnesium and iron carbonate minerals. The 'agate' is aragonite, a type of calcium carbonate formed in cavities as magma cools down (Swart & Marais, 2009). Its formation and characteristics were described in great detail by Miller (2000).

Agate Mountain (MC)

Aragonite found at Agate Mountain

Rocky hills at Sarusas Mine (MC)

(MC)

DYKES

(US spelling: dikes)

Dykes are sheet-like near-vertical intrusions of igneous rock that cuts across older host rocks. They form when magma is forced into fissures in the host rock. In the SCNP, dykes – usually of dolerite or marble and often covered by desert varnish – form long, low, dark ridges where weathering and erosion has removed softer host rocks. They provide different growth conditions than the surrounding plains, containing more lichens and succulents.

As in the case of inselbergs, dykes generally support more and different vegetation than the surrounds, due to different soil properties and more fog accumulating on the rocks.

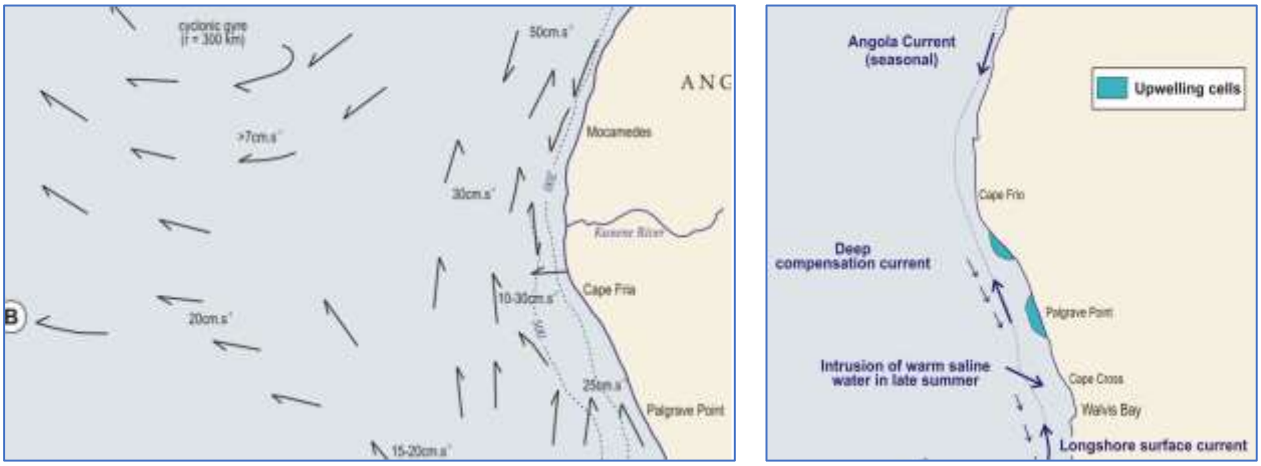
Dolerite dykes (MC)

COASTLINE

BENGUELA CURRENT

The Skeleton Coast coastline owes its N/NNW-S/SSE orientation to the break-up of Gondwana about 127-132 Ma ago. There are no deep, sheltered inlets; just some small bays north of (rocky) headlands. Several elongated coastal salt pans are found parallel to and just inland of the coastline.

Both the Skeleton Coast marine and terrestrial environments are strongly influenced by the cold Benguela Current that constitutes the eastern leg of the anti-clockwise South Atlantic Gyre. It flows north along the Namibian coast at a speed of 10-30 cm (offshore of Walvis Bay; slower further north), until it meets and subducts under the warmer Angola Current at the Angola-Benguela Front. The Benguela current is enhanced by strong southern winds generated by the South Atlantic Anticyclone and steep local air pressure gradients due to daytime heating of the land.



