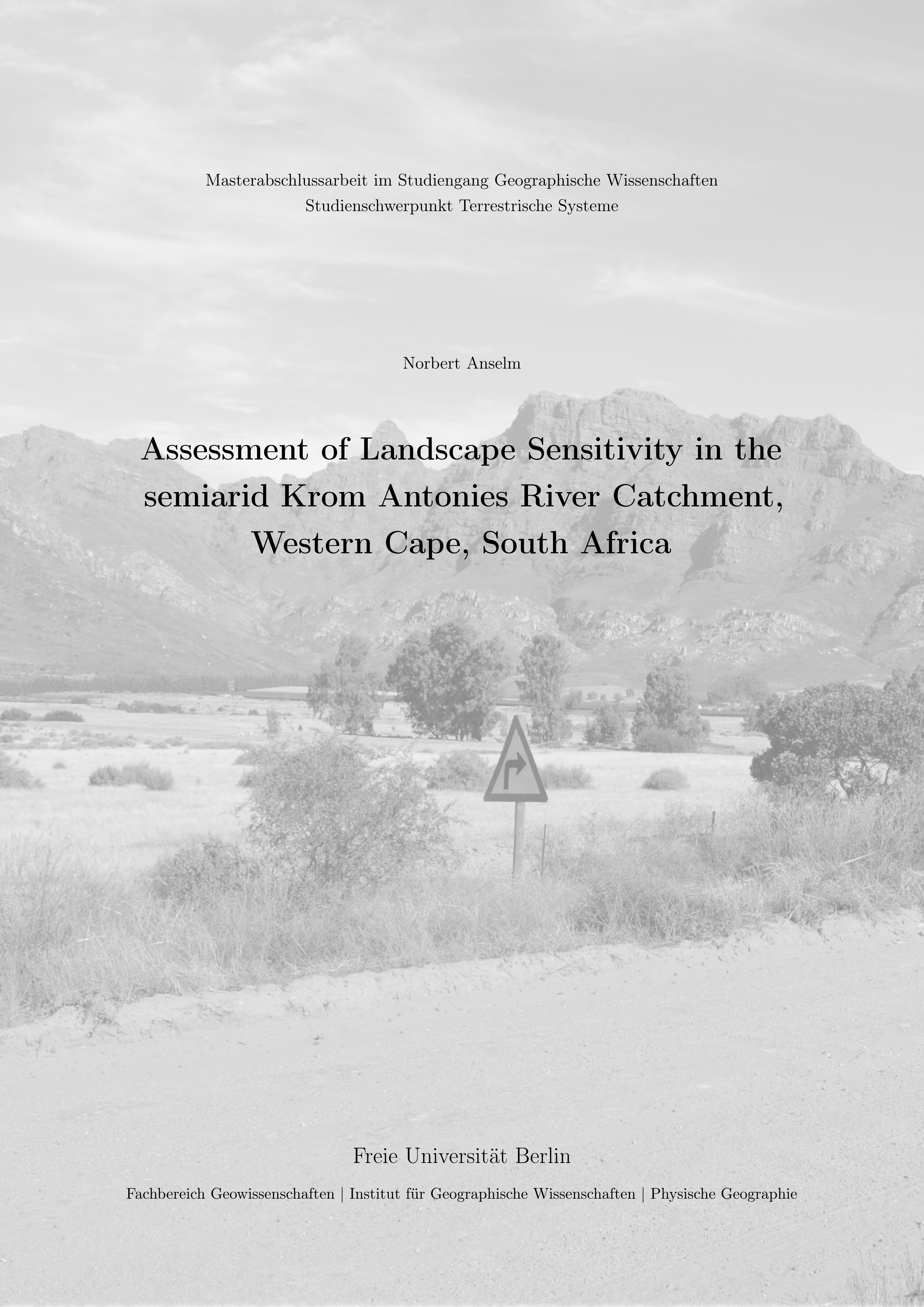

Masterabschlussarbeit im Studiengang Geographische Wissenschaften
Studienschwerpunkt Terrestrische Systeme

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Assessment of Landscape Sensitivity in the semiarid Krom Antonies River Catchment, Western Cape, South Africa

Freie Universität Berlin

Fachbereich Geowissenschaften | Institut für Geographische Wissenschaften | Physische Geographie

Assessment of Landscape Sensitivity in the semiarid Krom Antonies River Catchment, Western Cape, South Africa

M. Sc. Thesis in Physical Geography

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27th February 2015

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This document was proudly compiled using LaTeX, KOMA-Script and R.

Acknowledgement

I would like to express my gratitude to Prof. Dr. Brigitta Schütt and Prof. Dr. Karl Tilman Rost – and especially to Tilman Rost for giving me the chance, space and time to develop my own interests and projects.
I also would like to express my gratitude to the kind people in the Krom Antonies valley – especially Jacobus Smit and Herman Coetzee for their valuable insights in the past and present of South Africa and the valley in particular – and Prof. Jenny Day for her support and hospitality and for the pizza.
The German Academic Exchange Service gave the scholarship in the programme *Integrated Watershed Management Research & Development Capacity Building* better known as *Welcome to Africa*, thus cheers to the german tax payers.
I would like to thank all employees of the working groups Schütt and Böse for interesting discussions, important hints and sometimes just nonsense, especially that is

- Dr. Jonas Berking for the intro,
- Nicole Lamm for paying the ticket,
- Anette Stumptner and Stefan Thiemann for *Welcome to Africa*,
- Dr. Philipp Hoelzmann for the straightforward help and advice,
- M. Scholz^2 for the coze while the laser diffractometer did its job,
- Vivian Fehn for helping me out with the grain-size samples,
- Dr. Daniel Knitter for Emacs,
- Dr. Brian Beckers for SPEI and
- Fabian Becker for the office and the olympic swimming pools full of coffee we shared.

Finally and most important Nestor and Monique – I know you are completely fed up with South Africa and R and GIS and [put in what you like]. Thank you very much for the time, the confidence and the patience — I owe you both a lot.

Norbert Anselm
Berlin, February 2015

Abstract

In the DAAD funded project *Integrated Watershed Management Research & Development Capacity Building* the Landscape Sensitivity of the Krom Antonies River Catchment (Western Cape, South Africa) was assessed. The focus was given to climate variability, landuse and landcover and soil erosion and deposition, since this Mediterranean landscape is used intensively for agricultural production and shortage of water and/or degradation of soil were considered to affect the catchment grossly.
Climatic variability was investigated using the Standardized Precipitation-Evapotranspiration Index (SPEI) to detect droughts on various time-scales from 1948 to 2011. A magnitude-frequency-analysis (MFA) of daily rainfall data from 2001 to 2014 was conducted and aridity indices were calculated. Landuse and landcover was mapped during fieldwork and from aerial photographs. Soil erosion and deposition was modelled using Unit Stream Power Erosion Deposition (USPED).
The study showed that the rainfalls are strongly seasonal, droughts are frequent and seem recurring on several cycles. Rainfalls are spatially and temporal highly variable within the valley. Comparison to other data implied an own rainfall regime. Landuse and landcover is linked to availability of water and thus hydrology and geomorphology. Waterintensive agriculture became manifest in the upper course and along the river, the lower course is characterized by extensive pasture.
The erosion and deposition modelling showed that especially in the headwater areas large areas are characterized by severe erosion. Prominent areas for deposition appear to the footslopes. The valley bottom held no excessive values for both processes and is interpreted as morphologically stable. No landuse and landcover class could distinctively associated with either erosion or deposition exclusively.
Several (sub-)factors were estimated from sediment samples using laser diffractometry to obtain grain-size fractions. Its spatial variation agreed with field observations. The derived hydraulic conductivity showed reasonable results. The attempt to derive soil organic carbon from photographs via Partial Least Square Regression (PLSR) failed.

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List of Abbreviations

AET	Actual Evapotranspiration
ARC	Agricultural Research Council
CFR	Cape Floristic Region
C	Cover Factor
DEM	Digital Elevation Model
DWA	Department of Water Affairs
ENSO	El Niño Southern Oscillation
IWM	Integrated Watershed Management
IWRM	Integrated Water Resources Management
IQR	Inter Quartile Range
K	Soil Erodibility Factor
LS	Slope Length and Slope Steepness Factor
LD	Laser Diffractometry
m	Rill-Interrill Factor
MFA	Magnitude-Frequency Analysis
MFI	Modified Fournier Index
NCEP	National Center for Environmental Prediction
NDMC	National Drought Mitigation Center
NDVI	Normailzed Differential Vegetation Index
PET	Potential Evapotranspiration
PLSR	Partial Least Square Regression
PTF	Pedotransfer Function
R	Rainfall-Runoff Erosivity Factor
RI	Recurrence Interval
RST	Regularized Spline with Tension
RU	Response Unit
(R)USLE	(Revised) Universal Soil Loss Equation
SFD	Single Flow Direction Algorithm
SLEMSA	Soil Loss Estimation Model for Southern Africa
SPEI	Standardized Precipitation-Evapotranspiration Index
UNEP	United Nations Environmental Programme
USDA	United States Department of Agriculture
USGS	United States Geological Survey
USPED	Unit Stream Power Erosion Deposition

1. Introduction

South Africa is conflicting, while on the one side it faces a long and in parts dire history, it is on the other side a young, proud and economically ascending nation that strongly differs from the entire African continent. For example it is the only African nation that is internationally considered to be of economical weight, what found expression in joining BRIC in 2011 (Smith 2011). But as mining and natural resources (Yager 2010), and manufacturing (OECD 2013) are large contributors to the GDP, the agricultural industry is an important but rather small contributor (World Bank 2011).

Not so in the Western Cape province. While it covers only 12 % of the total agricultural land, up to 60 % of South Africa's exports come from that province and Western Cape still contributes 20 % to the total agricultural production. The Western Cape and the Swartland region in particular are also known as the breadbasket of South Africa (Noemdoe et al. 2006; Bugan et al. 2012) and wide areas are dedicated to wine, fruits and potatoes (Elsenburg 2015).

But South Africa and the Western Cape province are facing some serious environmental problems. Agriculture and its irrigation in particular accounts for more than half of the whole nation's water requirements (Frenken 2005). Additionally in such latitudes climate is often triggered by anomalies like droughts, flooding, even tornadoes are possible (Fauchereau et al. 2003). Such climate anomalies are a threat to people, to the region's food and water security (Hulme 1996) and to the economic development.

Another point is that South African soils are generally predisposed to soil erosion due to the erodible soils and a lack of good farming practice (Le Roux et al. 2007). Soil erosion is not a new phenomena. Already in colonial times the concerns about the degradation of the land arose (Beinart 1984) and obviously the conservation efforts in the Swartland were not sufficient (Meadows 2003). In a region that is almost exclusively dependent from agricultural production the loss of invaluable topsoil is a severe ecological and economical menace.