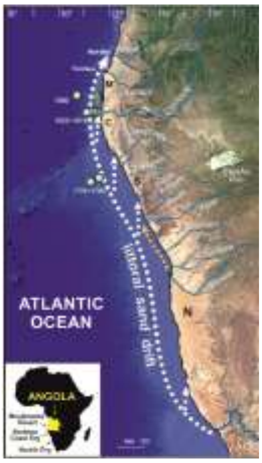
Surface currents (left) and upwelling centres (Midgley, 2012)

The Coriolis effect causes deflection of the Benguela to the northwest (Ekman current / transport) which is compensated for by upwelling of cold, deep water. Upwelling along the Namibian coast is strongest off Lüderitz (Lüderitz Cell), with smaller cells off Cape Fria (Northern Namibia Cell), Palgrave Point (near Torra Bay) and Conception Bay (Central Namibia Cell) (Robertson *et al*., 2012; O'Toole, 1997). Upwelling is both biologically significant – bringing nutrients to the photic zone and thus stimulating primary productivity – and as a source of marine sulphates for the formation of gypsic soils. A counter bottom-current flows southward near the edge of the continental shelf.

The Walvis Ridge separates the Angola Basin from the Cape basin and presents an obstruction to deep ocean circulation (Robertson *et al*., 2012). It is a submarine mountain chain of extinct volcanoes from off Cape Fria to the mid-Atlantic Ridge in the direction of Tristan da Cunha and Gough Island, with a length of > 2,500 km and height of > 4,000 m above the abyssal plain. The Namibian continental shelf is generally 100-140 km wide south of the Walvis Ridge, but narrower north of it, and only 30 km wide in the vicinity of the Kunene River.

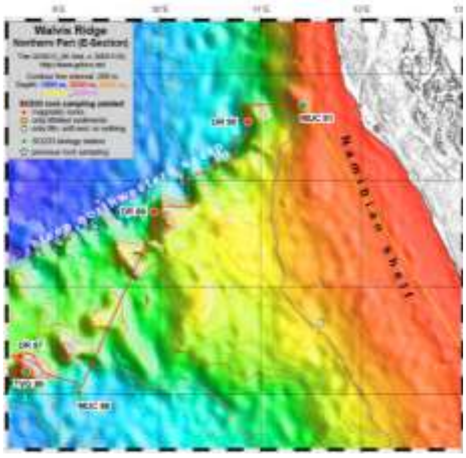
The tidal range along the SCNP coast is ± 1.4 m.

Sea levels on the Namibian coast has risen, on average, about 1.87 mm per year since 1959 (Mather, Garland & Stretch, 2009). Though coastal erosion happens in places and is likely to be exacerbated by sea level rise due to climate change, many parts of the Namibian coastline are aggrading.



Littoral sand drift

Offshore topography (Robertson *et al.*, 2012)



Bathymetry of the northern part, eastern section of the Walvis Ridge (GEOMAR, 2014)

LITTORAL SAND TRANSPORT

Sand brought from the southern African interior by the Orange River is transported more than 1,750 km north along a < 3 km wide subtidal swath of the 100-150 km wide continental shelf. It is powered by swell waves, driven by strong southerly winds, in the longest littoral sand drift cell so far recorded on Earth (Bluck *et al*., 2007; Garzanti *et al*., 2014). Changes in the orientation of the coastline allow sediment to come ashore and accumulate in four consecutive dune fields: the Namib, Skeleton Coast, Kunene and Curoca-Bahia dos Tigres (Moçâmedes) Ergs. One permanent river (the Kunene) and a series of ephemeral rivers interrupt the northward progress of each dune field, but sediment flushed back to the ocean is continually transported further north by the 'submarine sand highway' as it is called by Garzanti *et al*. (2017).

LITTORAL ZONE

The littoral zone encompasses the following subdivisions:

- The **supralittoral** (supratidal / splash / spray) zone is the area above the spring high tide line that is only occasionally inundated by storms during exceptionally high tides, but receives sea spray regularly.
- The **eulittoral** (midlittoral, mediolittoral, intertidal, foreshore) zone lies between the seldom-inundated spring high tide and seldom-drained spring low tide lines.
- The **sublittoral** zone extends from the spring low tide line to the part of the continental shelf where sunlight still reaches the seabed, ± 30 m deep (the euphotic zone).