Since soil erosion is an issue it is unlikely that a single factor is able to explain a regional denudation pattern (Le Roux 1990). Amongst others the landuse and landcover has been found one or even the most important factor that determines soil erosion in Mediterranean environments (García-Ruiz 2010). The effect of nonnatural landcover (like agriculture) on runoff and erosion rates – especially in hilly areas – is known (Kosmas et al. 1997). Yet some more than 78 % are still considered to be natural in Western Cape (67 % in the West Coast Municipal) but in the recent years significant transformation could be observed so that the share of cultivated land is increasing (Gillian & Govender 2013).

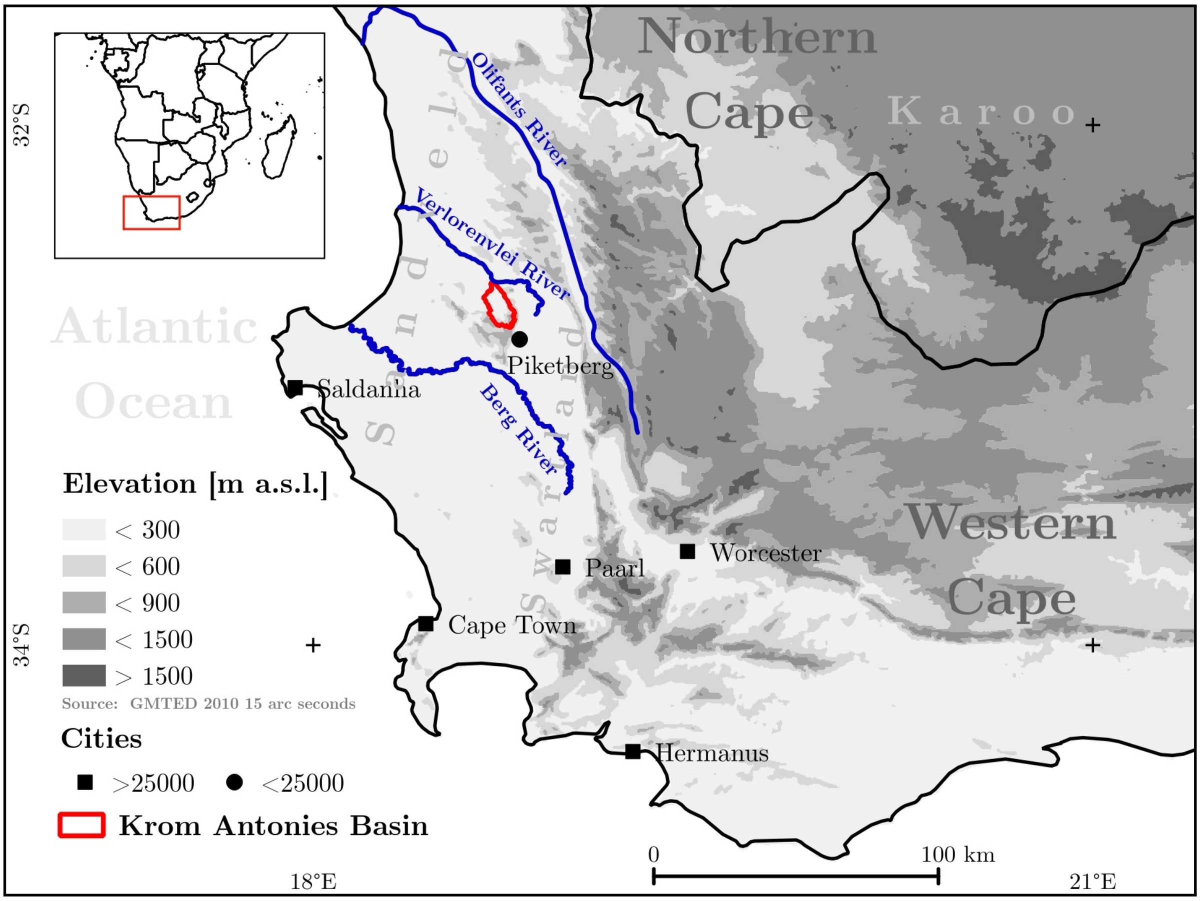


Figure 1.1: Overview map showing the wider area of the Krom Antonies river catchment in the Western Cape province.

Against the background of this issues the concept of Landscape Sensitivity (Brunsden & Thornes 1979) is applied in the small catchment[1] (approximately 119 km^2 – Figure 1.2) of the Krom Antonies river in the West Coast Municipal (Figure 1.1). Landscape Sensitivity – in brief – is the likelihood of a landscape system to alter due to perturbations in the controlling environmental processes (Thomas 2001). In fact Usher (2001) proposes to think about landscape holistically and to consider all relevant factors and triggers, but in the context of this study the focus is given by the environment of the Krom Antonies river catchment itself: climatic variability, landuse and landcover and soil erosion. Hence this selection of controlling processes and factors is used to work on the overall aim to estimate the vulnerability of the Krom Antonies river catchment. So from there the three main research questions are:

[1] The catchment boundary as shown on various maps in that thesis might deviate from administrative boundary of the local Water User Association. It was derived from an Aster DEM (Tachikawa et al. 2011) using the upslope area of the location of the obvious outlet of the Krom Antonies into the Kruisman River.

1. How does rainfall variability affect the valley of the Krom Antonies?
    a) How prone is the catchment to droughts?
    b) How reliable are the rainfalls?
    c) What temporal and spatial patterns occur?
2. What is the spatial pattern of landuse and landcover in the catchment?
3. What is the soil erosion and deposition budget?
    a) Can an USLE-based erosion model be applied to such a landscape?
    b) What landuse and landcover classes have a propensity to either erosion or deposition?

This work is embedded in the German Academic Exchange Service (DAAD) project *Integrated Watershed Management Research & Development Capacity Building*. The thesis definitely fails in capacity building and development. It surely will also not solve the ubiquitous societal tensions of South Africa that come along with the National Water Act implementation of 1998 (Republic of South Africa 1998) or that are inherent to societies in vulnerable ecosystems (Vicente-Serrano et al. 2012b; Hoff 2013). But the research is believed to be a humble contribution to the understanding of some key environmental issues of the Krom Antonies valley. This contribution might be useful to fall back on when considering new strategies of environmental management, like freshwater and soil protection, promotion of sustainability and ecology, as well as the promotion and protection of biodiversity.

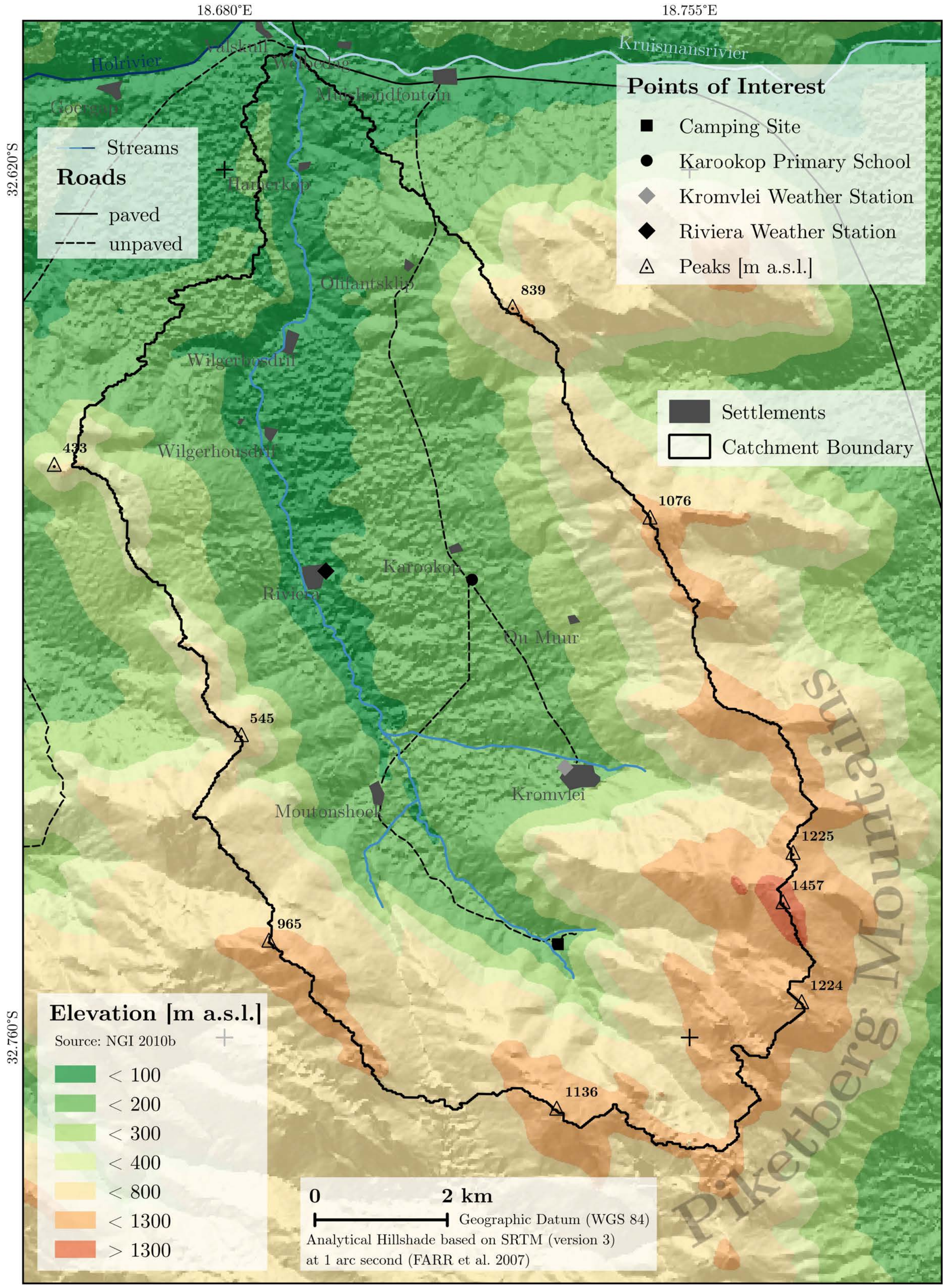


Figure 1.2: Overview map of the Krom Antonies river catchment in Western Cape.

2. Study Area

2.1. Climatic & Hydrological Characterization

The atmosphere over the southern most African continent is mainly affected by three macroscale features: by a high pressure system that develops above the subtropical South Atlantic, by cyclones moving eastward and by a low pressure system over the South African continent (Nelson & Hutchings 1983). Under the influence of the atmospheric circumpolar wave in winter the westerlies prevail in the southern Cape Province. In spring, summer and fall the semipermanent anticyclonic high pressure moves periodically to the south of the continent so that south-easterly winds gain more influence during that season (Andrews & Hutchings 1980). Additionally in austral summer a continental high pressure develops that increases the influence of the south-eastern tradewinds. The so-called berg winds occur (Nelson & Hutchings 1983).

Thus the resulting winds either press the ocean water against the continental shelf (mainly in winter) or the offshore south-easterlies force the surface water seawards (mainly in summer). Either way coastal upwelling of cool and nutrition-rich deep water occurs (Holmes et al. 2002) what in turn affects the whole south-western African oceanic and atmospheric system. The Benguela Current is the eastern boundary current of the South Atlantic subtropical gyre that flows northwestward along the westcoast of southern Africa (Stramma & Peterson 1989; Hutchings et al. 2009). The current is also influenced by the intensity of the wind vectors. It weakens in fall and winter and gains strength again in October with the maximum gradient in summer resulting in summer sea-surface temperatures as low as in winter (Bakun & Nelson 1991; Roy et al. 2001).

There is a latitudinal shift in the area of influence of the cyclones throughout the year so that the number of cyclones north of 45°S almost doubles in winter (Sinclair 1994). Nevertheless the westerlies and its associated cold fronts influence the southern subcontinent to a large degree (Kruger et al. 2010). Thus large portions of the rainfalls in the western part of the Cape Region occur in austral winter (Tyson et al. 1975) with a significant coupling to the Antarctic Oscillation and the El Niño-Southern Oscillation (Lindesay 1988; Reason & Rouault 2005). The winter rainfall area in the south-west Cape Province and a narrow maritime round-the-year rainfall zone along the southern coast are comparatively small. The majority of South Africa receives summer rainfall (Tyson 1984; Taljaard 1986).

According to the Köppen-Geiger climate classification (Köppen 1923) with criteria after Kottek et al. (2006) the region has a Csb climate which stands for warm temperate climates with dry warm summers and moist cool winters. There is a distinct seasonality and as can be seen in Figure 2.1 the time from October to March is clearly arid. Following Maliva & Missimer (2012) the latter research area is to be classified as semiarid since Figure 2.1 shows less than 500 mm of annual precipitation. Frost only occurs in mountainous regions and under the ascending effect of continental climate.

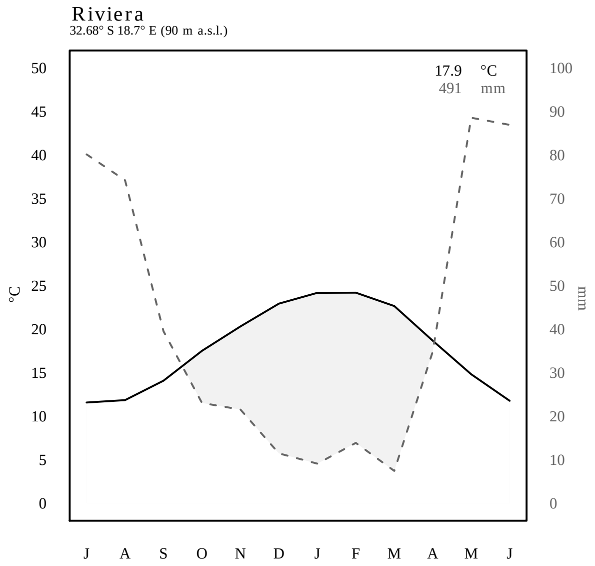


Figure 2.1: Walter-Lieth diagram of the Riviera weather station.

The precipitation has a strong gradient increasing from the west to the east with approximately 300 mm in the coastal planes and up to 1400 mm in the mountainous regions in the hinterland (Stuckenberg et al. 2013). The evapotranspiration shows a decreasing trend from north to south. The northern region of Western Cape holds values up to 2500 mm per year while the Cape Peninsula holds approximately 1250 mm of potential evapotranspiration per year. The actual evapotranspiration shows a similar trend (Wellington 1955). So the region suffers from a longterm water deficit.

Western Cape is not a part of the major continental drainage basin. The largest South African river – the Oranje river – borders South Africa on Namibia (Goudie 2005). All rivers north of the Cape Peninsula drain into the Atlantic Ocean. Some of them are largely affected by hydro-engineering activities. Most of the rivers tend to fall dry in summer and only two are reported to be perennial. Their course often does not reach the ocean during rain season. They develop vleis – groundwater supplied, arheic, brackish and retarded estuaries (Molter 1966; Sinclair et al. 1986), which are largely affected by high evapotranspiration losses during summer (Hoffman & Rohde 2011).

The area comprises unconsolidated primary and fractured-rock secondary aquifers. The groundwater abstraction is relatively high but the enormous recharge rates compensate the amounts. Normally the groundwater level mimics surface topography, but groundwater divides mostly do not agree with the surface water divides (Conrad et al. 2005).

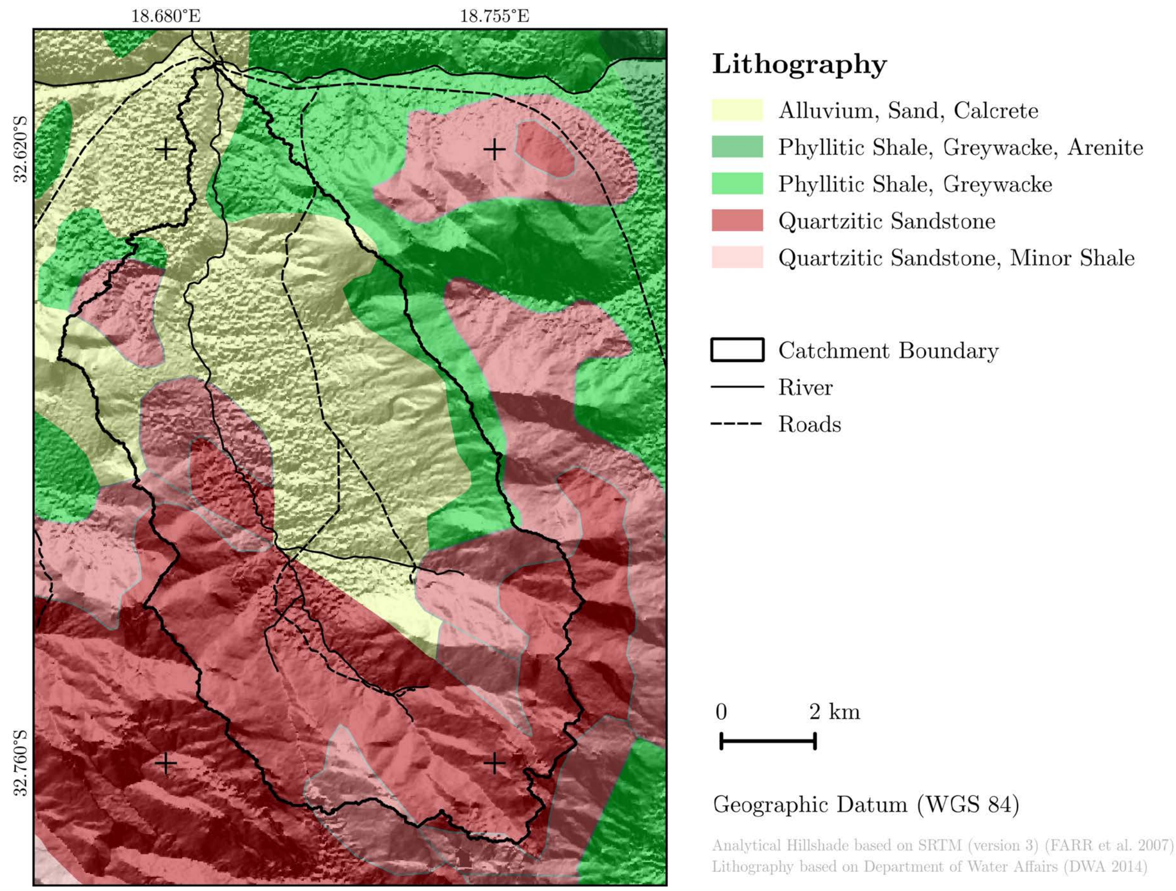


Figure 2.2: Geological Map of the Krom Antonies Catchment. Lithography was derived from vector files on a scale of 1:100 000 obtained from DWA (2014).

2.2. Geomorphological Description

The south-western Cape Province is occupied by a belt of sub-parallel mountain ranges. The southern most of them strikes from east to west in arcs concave to the south and in the west the striking goes from north-north-west with arcs concave to the west (KING 1951). The region is either dominated by superficial deposits of Tertiary to recent age or it is of solid rock from the Malmesbury or Table Mountain Group (COLE 1961). The Malmesbury Group is the oldest rock formation in the region. It consists of low-grade metamorphic sedimentary rocks of precambrian age (BELCHER & KISTERS 2003). Some 510 Ma ago the Cape Granite intruded into the Malmesbury Group (VILLAROS et al. 2006).

Since that time a stable system of steady sedimentation created a characteristic basin-and-swell structure in the center of the Gondwanian continent. The two major basins were the Cape and the Karoo (PARTRIDGE 1998; CATUNEANU et al. 2005). They were pronounced by permanent uplift and volcanism and until early Jurassic this basin developed as a tabular cratonic cover

(Tankard et al. 2009). When Gondwana started to break up the Falkland plateau drifted into the latter South African continent and caused the compression of the Cape Supergroup associated sediments. When the oceanic plate subducted under the continental plate a massive mountain range piled up consisting of the predominantly sedimentary strata of the Cape Folding Belt. That is a composition of the paleozoic Cape Supergroup, mesozoic Karoo Supergroup and some precambrian basement rocks (Lock 1978; Booth 2011).

The deformation of the Cape strata and associated sediments took place in Permian and Trias (namely between 278 Ma and 230 Ma). That orogeny created the well-known north-verging thrusts and folds in the southern Cape Fold Belt (Fagereng 2012). The newly formed mountains caused the plate to sag so that all the Karoo deposits buried the Cape Supergroup (Catuneanu 2004). After several phases of uplift since Jurassic (180 Ma) most of the several thousand meter thick layers of Karoo deposits have been eroded and the actual landscape developed (Shone & Booth 2005).