J-BAYS AND SAND SPITS

Coastal geomorphology is greatly affected by a pair of strong wind-driven wave swell-regimes, one originating in the Atlantic Storm Belt (the 'Roaring Forties') in the Southern Ocean and the other generated by the South Atlantic Anticyclone. They produce a persistent, northward longshore current along the inner continental shelf that forms a powerful littoral transport system, stretching from Cape Town to the Gulf of Guinea (Robertson, *et al*., 2012; Garzanti *et al*., 2014). Wave fronts meet the shore obliquely. Wherever they encounter a rocky headland, waves are refracted and lose some of their energy and thus their ability to keep sediment in suspension and motion (Garzanti *et al*., 2014). Sand is deposited as northward-pointing sand spits, and in J-bays behind headlands.

J-bays (BA)

SANDY BEACHES

SCNP beach sands contain 70-80% quartz. According to Garzanti *et al*. (2014), around a third of beach sands between the Namib Sand Sea and Skeleton Coast Dune Field were supplied by the Orange River, with the remainder mostly contributed by the Swakop River that drains the Damara Supergroup of central Namibia. The arid climate and weathering-resistance of volcanic rocks from the Etendeka mean that they constitute, on average, a mere 4% of Skeleton Coast beach sand north of Torra Bay, though concentrations of 10-20 % are reached in some gravel-dominated pocket beaches at Terrace Bay and Möwe Bay respectively (Garzanti *et al*., 2014).

Sandy beaches (MC)

PEBBLE BEACHES

Well-developed pebble (cobble) beaches indicate high energy wave action. De Decker (1988) reported that cobbles (up to 10 cm ∅) and medium pebbles (up to 1 cm ∅) are effectively transported at depths of ≤ 15 m and ≤ 30 m, respectively, by the longshore current (De Decker, 1988).

Some date from the period about 5,500 years ago, when relative sea levels were about 1.5m higher than present. The Etendeka Volcanic Province contributeD substantially to pebbles and gravel on beaches north of the Uniab River Mouth (Garzanti *et al*., 2014).

Pebble beaches north of Torra Bay (MC)

RAISED BEACHES / MARINE TERRACES

Lying coastal areas preserve a record of Neogene and Quaternary fluvial and linear marine terraces (Miller, 1988, 2008).

Longshore drift has been transporting fluvial sand, gravel and larger cobbles northwards from the mouths of West Coast rivers for millions of years. This material was deposited as gravel and pebble beaches, and some were left stranded as raised beaches (marine terraces) far above the current high-water mark by marine transgressions and regressions. Along the northern Skeleton Coast, these marine terraces often consist entirely of gravels and pebbles, predominantly basalt from the eroded Etendeka plateau. Material tends to be sorted in bands – from lowest to highest elevation – of subtidal, sandy gravel, interbedded pebbly sand in the eulittoral (intertidal) zone, and gravel with very little sand in the supralittoral zone (storm beach) (Miller, 2008, pp. 25-35, 25-36).

Raised beaches are found at heights of 3-5 m and 8-10 m above sea level north of the Ugab and Huab River mouths, at 3-5 m north of the Koigab, Uniab, Hoarusib and Khumib River, and at 8-10 m between the Uniab and Hoanib Rivers. Terrace Bay got its name from the 3 m marine terrace in that vicinity. These terraces are probably of Late Pleistocene to Holocene age. (Miller, 2008, pp. 25-40, 25-45, 25- 46)

Marine terraces near the Huab River (MC)

Older, gravelly marine terraces that have been cemented by pedogenic gypsum and/or calcrete, occur at 13-23 m and at 30 m above sea level at various locations along the SCNP coast. Well-developed 13-23 m terraces are found between the Uniab and Hoanib River mouths. Terraces of 1 m thick are found at 20-30 m elevation in the Möwe Bay area and further north towards the Hoarusib. At Rocky Point, north of the Hoarusib, and north of the Khumib River delta, terraces occur at 13 m. Low concentrations of small diamonds have been found, mainly in the 8-10 m and 13-23 m terraces, transported from the Orange River by swell-generated longshore drift. The highest terraces are probably of Pliocene age (3-2.5 Ma), with lower terraces being progressively younger.