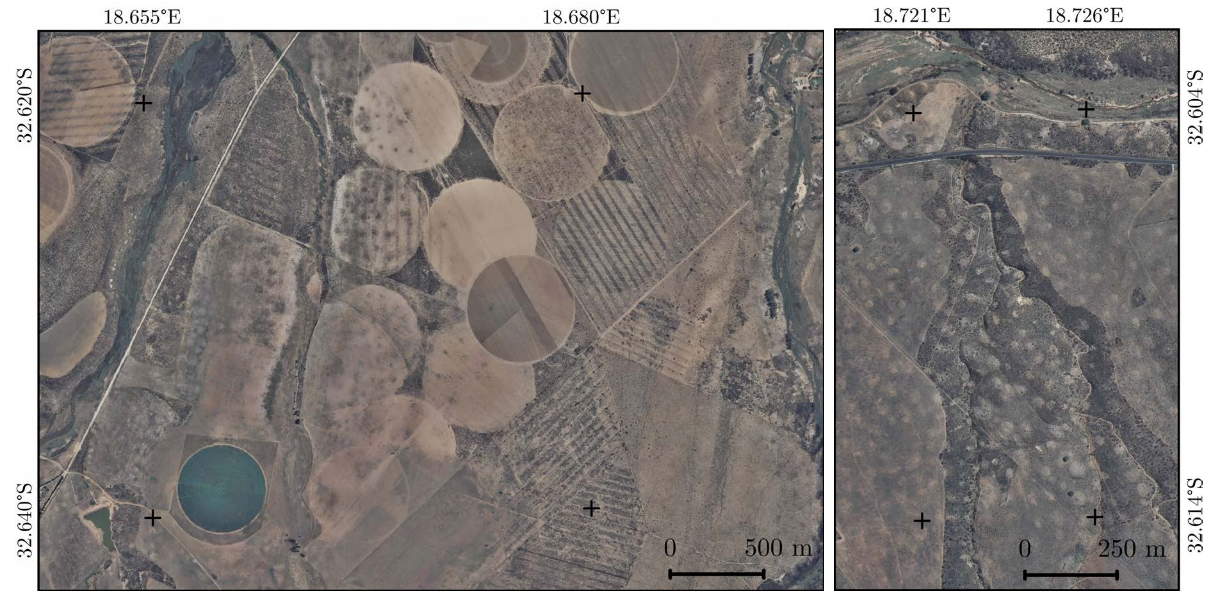
The coastal planes consists to a large scale of loose sediments. The region between the rivers Berg and Olifants and up to Piketberg and Olifants Mountains is locally termed the Sandveld – giving a hint what is the predominant soil type (King 1951). The Sandveld's soils are classified as cumulic soils after South African Soil Classification. They are immature, well drained, mostly unconsolidated material from various geomorphic processes (Fey 2010b; Bugan et al. 2012). It is a widespread soil type in South Africa – especially in Western Cape. These soil types correlate to world reference base (WRB 2006) cambisols, acrisols, fluvisols and arenosols (Fey 2010a).

The complete geological setting of the Western Cape can be seen on the geological map of the Krom Antonies catchment in Figure 2.2 on a smaller scale. While most of the mountain ranges in the south and southwest consists of quartzitic sandstone from the Table Mountain Group the morphometrically more distinctive ranges in the east and the plains beyond it are of Malmesbury origin. The center of the valley as the rest of the Sandveld in northern direction are mostly covered with quaternary alluvial sands (DWA 2014). Generally the Krom Antonies valley is well defined by the surrounding mountains.

2.3. Vegetation & Landuse in Western Cape

The Cape Floristic Region (CFR) is extremly rich in species and profoundly endemic. It even succeeds other Mediterranean biomes according to that attributes (Richardson et al. 2001; Linder 2005). The reason for the high biodiversity is multifactorial. Bond et al. (2003) got to the point asking What controls South African vegetation – climate or fire? since biodiversity is a function of rapid past aridification (Linder & Mann 1998), geomorphological forcings due to Miocene uplift (Cowling et al. 2009), high soil diversity with distinct physiographic conditions

on each mointain range (Johnson 1996) and a fire ecology with its reproductive isolation of populations in niches (Bond et al. 2003).
Most of the region is occupied by sclerophyllos bush – locally termed fynbos – what forms the fynbos biome (Cole 1961; Fairbanks et al. 2000). It borders the succulent karoo biome in the north and to the east (Linder 2003; Desmet 2007). The fynbos is often referred to as the macchia of the Mediterranean of the Southern Cape (Molter 1966). The Sandveld consists mostly of coastal fynbos in the plaines and mountain fynbos in higher altitudes. The first is a more wide-spaced narrow xerophilous type and the latter is denser and in a more hygric environment (Acocks 1988). A famous subcategory is the renosterveld. Structurally it ranges from grassland to sclerophyllous shrubs. Thus it is highly transitional and has a lower level of endemism than most fynbos. Man has a huge impact on the renostervelds. Though in Western Cape undisturbed renosterveld occupies only 6 % of its former extent (Cowling et al. 1986). Generally the impact of man (aboriginal khoisan but especially European settlers) is huge. "European man has destroyed large areas of the coastal foreland vegetation through his unrelenting quest for grazing, for fuel and for arable land. This vegetation has been depleted to the extent that at most only 14.7 % of the western coastal foreland vegetation remains in a reasonably natural state today" (Boucher 1983: p. 669).
Trees and forests are unusual in the CFR. They are restricted to southward exposed and climatically favored habitats. But most trees nowadays are invasive species (Molter 1966). Invasive species are a threat to most of the Capensis (Neke & du Plessis 2004; Milton 2004) and their impact is twofold. On the one hand they are an excessive consumer of the scarce resource of water (Le Maitre et al. 2000; Görgens & van Wilgen 2004) one the other hand it is a severe threat to the biodiversity hotspot (Heijnis et al. 1999; Baard & de Villiers 2000). Additionally the impact of climate change on the CFR must be serious since extinction and decrease in biodiversity is most likely (Rutherford et al. 1999).
From an agricultural point of view the Sandveld is extremely important for the South African economy. Potatoes and rooibos are grown there to a large extent (Archer et al. 2009). The neighboring Swartland region experienced a shift from traditionally prominent grain production towards wine grapes (Halpern & Meadows 2013). Usually rooibos is not irrigated while potatoes and wine grapes are irrigated intensively (de Wit & Crookes 2013). The Sandveld alone is responsible for one third of the South African potato production (Münch 2004).
The techniques of agriculture shifted from traditional and mostly labor-intensive to highly industrialized and market-oriented. The old way of growing grains or rooibos in strips was succeeded either by plantations (mostly grapes) or huge center pivot irrigation fields (Lotz 2012). Figure 2.3a depicts that strikingly. The stripes are remains from the former methods but their spatial patterns still persist even under pivots in use. Furthermore the figure indicates that most pivots are managed with a special cropping sequence.



(a) Pivots **(b)** Fairy Circles?

Figure 2.3: Aerial Pictures showing remains of the traditional field techniques (stripes) and the modern day non-permanent center pivot irrigated fields (NGI 2010a). The lower left corner shows a water storage dam.

Figure 2.3b shows a phenomena presumably not related to agriculture. If that aerial photograph shows the so-called fairy tales remains unclear. They are well known from more deserted areas in the north but all aerial images showed these pattern. If their origin is from ants (Picker et al. 2012), termites (Juergens 2013) or if that is a self-organized pattern of the surplus biomass-water feedback from an extended root system (Getzin et al. 2015) remains unclear as well.

Also important but to a lesser extent than agriculture is livestock. Already the earliest settlers practiced livestock farming (Mitchell 2002). The pastures are not irrigated and the grazing capacity is generally low (WCDA 2012).

3. State of the Art

3.1. Landscape Sensitivity

The concept of Landscape Sensitivity is an advancement of classical modern studies of geomorphological processes like Strahler (1952) or Chorley (1962). It is strongly influenced by the British School of Geography and Geomorphology respectively – namely by Brunsden & Thornes (1979) who introduced the concept in 1979. The concept has a long tradition in Geography but nevertheless several authors agree that landscape sensitivity is not an easy to define term (Allison & Thomas 1993; Usher 2001).

> The sensitivity of a landscape to change is expressed as the likelihood that a given change in the controls of a system will produce a sensible, recognizable and persistent response. The issue involves two aspects: the propensity for change and the capacity of the systems to absorb the change.
>
> Brunsden & Thornes 1979: p. 476

So it does not matter if it is the landscape itself as sum of all geosystems (e.g. Usher 2001) or specific aspects and subsystems of the landscape (e.g. Sauchyn 1997; Miles et al. 2001). Change occurs as an ordinary process-response function. It is associated with transportation of material, morphological evolution and structural rearrangement in case of crossing a threshold (Knox 2001). A stable system has the ability to adjust the minor perturbations within all the subsystems and external forces. In case of increasing the magnitude or the rate of stress applied to the system it tends to instability. Then a readjustment of the various subsystems sets in until a new stage of equilibrium is established (Thomas 2001). According to Phillips (1992) most earth-surface systems are rather dynamically unstable and deterministically chaotic than in a prolonged equilibrium.

The sensitivity of a landscape regarding to intrinsic and extrinsic generated changes can be expressed as a transient-form ratio. In equation 3.1 t_a is the mean relaxation time and t_f is the mean recurrence time of the changing events. If the ratio is greater than one it is likely a predominantly transient system. In case of a ratio of less than one the system tends to prevail (Brunsden & Thornes 1979).

$$\mathrm{TF_r} = \frac{t_a}{t_f} \tag{3.1}$$

In a geomorphic system the changing forces are considered to be the tectonic, climatic, biotic and anthropogenic controls of the corresponding geological, hydrological and morphological

framework of the system (Brunsden 2001). The responses of a system are uncommonly linear because the systems are complex and they are determined by a mix of external drivers, negative and positive feedback loops and various forcings (Tucker 2004; Williams et al. 2011). Fryirs et al. (2007) go into detail about the (dis-)connectivity of landscape compartments.
Change from a geomorphological perspective is mandatory scale contingent. Fluctuations can be seen either as a sequence of discrete events or as continuous, just depending on the resolution (Phillips 2009). Schumm & Lichty (1965) stated that cause and effect in landform processes is highly related to the temporal and spatial scale. Yet alterations in scale can obscure or even reverse the cause-effect relationship what makes it hard to concecptualize and operationalize in real-world studies (Phillips 1999; Viles 2001).

3.2. Water Management in Drylands

Drylands

Drylands are according to Nicholson (2011) arid and semiarid environments where evapotranspiration potentially balances or exceeds precipitation. Another approach comprises low and irregular to erratic rainfall between 300 – 600 mm per annum, infertile deplenished soils and a certain degree of overuse by man (MEA 2005).
Drylands are described as non-equilibrium ecosystems (Behnke et al. 1993) and they deserve special attention because they display vulnerability-creating mechanisms to ecosystems and man (Sietz et al. 2011). Rainfall variability (Borgogno et al. 2007; D'Odorico & Bhattachan 2012), climate change (Dougill et al. 2010) and fast growing populations in dryland regions (Enfors & Gordon 2007) are only some aspects to mention. Many of them are home made because man affects land use and land cover significantly. And it is quite evident that the landuse and landcover in a certain area is decisive for a catchment's hydrology (Chhabra et al. 2006). This includes especially groundwater quality and recharge (de Fries & Eshleman 2004; Scanlon et al. 2005; Ahearn et al. 2005), river discharge (Poff et al. 1997; Costa et al. 2003; Schilling et al. 2010), riverine and riparian ecosystems (Allan 2004) and nutrient balances and chemical pollution (Sliva & Dudley Williams 2001; Lehrter 2006).
The majority of drylands (globally) is covered with rangeland and so livelihood strategies, pastoralism and mitigation to shifts in the climate regime are closely connected (Rockström 2000; Thomas 2008; Falkenmark & Rockström 2008). Recent research just revealed that some ecosystems show a higher degree of resilience to external stressors (Bestelmeyer et al. 2013; Stromberg et al. 2013). But since almost half of the Earth's solid surface can be classified as dryland (Tooth 2000) and the majority of the world's population faces the threat

of precarious water security (Vörösmarty et al. 2010) it is fundamental to manage the scarce resource.

Water Management

As well as a land manager, a farmer and a politician have different opinions on what watershed management might be, the scientific community is at strife about what might be the right way of managing water. Nevertheless all of them might agree on the emerging issues and challenges that arise from population growth and higher living standards (Bouwer 2000). The idea to manage the resource water can be traced back several centuries (Butzer et al. 1985; Ferrio et al. 2005). But the modern approach has its origin either in the spanish *confederaciones hidrográficas* from the 1920s or in the water resources plan of the Tennessee Valley Authority in the 1940s (Rahaman & Varis 2005).
The topic of water management gained more attention since the 1970s when the first UNESCO International Conference on Water took place at Mar del Plata, Argentina 1977 (Jeffrey & Gearey 2006). 15 years later on a similar conference in Dublin the guiding principles for a responsible treatment of water resources were formulated in The Dublin statement on water and sustainable development (IWCE 1992):

- Fresh water is a finite and vulnerable resource, essential to sustain life, development and the environment.
- Water development and management should be based on a participatory approach, involving users, planners and policy-makers at all levels.
- Women play a central part in the provision, management and safeguarding of water.
- Water has an economic value in all its competing uses and should be recognized as an economic good.

Since the 1990s two competing approaches arose: Integrated Water Resources Management (IWRM) and Integrated Watershed Management (IWM) and both are very closely related. In parts it cannot be distinguished which author follows IWM and which IWRM. The main principles apply to both and can be summarized as follows:
IW(R)M is a holistic approach that comprises socio-economical, environmental and political aspects (e.g. Ffolliott et al. 2003; van der Zaag 2005). It is a framework and a tool for the purpose of development (Förch & Schütt 2004) of all resources within a watershed on a watershed-scale (Walmsley et al. 2001; Ffolliott et al. 2003) because like everywhere else in the geosphere "everything is related to everything else, but near things are more related than distant things" (Tobler 1970). It has a strong focus on sustainability (e.g. Jaspers 2003; Biswas 2004; Rockström et al. 2004) and is always meant to be a participative approach that interlinks all locals – from water users, politicians to administration and scientists (Thomas

& Durham 2003). This especially includes capacity building and traditional knowledge (e.g. Förch & Schütt 2004; Evans & Varma 2009).
There is critique about the approach because sometimes it just fails for various reasons (Heathcote 2009) and despite the broad fostering from the international scientific community and from the UNEP (2012) the approaches lack a wide implementation that encompass more than an experimental stage (Blomquist & Schlager 2005). Moreover Jeffrey & Gearey (2006) state that the complexity of the field of managing a socio-economical unit like a watershed puts the holistic manner under tension. Additionally the holistic nature of IW(R)M makes all the formulation of the concept vague and blurry (Biswas 2008) and the underlying concept of stable natural systems fluctuating within a certain variability also known as stationarity is considered obsolete (Milly et al. 2008).

Collecting Techniques

Natural water sources in drylands can be roughly classified into autogenic and allogenic ones (Beckers et al. 2013a). First describes a locally generated source like rainfall, local runoff or ephemeral streams like *wadis* (Roberts 1977). The latter specifies fossil groundwaters or major streams originated in humid regions flowing through arid areas as a main source of water (Bull & Kirkby 2002; Woodward 2009). In Drylands usually water is the most fundamental limiting factor of agricultural production. Thus special techniques were established to mitigate the seasonal scarcity of water, that implies

- special ways how to get water,
- how to store water and
- how to irrigate

and a combination of all three points.
The basic principle to get water is to deprive a part of the land of its share of rainwater which is very small and usually non-productive and add it to another piece of land. Therefore the amount of water available comes closer to the crop water requirement or any other beneficial use (Oweis & Hachum 2006). That approach has a long tradition in almost every (semi)arid environment in the world (e.g. Ferrio et al. 2005; Mächtle et al. 2009; Puy & Balbo 2013; Beckers et al. 2013b).
Water harvesting and storage of water is closely related and sometimes interdepend. Boers et al. (1986) show that water harvesting using a micro-catchment is very successful in desert margins. The key is a relatively large micro-catchment thats runoff is routed to a specific point where a small amount of crops benefit. A special case is the harvest of rainwater. It is cheap, simple and in combination with various soil conservation measures it is possible to improve agricultural productivity (Vohland & Barry 2009). Especially in urban areas of less developed countries

or in areas where conventional water supply fails the approach to harvest rainwater from the rooftops or similar planes not in use is widespread and promising (Handia et al. 2003; Abdulla & Al-Shareef 2009).
Runoff farming water harvesting is a method of collecting surface runoff water using channels, dams and diversion systems. Soil crusts increase the effect because the loss of water through infiltration is reduced (Abu-Awwad & Shatanawi 1997). Either the water is stored in a reservoir or routed directly into the root zone of a farmed area (Boers & Ben-Asher 1982). It is common that runoff farming is coupled to dryland geomorphology (e.g. Giráldez et al. 1988; Beckers & Schütt 2013).
Another widely used technique is the pumping of groundwater. It came up during the last 50 years especially where agriculture is largely industrialized (Foster & Chilton 2003) and is considered to be the key component of water resources in regions with water scarcity (Hernández-Mora et al. 2001). It makes a huge difference if the aquifers are of fossil (de Vries et al. 2000) or modern origin (Rodell et al. 2009). Recently the development goes into the direction of recharging aquifers (Scanlon et al. 2012).

Irrigation Techniques

According to Kreeb (1964) irrigation methods can be classified as

1. flood irrigation system,
2. furrow irrigation,
3. subirrigation and
4. sprinkler irrigation.

Flooding the meadows is a natural behavior of a fluvial system (Wolman & Leopold 1957). These predictive characteristics have been used for agriculture since early Quaternary and one of the widest known example is the Nile and its floodplains (Lawton & Wilke 1979; Hassan 1986; Kuper & Kröpelin 2006). The controlled flooding is closely related and derived from the above mentioned method. The agricultural field is either criss-crossed by little dams and channels to route the water from the well to the other side of the field. This is mostly restricted to one season. Or an enhancement of that technique is a more or less permanent tessellated pattern of fields (Clemmens 1981).
The reduction of the flow accumulation is another widely used technique. This is usually done by slowing down the water on its way down a slope and thus prolonging the time to infiltrate. The contour-parallel basins equal the agricultural fields (Khanna et al. 2003). A special case of that method are terraces like in the Mediterranean (Thompson & Scoging 1995; Stanchi et al. 2012) and on various types of Asian rice fields (Bambaradeniya & Amarasinghe 2003; Lennartz et al. 2009).

The furrow irrigation method consists of a dense network of small channels so that water bypasses the plants. Usually the plants do not have contact to the water since bed-planting is practised (Kreeb 1964; Fahong et al. 2004). Subsurface irrigation has a long tradition (Lightfoot 1996). Perforated pipes are buried and the water is routed to the rooting zone directly. Recent development tends to high precision subsurface drip irrigation that applies water to the root zone directly (Kandelous et al. 2012; Gao et al. 2014).
Very common – even in humid climates – are sprinklers. They are cost and water intensive which implies an usage in the field of agro-industrial crop production. A special form of sprinklers are center pivot irrigation systems like shown in Figure 2.3a. Recent development shows a diversification and an increasing trend towards adaptability and site specific gear (Moreno et al. 2012; Valín et al. 2012). All of the mentioned methods hold various subcategories and varieties and its distinctions are transient (Kreeb 1964).
Many of the irrigation techniques suffer from severe disadvantages that affect dryland environments in a special way. These are extreme evapotranspiration losses up to 85 % (Greenwood et al. 2009), negative impact on the water quality in the aquifers by fertilizer mismanagement (Cassel Sharmasarkar et al. 2001), soil salinization (Xie et al. 2011) and depending on the technique erosion of invaluable topsoil (Gurbanov 2010; King & Bjorneberg 2011; Lehrsch et al. 2011; Boulal et al. 2011) or the depletion of groundwater resources because of overabstraction (Konikow & Kendy 2005).

3.3. Droughts & Drought Indices

Droughts

Droughts can affect plant's growth in general (e.g. Buell et al. 1961; Grant 1984) and their metabolism in particular (e.g. Pandey et al. 2003; Fresneau et al. 2007). That results in high agricultural and thus economical losses (e.g. Liverman 1990; Endfield & Tejedo 2006). In order to quench global food demand the design of drought resilient plants is discussed, but their practical implementation seems to be in distant future (Farooq et al. 2009). At worst droughts lead to conflicts between competing water users and/or can be a severe threat to existence of life (e.g. Brooks 1971; Robertson 1986; Olsson 1993). Traditional communities deeply rooted in drylands developed strategies to cope with recurring droughts, but on the contrary if these strategies fail the impact can be even worse (e.g. Caldwell et al. 1986; Oba & Lusigi 1987; Nyong et al. 2007; Rukema & Simelane 2013). Furthermore droughts have an impact on hydrologic regimes of a river what on the one hand controls river morphology (Hakala & Hartman 2004) and thus on the other hand influences riverine ecosystems (Acuña et al. 2005; Power et al. 2008).