Miller (pers. comm., 2021) that the surfaces of raised beaches are covered by large pebbles and a lag of finer pebbles between them, with a substantial number of the large pebbles split into several parallel slices by salt weathering. Fog dissolves salt on the pebbles; this seeps into fine fractures, crystallizes and causes the pebbles to split along the fractures.

Since the epeirogenic uplift of the continental margin after separation from South America, sea levels have fluctuated widely between 200 m above current sea level around 70 Ma, and 400 m below current sea level around 30 Ma. Sea level was ± 120 m lower and the coastline 10-50 km further west 18,000 years ago, during the last glacial maximum, when much water was locked up in polar ice caps. Some 5,500 years ago, sea level was 1.5 m higher than now (Robertson *et al*., 2012; Sieser & Dingle, 1981).

SOIL CRUSTS

Abiotic soil crusts occur in arid environments where the rate of evaporation exceeds precipitation, and soluble salts such as carbonates, sulphates and silicates accumulate at or near the soil surface.

GYPSUM CRUSTS

Gypsum crusts are most strongly expressed in the surface gravels and sands of the first 5-10 km from the coast. Older deposits tend to be denser with compact surface and subsurface horizons. Large polygonal patterns, up to 20 m in diameter, are on occasion observed on the soil surface.

Gypsum crust (MC)

From 30 to 50 km inland, calcic soils (containing more calcium carbonate/lime) gains dominance over gypsic soils (containing more calcium sulphate/gypsum), though signs of gypsum have been observed up to 70 km from the coast. In the transition zone, calcic and/or petrocalcic horizons appear above gypsic and/or petrogypsic horizons in the soil profile. As calcium sulphate is more soluble than calcium carbonate, this points to downward leaching as the main mechanism of accumulation further inland.

The situation is reversed in the case of coastal pans with seawater seepage or depressions with a shallow water table: upward accumulation results in formation of the gypsic horizon above the calcic horizon. Horizontally bedded crusts containing 50-80 % gypsum with well-develop crystals can be found on the edges of coastal salt pans, and sometime desert roses. These gypsum rosettes are aggregates of interlocking, double-convex lenticular selenite (gypsum) crystals with sand inclusions that grow from gypsum rich soil or the rise and fall of the water table in brine-rich salt pans.

The main theories of Namib Desert pedogenic gypsum crust formation are summarised by Goudie and Viles (2015, pp.103-105), Miller (2008, pp. 25-50 – 25-51) and Eckardt and Spiro (1998). The leading theory is deposition of atmospheric sulphates of marine biogenic origin. Dimethyl sulphide is produced by phytoplankton during photosynthesis, oxidised to sulphate, carried as an aerosol by onshore winds and deposited on stable pediments and bedrock surfaces. Eckardt *et al*. (2001) propose that it proceeds through primary deposition of marine aerosols, evaporitic concentration at, and subsequent deflation of inland playas and coastal sabkhas, and eventual redeposition of gypsum on gravel plains. Fog provides the necessary moisture to carry surface deposits down into the soil. The SCNP's stronger winds (to destabilise / mobilise pediments), lower fog incidence (to provide moisture for vertical transport of surface sulphate deposits deeper into the soil) and less intense upwelling (to supply sulphates) account for the weaker expression of gypsic soils compared to the central Namib.