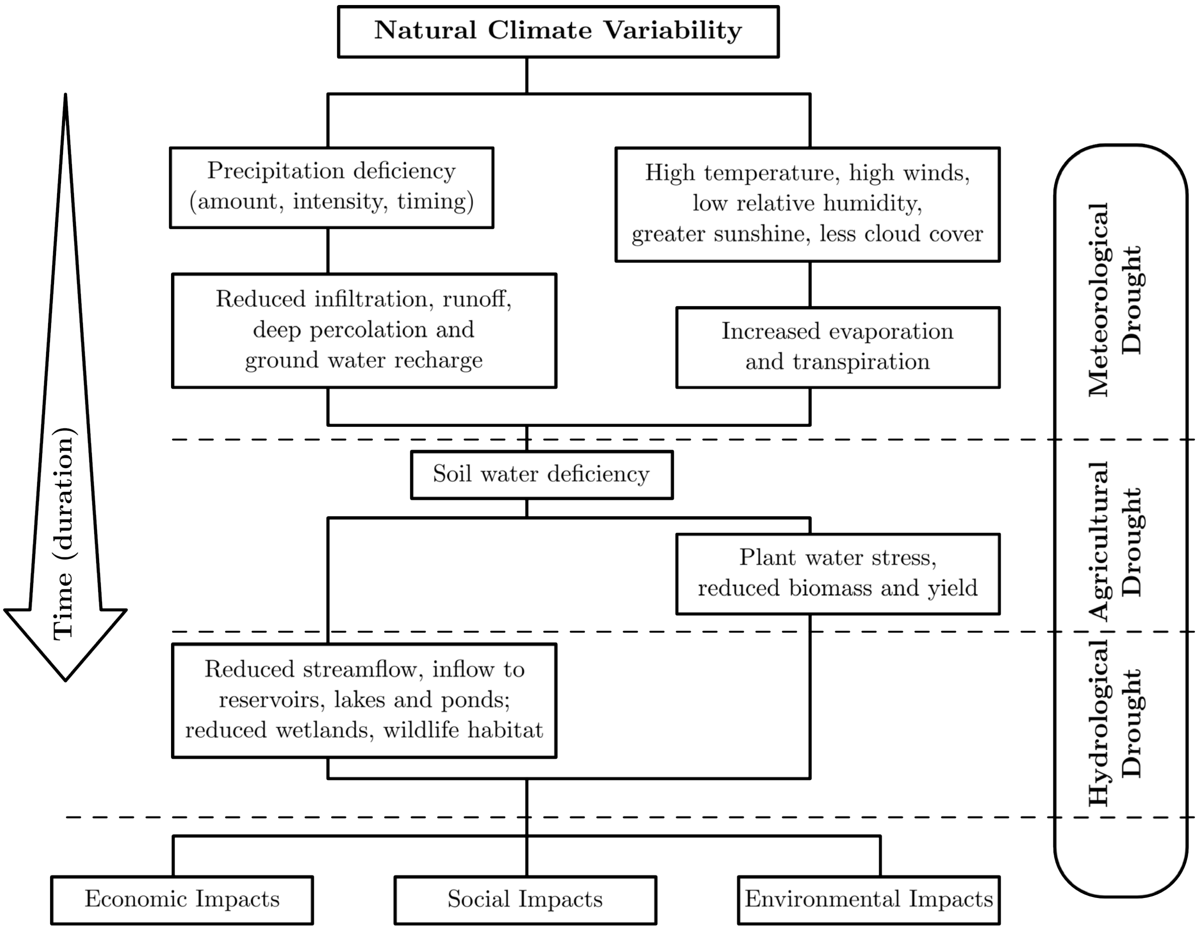


Figure 3.1: Types of drought as presented by NDMC (2015). Note the cascading principle of the phenomena as all droughts result from a deficit in precipitation but other types of drought can result from such an initial drought.

To define drought is a difficult task. This difficulty is fourfold:

- different views on droughts from different disciplines,
- different definitions in different parts of the world,
- established drought definitions and studies often do not cover precipitation **and** runoff
- and a lack of symmetry in hydrologic terminology (Dracup 1991).

Mostly drought refers to a long-term balance between rainfall and evapotranspiration in a particular area that is referred to as normal. Any imbalances do either reflect drought or above-normal humid conditions (Wilhite & Glantz 1985). A very broad definition according to Matalas (1963) is that a drought is an extended period of dryness. It is considered to be a "creeping phenomena" since the exact onset and end can not be distinguished precisely (Gillette 1950). Droughts are recurrent features of climate that occur in all regions of the

world (Lloyd-Hughes & Saunders 2002), but the likelihood to be observed in regions with relatively high interannual rainfall variability is much bigger. In such arid and semiarid areas recurrent droughts are part of the climate and are not unexpected events (Glantz & Katz 1977). That is why the temporal resolution differs according to climatological regime under study (Hisdal & Tallaksen 2000).
There are four widely accepted types of drought. A meteorological drought is like any other drought specific to a certain region. Prevailing atmospheric conditions that result in deficiencies of precipitation and/or high evapotranspiration rates cause such a drought (Palmer 1965). A hydrological drought affects two compartments of the hydrological cycle – streamflow and groundwater. It is considered a hydrological drought when the established uses can no longer be supplied in the current management system (Wilhite & Glantz 1985). Distinction must be drawn between a streamflow drought and a low flow. It can be part of a drought, but it is also part of the normal seasonal discharge hydrograph (Hisdal et al. 2000). Only major meteorological droughts can affect hydrology, because of the relatively high latency of the system (Peters & van Lanen 2000; Hisdal & Tallaksen 2003).
Agricultural droughts in comparison to hydrological ones occur in larger areas and their impact can not be measured precisely until the crops are harvested. Generally this type refers to the soil moisture and not to the whole hydrological cycle and its water content. Hydrological and agricultural droughts are considered to be manifestation of meteorological droughts while meteorological droughts are just indicators for precipitation deficiency (Andreadis et al. 2005; Boken et al. 2005). Socio-economic droughts result from at least one of the above mentioned droughts. They are associated with the supply and demand of some economic good (Alvarez & Estrela 2000).

Drought Indices

The characterization of a drought depends on the foci of the study and the perspective of the beholder and is hence a controversial undertaking (Paulo & Pereira 2006).
A widely applied index was formulated by Palmer in 1965 what later became the Palmer Drought Index (PDI) and Palmer Drought Severity Index (PDSI) respectively. It is based on the supply and demand concept of water and integrates rain and temperature in a series of water balance terms. The hypothetical soil moisture content is derived and compared to a reference water balance term. This comparison results in a dimensionless index (Keyantash & Dracup 2002). Several Palmer-based indices evolved like the Palmer Hydrological Drought Index (PHDI) (Karl 1986), the modified PDSI for real time monitoring (Heddinghaus & Sabol 1991), and the self-calibrated PDSI (sc-PDSI) where empirical constants are replaced by dynamically calculated values for the specific area (Wells et al. 2004).

The PDSI-family suffers from several limitations. It has a fixed temporal scale what makes it difficult to work on hydrological droughts. It assumes rain to be the only precipitation (it is not multiscalar) so the results from mountainous regions and from winter months are at least questionable. Additionally runoff is underestimated because of the assumption that all soil layers are saturated before runoff occurs. And the PDSI responds slowly on developing and decreasing droughts (Alley 1984; Vicente-Serrano et al. 2010a).

Another virtual standard to quantify droughts – especially in North America – is the Standardized Precipitation Index (SPI) (McKee et al. 1993). The index is based on a time-series of precipitation that is fitted to a probability distribution of precipitation. Then it is transformed to a normal distribution so that the mean SPI is zero. It compares the precipitation values of a given year with the past year regarding the median precipitation in order to indicate flood or drought years (Edwards 1997; Lu 2011). The SPI outplays the PDSI because of the possibility to cover different time scales but it suffers from a severe disadvantage: it relies on precipitation as input data only so that other variables that can affect droughts are neglected (Vicente-Serrano et al. 2012a).

To overcome the shortcomings of SPI and PDSI (and its derivates) the Standardized Precipitation Evapotranspiration Index (SPEI) was developed (Vicente-Serrano et al. 2010a; Vicente-Serrano et al. 2010b). SPEI relies on a water balance from $\mathrm{D_i = P_i - PET_i}$ to measure a deficit or surplus of water. The following procedure agrees with the computation of SPI. Direct comparison of SPI and SPEI shows the improvement and the supremacy of SPEI (McEvoy et al. 2012). Vicente-Serrano et al. (2012a: p. 5) summarize that the "SPEI, based on precipitation and potential evapotranspiration, combines the sensitivity of PDSI to changes in evaporation demand, caused by temperature fluctuations and trends, with the simplicity of calculation and the multitemporal nature of the SPI."

There are many more indices dealing with droughts like RAI (Rainfall Anomaly Index) (van Rooy 1965), SWSI (Surface Water Supply Index) (Shafer & Dezman 1982), and the RI (National Rainfall Index) (Gommes & Petrassi 1994). Many of them are summarized by Byun & Wilhite (1999), Heim (2002), and Mishra & Singh (2010).

3.4. Modelling of Soil Erosion

Soil Erosion

Soil erosion in principal is a degradative biophysical process that is exacerbated by socio-economic and political factors (Lal 2001). Soil erosion has a twofold effect: on-site and off-site. On-site effects include redistribution of soil material on fields, loss of soil, breakdown of soil structure, decline of soil organic matter and nutrients which means reduction in cultivable soil

depth and decline in fertility. Altogether it leads to loss of productivity that either needs to be compensated by fertilizers or it leads to abandonment of land (Morgan 2009). Off-site effects seem less obvious but clogging of drainage ways, siltation of rivers results in higher costs for facility maintenance, facility replacement or mitigation and prevention which imposes signifant costs upon society (Moore & McCarl 1987; Pimentel et al. 1995). Additionally new arable land needs to subdued as compensation.
Soil erosion is a natural phenomena on a temporal low scale. However lots of past agricultural civilizations were confronted with the issue of increasing soil erosion rates (Lal 1998). And since the agriculture became industrialized the soil erosion rates in the respective parts of the world increased dramatically (Pimentel et al. 1987; Montgomery 2007). Approximately 80 % of the world's agricultural land suffers from moderate to severe erosion (Oldeman 1994). Cerdan et al. (2010) report on mean erosion rates for different landuse classes in Europe with a mean of 17.4 t ha^{-1} yr^{-1} in vineyards and orchards. Under Mediterranean conditions in a similar landuse class Kosmas et al. (1997) show much lower values for vines.
While some authors show severe effects on the soil properties by overgrazing (Oztas et al. 2003) other show that dryland grazing systems are dynamic ecosystems where soil is not necessarily degrading by exceeding carrying capacities (Rowntree et al. 2004).