Gypsum crusts stabilise the soil surface against erosion, allowing the development of biological crusts. The gypsic soils and crusts of the Namib Desert are very old and they form extremely slowly. High soil temperatures and repeated cycles of partial dehydration of gypsum [$CaSO_4.2H_20$] to hemihydrate / bassanite ($CaSO_4.½H_20$) or complete desiccation to anhydrite [$CaSO_4$], and rehydration as from coastal fog, develop a characteristic puffy structure that feels spongy underfoot (Loeppert & Suarez, 1996). Any weight causes rupture of the soil crust and collapse of the puffy structure, thus off-road driving across gypsic gravel plains leaves highly visible vehicle tracks and expose the sandy substrate to wind erosion.

Impact of vehicles on gypsum crusts (MC)

The mapping of gypsic soils in the SCNP is incomplete as access is limited by the need to protect these crusts from vehicle damage. Soil maps also under-represent the presence of gypsic soils, as a consequence of the stringent requirements of the classification system. In the World Reference Base (WRB) soil classification system (IUSS, 2015) the shallow and stony Leptosols key out ahead of Gypsisols. Thus, large areas of the SCNP with shallow bedrock are classified as Leptosols though they clearly contain significant amounts of gypsum and have gypsum crusts. Gypsic soils are widespread and well developed at least as far north as Terrace Bay and have been observed up to Agate Mountain near Cape Fria. Most soils of the park, except sand dunes and areas of bare rock dunes, contain some form of gypsum.

DURICRUSTS

Durisols (also known as 'dorbank' in southern Africa) with silica-cemented duricrusts (petroduric horizons) may be present in the study area, but it has not been confirmed. They are strongly weathered soils developed on old, stable land surfaces, in silicate-rich alluvial and colluvial deposits. They develop over long periods when the soil reaction is so alkaline (pH > 8) that silicon becomes mobile, is translocated downward and precipitates as amorphous (opaline) or microcrystalline forms of silica in weakly cemented to indurated nodules or concretions known as durinodes. On level to gently sloping erosion surfaces of arid regions, Durisols are often eroded down to their resistant petroduric horizons, forming stable landscapes. They can grade laterally into or occur in association with gypsic, petrogypsic, calcic and petrocalcic horizons.

Duricrusts from the Central (left) and Southern (right) Namib (MC)

VESICULAR CRUSTS

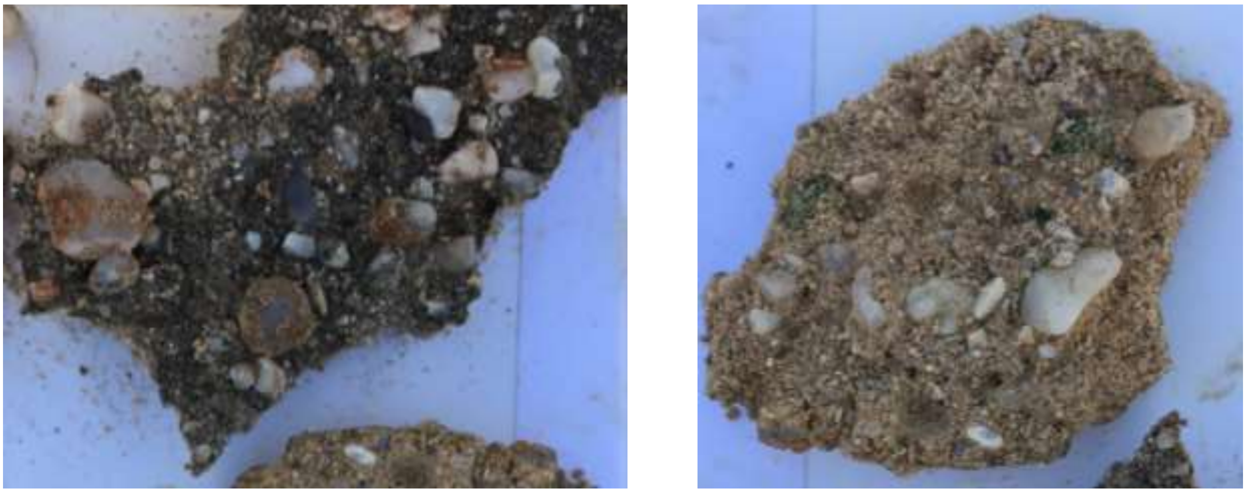
Gravel embedded in a thin, weak vesicular crust with 'bubble' or 'foam-like' appearance is frequently found in the upper few centimetres of unvegetated, structurally unstable silty loam soils with a desert pavement in arid areas (Ellis, 1988; Ellis, 1990; Lambrechts & MacVicar, 2004). The loamy vesicular layer often displays a polygonal network of desiccation cracks, extending into the underlying layers, filled with finely textured in-blown material. Low amounts of rainfall or fog moistens does not penetrate far into the soil and the fine mud dries out quickly. It traps air bubbles from soil macropores that did not have time to work their way to the surface. There are some theories that cyanobacteria may be involved in producing these bubbles. Desert varnish, ventifacts, soluble mineral accumulations, layers with weak to moderate platy structure, thin aeolian sand or loess may occur on the surface above the vesicular layer.

Vesicular crust (MC)

BIOLOGICAL CRUSTS

Biological crusts help to stabilise gravel plains, pediments and alluvial fan deposits in hyper-arid to semi-arid areas. They also occur on rock fragments and gravel. These crusts are widespread in the SCNP, though not as well-developed as between the Kuiseb and Ugab Rivers further south. They are mainly associated with Gypsisols and Calcisols, with the soil particles aggregated in a fragile surface crust by a combination of lichens, bryophytes (mosses and liverworts), green algae, microfungi and cyanobacteria. Biological crusts are ecologically important, inter alia for their drought tolerance and ability to reduce or eliminate wind and water erosion, trap moisture (fog being an important source of moisture in the Namib Desert), fix atmospheric nitrogen, cycle other nutrients, and provide shelter for vascular plants to germinate and grow. Their fragility and slow recovery rates after mechanical disturbances render them highly susceptible to long-term damage by off-road driving and mining. (Belnap, 2001; Lalley 2005; Warren, 2014)

Biological crusts (MC)



Top and bottom of a biological crust (MC)

WEATHERING, EROSION, DEPOSITION

WEATHERING

Rainfall, seasonal and diurnal temperature fluctuations and wind action all influence the rate of weathering. In arid regions like the SCNP, chemical weathering is inhibited by the shortage of moisture, so that physical weathering is the primary process.

The high solar radiation received during the day – enhanced by the Namib Desert's position in the tropics and virtual absence of clouds – heats the ground to very high temperatures. The ground colour determines its response to radiation: the albedo of desert soils is typical 24 (darker soil, absorbs more heat) to 28 (lighter soil). At night, the dry air allows the desert to cool down dramatically through conduction and longwave radiation. Cold, dense air collects in the lowest part of the landscape. The large diurnal temperature variation induces shear stress in rocks and minerals, thereby hastening physical weathering.

Exfoliation (or 'onion-skin weathering') – flaking-off of outer layers – is caused by different rates of heating and cooling through the bulk of the rock. This is particularly noticeable in granites. In some places, thick sheets (≥ 20 cm) may spall off the granitic core of inselbergs. The process is hastened if water enters cracks, freezes and expands on very cold nights.

Airborne salt (aerosols in sea breezes, fog) seeps into pores and cracks of rock. Evaporation lets salts crystallise and expand, which exert pressure on the walls of the rock pores. If the pressure exceeds the tensile strength of the rock, granular physically (mechanical) weathering occurs – a process known as 'haloclasty'. The salt crystals physically pry mineral grains apart. Wind also plays a role in the process, by promoting evaporative salt crystal growth. Once cavities form, a reduction in air pressure within the cavities results in increased wind speed and enhanced evaporation – a positive feedback. Goudie and Viles (2014) suggest that salt weathering may be implicated in the process of 'haloplanation' - creating the large, relatively flat coastal plains of the Namib.