Models and Approaches

Models and simulations are closely related. A model is considered to be a set of assumptions about a system under study. It consists of a general theory and a special description of an object. It is highly abstracted and sometimes reduced to fundamental characteristics (Rosenblueth & Wiener 1945). On the contrary a simulation is accepted to be the implementation of dynamic models where all subsystems and variables are solved (Hartmann 2005). During the last decades modelling became a virtual standard method in almost every scientific disclipine like political sciences (Saperstein 1984), epidemiology (Okongo et al. 2013) or astronomy (Hammer et al. 2010) just to name a few.
There are plenty of modelling approaches. The most relevant ones in the field of soil erosion are empirical models, conceptual models and physical-based models according to Wheater et al. (1993). Empirical models are derived from empirical observations solely. There is no need for assumptions concerning the relationships between variables or on physical principles. A conceptual model is like a framework including all dependent and independent variables in the system and their relationships. Usually conceptual models lump the processes over the modelled scale (Merritt et al. 2003) and their parameters are a result of calibration against empirical data (Abbott et al. 1986). Conceptual models play an intermediate role between empirical and physical-based models (Merritt et al. 2003). Physical-based models are based on the

understanding of the physical principles and processes that cause effect on a system. Thus they are very flexible and robust but hard to parameterize (Lilly et al. 2009). The transition from one type to another is continuous because many variables are measured or empirical and a concept is the foundation of all models (de Vente & Poesen 2005).
Traditionally a model is not necessarily related to computation. The earliest works on the estimation of soil erosion took place in the 1930s and 1940s and the models were crude and all the calculation was conducted by hand (Renard et al. 1997a). But especially during the last decades most of erosion modelling is conducted using either GIS or various computer programs and languages (e.g. Hofierka et al. 2002a; Finlayson & Montgomery 2003; de Vente et al. 2009).
There are plenty of erosion models following different approaches. To name a few: SLEMSA (Soil Loss Estimation for Southern Africa) (Elwell 1978), MMF (Morgan Morgan & Finney Method) (Morgan et al. 1984), EUROSEM (European Soil Erosion Model) (Morgan et al. 1998), SHE (Système Hydrologique Européen) (Abbott et al. 1986) and WEPP (Water Erosion Prediction Project) (Nearing et al. 1989). Overview of presumably not all erosion models is given by de Roo (1993), Jetten et al. (1999), Merritt et al. (2003), Jetten & Favis-Mortlock (2006), and Govers (2011).

USLE and USLE-Derivates

The Dust Bowl in the US of the 1930s was a tremendous environmental catastrophe (Schubert et al. 2004). Amongst others wrong management of agricultural soils triggered a vast land degradation (Cook et al. 2009). Since that time the endeavor to ascertain the reasons why and how soils degrade was triggered and systematic research began (USDA 2015). After the development of some predecessors in the 1940s and 1950s the well-known Universal Soil-Loss Equation (USLE) was developed in the late 1950s with empirical data and presented in an Agricultural Handbook in 1965. It is meant to be generally applicable to the geographic and climatic regions east of the Rocky Mountains (Wischmeier & Smith 1965). Later it was enhanced and revised to overcome some of the limitations (Wischmeier & Smith 1978). It is the product of the factors

- climate erosivity – R,
- topography – L and S
- soil erodibility – K and
- landuse and management – C and P.

The R factor is directly proportional to the EI_{30} index. It reflects the kinetic energy times the maximum intensity of a thirty minute storm event (Wischmeier & Smith 1978; Renard & Freimund 1994). Since this is a very data demanding index and there are only a few

spots in the world with such temporal and spatial resolution alternative procedures are usually used to derive R – especially derivates from monthly data (e.g. de Santos Loureiro & de Azevedo Coutinho 2001; Angulo-Martínez & Beguería 2009). R was developed mainly for the continental US east of the Rocky Mountains and thus using RUSLE with R in other parts of the world is considered to be vague (Renard et al. 1997b). But there are plenty of trials to adapt R to other climatic regions (e.g. Millward & Mersey 1999; Wang et al. 2002; Hoyos et al. 2005).
A very common method is to use the correlation of R and the MFI (Modified Fournier Index) as Arnoldus (1977) showed. Other approaches include the relationship of EI_{30} to total rainfall, effective rainfall (exluding the first 12.5 mm of rainfall and consider them as non-erosive due to infiltration) or the so-called Burst Factor (Smithen & Schulze 1982). Most of the equations to calculate R overestimate the kinetic energy in comparison to direct measurement in coastal influence and underestimate it in sub-humid to semiarid locations (van Dijk et al. 2002). Additionally Mikoš et al. (2006) showed that most equations are not necessarily valid in alpine environments.
The topography factor LS consists of the slope length (L) and the slope factor (S). The original USLE is plot based with a length of 22.1 m (72.6 ft) and a width of 1.8 m (6.1 ft) at an inclination of 9 % (Renard et al. 2010). The initial LS was an empirical ratio chart of the expected soil loss per unit area on a field slope to the corresponding loss on a standard plot (Wischmeier et al. 1958; Wischmeier & Smith 1965). The first revision of USLE in 1978 brought a different procedure. The slope length factor was defined as $\mathrm{L} = (\lambda/72.6)^{\mathrm{m}} \times (65.41 \sin^2\theta + 4.56 \sin\theta + 0.065)$ where λ is the length of a slope, θ is the slope, 72.6 is the standard plot length (ft) and m is the slope length exponent according to four slope classes (bigger than 5 %, 3.5 % – 4.5 %, 1 % – 3 % and less than 1 %). This is only valid for uniform slopes. The exponent m varies from 0.2 to 0.6 (Wischmeier & Smith 1978). Mostly because of the variations in experimental results and the poor explanation of these relationships LS in general and m in particular remained controversial (Laflen & Moldenhauer 2003).
Soil erodibility is complex. Bryan & Poesen (1989) state that the concept of soil erodibility occurs as a vague, empirical summation of the erosional response of soils. It was thought for a long time to be a constant that might be easily measurable, but it is more a set of highly complex response patterns that is strongly influenced by intrinsic soil characteristics and by macroenvironmental variables extrinsically (Bryan 2000). It is a measure of soil susceptibility to interrill and rill erosion. The factor is defined as the soil loss rate per erosivity index unit on the USLE standard plot (Römkens et al. 1997). In practice it is the integrated long-term average soil response to the erosive power of rainstorms (Torri et al. 1997).
In this context the soil erodibility (K) is considered one of the most important factors that is hardly understood but always in use (Auerswald et al. 2014). The original procedure

to estimate K is either the nomograph by Wischmeier & Smith (1978) or its algebraic approximation (see equation 4.13 in chapter 4.3.3) where silt does not exceed 70 %. Alternative approaches almost always make use of the soil particle size distribution and soil organic carbon (Zhang et al. 2008) or of the application of geostatistical methods on K values from soil samples (Wang et al. 2001). The soil erodibility factor works well for regions where the empirical relationships to derive the equation of K were collected (which is the Midwest of the USA) (Bork & Schröder 1996), but it needs calibration and conversion for soils in other parts of the world (e.g. Rejman et al. 1999; Lu et al. 2001; Beskow et al. 2009).
The USLE and its factors is the foundation for many derived and adapted formulas to calculate soil erosion.
In MUSLE the rainfall energy factor R is replaced by a runoff factor that needs calibration to a gauged basin to determine a soil moisture coefficient. Thus it is always restricted to a basin scale but it can differ in spatial extent (Williams 1977). MUSLE is applicable on different time scales (annual, monthly, daily) and to specific storm events (Zhang et al. 2009) and is said to be superior in steeply forested areas (Sadeghi et al. 2007). MUSLE is not to be confused with USLE-M after Kinnell & Risse (1998). USLE-M also differs in the treatment of the erosivity index (R). The runoff amount and sediment concentration is introduced as an exclusive term that is coupled to time – the result is an erosivity index per event ($Q_R EI_{30}$) after Kinnell (1997). So USLE-M uses an event based erosion index to compute USLE (Kinnell 2001).
Presumably the most popular model is the revised USLE (RUSLE). It follows the same empirical principles as the original USLE. The factors are identical but their computation differs. Additionally the computerization was taken into account and an user-friendly software was developed and distributed (Renard et al. 1991). Nowadays speaking of RUSLE means either the model and its empirical relationships or a distributed software. There is the RUSLE1 that is very close to the original USLE factors and there is RUSLE2 that is still developed and maintained. That includes frequently calibration, removal of possible inconsistencies in the equations and implementation of latest scientific approaches (Renard et al. 2010). Another distinction between USLE and RUSLE is the treatment of LS. In USLE the recommended values of the slope length exponent (m) vary from 0.2 to 0.5 for slope levels between 1 % and 5 % and there is no change beyond an inclination above 5 %. In RUSLE m is a function of the erosion ratio of rill (caused by flow) to interrill (caused by raindrop impact) and m continues to increase with increasing slopes (McCool et al. 1997; Gabriels 1999; Hoffmann Oliveira et al. 2013). One of the biggest advantages of RUSLE is the possibility to apply the equation to catchment scale because USLE is restricted to fields solely (Rodríguez & Suárez 2010).
The Areal Nonpoint Source Watershed Environment Response Simulation (ANSWERS) model (Beasley et al. 1982) is designated for the use in catchments with agriculture as primary land use during and shortly after rainfall events. It uses USLE K, C and P but additionally several

physically based mathematical relationships to represent field capacity, potential interception, soil moisture and several other soil variables (DE ROO 1993). This makes it superior to USLE in many ways but it has the disadvantage of theoretical weaknesses and the complexity of required input data (DE ROO et al. 1989).
The LS factor has been a widely discussed topic because it lacks a proper representation of real life topography (GRIFFIN et al. 1988; HICKEY 2000). Additionally several authors agree that the topography factor is the most sensitive one (e.g. BRYAN & POESEN 1989; LIU et al. 2001; LIU et al. 2011; KUMAR & KUSHWAHA 2013). To overcome some shortcomings of LS the RUSLE3D model uses the concept of unit stream power to compute the sediment transport capacity (T) based LS as follows: $\mathrm{T} = (\frac{\mathrm{A}}{22.13})^m \times (\frac{\sin(\beta)}{0.0896})^n$ where A is the upslope contributing area and β is the slope angle (MITASOVA et al. 1996a). RUSLE3D is a detachment limit method. That implies that a water flow transport can transport an infinite amount of sediment. Thus RUSLE3D is not capable to calculate deposition (MITASOVA et al. 1999). Close to RUSLE3D is USPED (Unit Stream Power Erosion Deposition) with the assumption that water flow can transport a limited amount of water (transport capacity limited) (MITASOVA et al. 1996b; MITASOVA et al. 1999). USPED substitutes LS from RUSLE with $\mathrm{LS} = \mathrm{A}^{\mathrm{m}} \times (\sin\beta)^{\mathrm{n}}$. Further computation is needed since USPED has the advantage to compute erosion and deposition as the change in transport capacity (WARREN et al. 2005).