Salt weathering resulting in honeycombed rock (MC)

Initiation of salt weathering (MC)

Tafoni (photo from central Namib) (MC)

Small hollows are called 'alveoli', closely clustered alveoli are referred to as 'honeycombs', while larger cavities with rounded entrances and smooth concave walls are known as 'tafoni'. These weathering forms are most frequently found in granular rock such as sandstone, granite and sandy limestone. It may be that the process is initiated by weathering out of large phenocrysts (large crystals) from the matrix. The salts mostly implicated in salt weathering are sodium chloride (NaCl) (halite), sodium sulphates such as mirabilite ($Na_2SO_4.10H_2O$), gypsum ($CaSO_4.2H_2O$) and other calcium sulphates, epsomite ($MgSO_4.7H_2O$), hexahydrite ($Na_2SO_4.6H_2O$), and sylvite (KCl).

EROSION

Mass wasting (downslope movement of material under influence of gravity), fluvial action (mainly unconfined sheet floods) and wind are the main erosive agents in the desert. Though wind is most prevalent, occasional rainfalls have a disproportionately large effect on redistribution of material due to the greater erosive power of water. Short, intense rainstorms loosen unprotected soil particles by rain splash, while the lack of vegetation and soil organic matter, and the presence of soil crusts and desert pavement make for fast sheet runoff, even on very gentle slopes. Seeking out the lowest positions in the landscape, sheet flow is concentrated into broad, shallow washes and, if sufficient water is present, into narrower streambeds (arroyos, wadis). The water can drop its sediment load in an alluvial fan on flat terrain, or end as a temporary playa lake in a depression where it will infiltrate and / or evaporate.

AEOLIAN TRANSPORT CORRIDORS (ATC)

Aeolian transport corridors are found where the coastal orientation changes from SE-NW to SSE-NNW, often at rocky headlands behind which beach sand accumulates (Lancaster, 1982). Very strong southerly winds mobilise sand to form trains of fast-moving barchan dunes that eventually coalesce into transverse dunes in continuous dune fields. (Garzanti *et al.*, 2014; 2017)

Sand grains up to 2 mm are moved by strong southerly (SSE) winds in a process called saltation, during which they can dislodge and move larger gravel – up to 6 times their own size and 200 times their own weight (Bagnold, in Miller, 2008c). Heavy minerals such as garnet, magnetite and ilmenite are moved more by creep than saltation. They concentrate before and behind any irregularity in the soil surface and on sand ripples.

Garnet (left) and basalt (right) grains concentrated on sand ripples and against obstacles (AD; MC)

AEOLIAN DEFLATION AND ABRASION

The process of deflation scoops out hollows in relatively unconsolidated material and can scour valleys down to bedrock. These deflation hollows vary in size from a few meters to basins of many kilometres long and tens of meters deep, some with have dune fields on the lee where the transported material is deposited.

Valleys scoured down to bedrock by wind (AD left, MC right)

Sand-laden wind abrades rock, creating several typical desert landforms, such as rock pedestals and yardangs. Rock pedestals ('mushroom rocks') have narrower 'necks' due to greater abrasion near ground level, as sand particles saltate rather than being carried aloft as is the case with smaller, lighter silt and clay particles.

The stoss side of rocks are often grooved or fluted, where softer material, such as carbonates, has been removed faster than more resistant minerals. These can form cones, spikes and upward-facing blades.

Grooves cut into rock by wind action (MC)

Grooves cut into stones by wind action (MC)

(MEGA)YARDANGS

The area south of the Kunene Dune Field consists of basement rock of the Swakop Group (570-900 Ma) that had been sculpted by corrosive (abrasive) wind erosion into narrow more-or-less parallel linear ridges, typically 8-10 km long and 300-350 m apart, known as yardangs. These are oriented SSE-NNW, in line with the dominant wind direction. Wind records from Möwe Bay indicate that in 62.5% of the time, winds blow from 157.5 ^{o}S-212.5 ^{o}S. Pre-existing jointing of the basement rock provides a structural framework which is enhanced by the sand-blasting effect of the wind. (Goudie & Viles, 2015; Goudie, 2007).

Yardangs S of Kunene Dune Field, N of Engo Valley

Yardangs S of Engo Valley, NE of Angra Fria

VENTIFACTS

Ventifacts (wind-faceted stone; 'windkanter') are individual stones, usually from hard, fine-grained rocks, that had been pitted, grooved, etched or polished by wind-driven sand. 'Dreikanter' are pyramidal ventifact with three facets on the surface, while the buried side is rounded or irregular.

Dreikanter with polished upper sides (left) and rough bottoms (left) (MC)

Grooved ventifacts (MC)

DESERT PAVEMENT

[also known as stone pavement; reg (western Sahara), serir (eastern Sahara); gibber plain (Australia); syrosems (some old literature)]

A desert pavement is a layer of densely packed, almost continuous surface fragments, usually one or two stones thick – known as the lag – set in or on a matrix of finer material.

The most common mechanism for its formation in arid environments is the removal of finer particles by wind, which leaves coarse sand, gravel and larger stones behind in a layer known as lag, residue, armour or pavement. This mechanism does not cause lateral movement of the larger particles. A second mechanism is dislodgement of finer particles by raindrop splash and horizontal removal by sheet wash. A third group of processes causes vertical movement of particles, with finer material migrating downward and coarser material upwards. This can be caused by (i) cycles of wetting (by fog or rainfall) and drying, (ii) cycles of freezing and thawing (less active in the Namib Desert), (iii) salt heave or (iv) lifting by air bubbles released by respiration of soil microbes, in clayey in-blown dust that had filtered down and underneath stones. (Goudie & Viles, 2015). Gravel plains covered in desert pavement often display yermic properties (see Soils section).

Most of the Skeleton Coast National Park pediments and gravel plains can be assigned the *Yermic* qualifier, which refers to low organic matter content, light soil colours, the presence of a desert pavement, evidence of aeolian activity such as ventifacts (wind-shaped gravel, stones, rocks), a loamy, vesicular layer, and/or the presence of needle-shaped clay minerals.




Gravel plains with desert pavement (DC; MC; AM; MC)

The scuff mark shows how the gravel / stones are concentrated in a thin surface layer (MC)

Gravel plain with desert pavement

DESERT VARNISH (rock varnish, patina)

Desert varnish is a thin, dark veneer on physically stable surfaces of rocks and stones. They consist of clay minerals cemented by oxides and hydroxides of manganese (birnessite) and iron (goethite and haematite), with some other trace elements and organic matter – all of which probably come from atmospheric deposits.

Manganese-oxidising bacteria play a crucial role in their formation. The layers are typically less than 100 nm thick (Thomas & Goudie, 2000). The surfaces are shiny when smooth and rich in manganese. Their colours are black when rich in manganese, red or orange when rich in iron and brown when manganese and iron concentrations are similar.

Desert varnish (MC)

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AD – Alex Derr

AM/RB – Alma Moller/Rolf Becker

BA – Bing Aerial satellite imagery

BA/MC – Bing Aerial satellite imagery with information added by Marina Coetzee

DC – Don Cowen

GB – Gaby Hansen

GSN – Geological Survey of Namibia, Ministry of Mines and Energy

CDK – Christine & Dudley Kelbe www.kelbephotography.co.za/blogs/namibia_the_coast/#

JJS – Peter Jacobson, Kathryn Jacobson, Mary Seely (Jacobson *et al*., 1995)

JP – John Paterson

MC – Marina Coetzee

PJ/R – P Jacobson, in Robertson *et al*., 2012

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