4. Data Used & Methods Applied

The field work took place in January and February 2014. All data were processed using R version 3.1 (R Core Team 2014), several additional software packages as stated in Table 4.1 and especially GRASS GIS Version 6.4.4 (Neteler et al. 2012; GRASS Development Team 2014). QGIS Geographic Information System Version 2.6 was used for cartographic purposes (QGIS Development Team 2014).
All laboratory works were conducted at the Laboratory for Physical Geography of the Institute for Geographical Sciences at Freie Universität Berlin. A generalized workflow of the working packages, input data and intermediate steps is depicted in Figure 4.1.

4.1. Climatic Variability

4.1.1. Climate Data

The used data sets are from the Agricultural Research Council (ARC 2014) which is in charge for the South African weather stations.
First weather station is Riviera which is exclusively in the middle of the catchment (see map in Figure 1.2). Despite the fact that the station exists since the early 1980s the obtained time series just ranges back till 2001. In addition the observations are partly incomplete and show severe gaps.
A second data set comes from the surroundings of Piketberg which is outside the catchment more upcountry (Figure 1.1). It shows less gaps but suffers from less observations. Fortunately it was possible to obtain daily rainfall records from the local farm Kromvlei (see Figure 1.2). It ranges back till early 2003 and shows no obvious data gaps.
All data sets range to January 2014. The two official stations give their measurements in floating-point numbers, data from Kromvlei are available as integers.
The above-mentioned data sets were considered to be not sufficient for some of the tasks since they cover 12 years maximum. The data set presented by the Climate Prediction Center & National Centers for Environmental Prediction (further called NCEP) was found to be sufficient for these tasks and has the advantages

- to be accessible freely,
- to contain the most relevant climate parameters,
- on a monthly basis from 1948 to 2011
- at spatial high-resolution (0.5°× 0.5° latitude/longitude)
- in native GeoTiff format (Fan & van den Dool 2004; Fan & van den Dool 2008).

Table 4.1: Package, version, references and brief description of used R packages for data processing.

Package	Reference	Description
Analytical Packages		
climatol 2.2	GUIJARRO 2013	Functions to homogenize climatological series and to list climatological summaries of the homogenized results
hydroGOF 0.3-8	ZAMBRANO-BIGIARINI 2014a	Goodness-of-fit functions for comparison of simulated and observed hydrological time series
hydroTSM 0.4-2-1	ZAMBRANO-BIGIARINI 2014b	Time series management, analysis and interpolation for hydrological modelling
pls 2.4-3	MEVIK et al. 2013	Special functions to calculate Partial Least Squares and Principal Component regression
raster 2.3-0	HIJMANS 2014	Reading, writing, manipulating, analyzing and modeling of gridded spatial data with basic and high-level functions
rgdal 0.8-6	BIVAND et al. 2014	Bindings for the Geospatial Data Abstraction Library (GDAL)
rgeos 0.3-6	BIVAND & RUNDEL 2014	Interface to Geometry Engine – Open Source (GEOS) using the C API for topology operations on geometries
SPEI 1.6	BEGUERÍA & VICENTE-SERRANO 2013	A set of functions for computing potential evapotranspiration and several drought indices including the Standardized Precipitation-Evapotranspiration Index (SPEI)
spgrass6 0.8.3	BIVAND 2013	Interface between GRASS 6+ geographical information system and R
zoo 1.7-11	ZEILEIS & GROTHENDIECK 2005	Infrastructure for ordered and irregular time series
Auxiliary Packages		
extrafont 0.16	CHANG 2014	Tools for using fonts in R
plotrix 3.5-4	LEMON 2006	A large number of specialized plots and accessory functions like color scaling, text placement and legends
plyr 1.8-1	WICKHAM 2011	Implementation of the split–apply–combine strategy for data analysis in R
RCurl 1.95-4.1	LANG 2013	General network client interface for R
stringr 0.6.2	WICKHAM 2012	stringr is a set of simple wrappers that make R's string functions more consistent, simpler and easier to use
xtable 1.7-3	DAHL 2014	Export tables to LaTeX or HTML

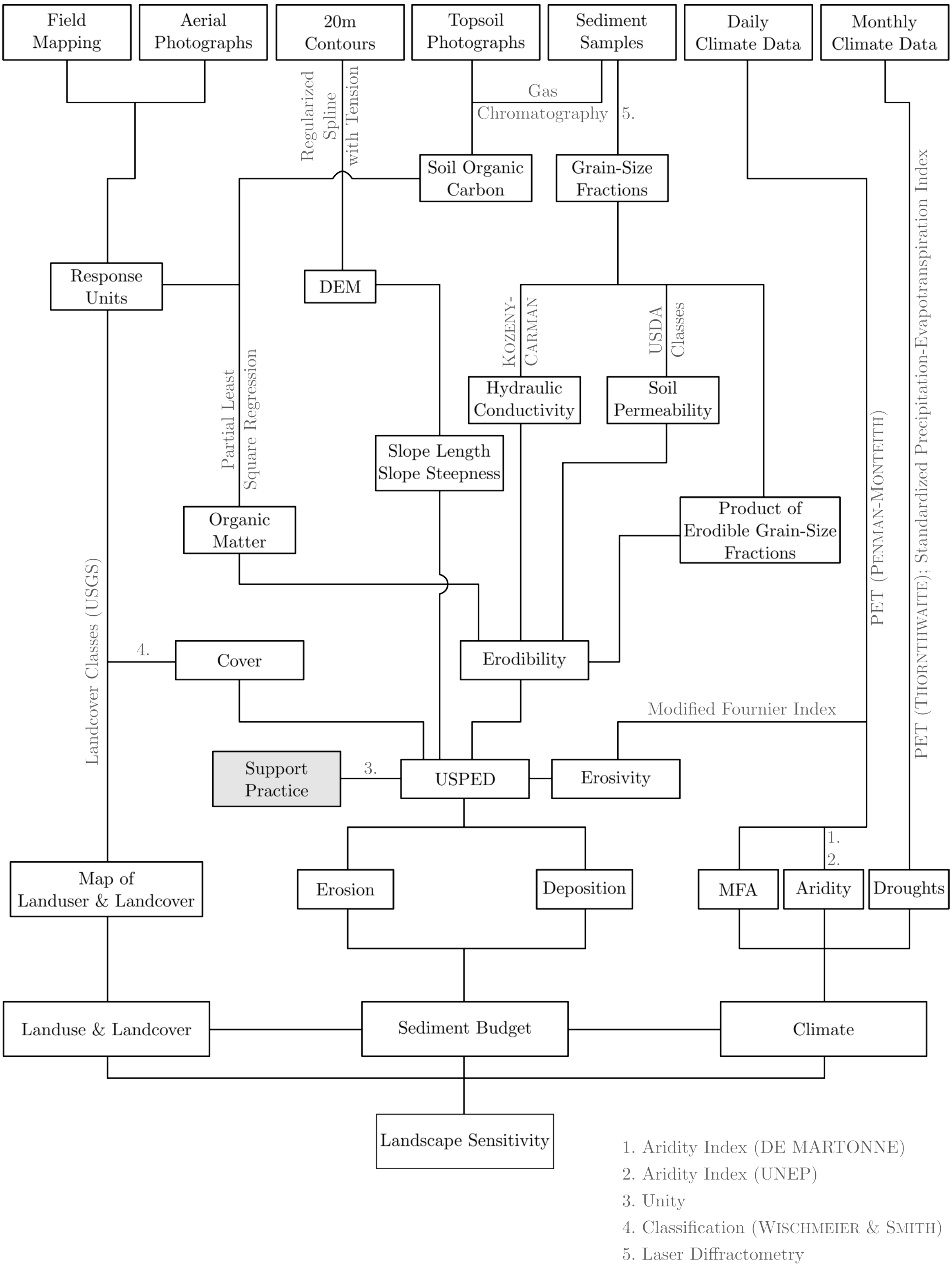


Figure 4.1: The workflow is showing the intermediate steps and procedures from input data at the top to results of the three main working packages at the bottom. Note that support practice was neglected.

To compute a drought index monthly rainfall and temperature data were obtained. The temperature data were used to derive potential evapotranspiration after THORNTHWAITE (1948).

4.1.2. Drought Index

In order to analyze the catchments vulnerability to the scarcity of precipitation a climatic drought index was calculated. Droughts are a spatio-temporal phenomena so if possible the time series of climate data should be as long as possible.
The Standardized Precipitation-Evapotranspiration Index (SPEI) (VICENTE-SERRANO et al. 2010a) was calculated using NCEP data. The SPEI is a simple multiscalar index using the difference between precipitation and potential evapotranspiration following equation 4.1 which is a straightforward adaption of a climatic water balance according to THORNTHWAITE (1948).

$$\mathrm{D_i} = \mathrm{P_i} - \mathrm{PET_i}\text{[1]} \tag{4.1}$$

To calculate the SPEI at different time scales a log–logistic distribution is used for the $\mathrm{D_i}$ values. The mean SPEI value is 0 with a standard deviation of 1. This means a SPEI of 0 represents 50 % of the cumulative probability of D (VICENTE-SERRANO et al. 2010a).

4.1.3. Aridity

In order to get an idea about the water balance in the course of the year aridity indices were calculated using Riviera station data:
The Aridity Index after DE MARTONNE is a classical index following the approach of the ratio of mean precipitation and mean temperature (QUAN et al. 2013). It was calculated for each month with equation 4.2 (DE MARTONNE 1926). P stands for precipitation (cm) and T is the mean temperature in °C. The index decreases with increasing aridity.

$$\mathrm{I_m} = \frac{12 \times \mathrm{P_i}}{\mathrm{T_i} + 10} \tag{4.2}$$

Since the Aridity Index after DE MARTONNE seemed very coarse for arid regions the UNEP Aridity Index additionally was calculated. The United Nations Environment Programme promotes the approach to use a ratio of precipitation (P) and potential evapotranspiration (PET) both in mm as in equation 4.3.

$$\mathrm{I_e} = \frac{12 \times \mathrm{P_i}}{\mathrm{PET_i}} \tag{4.3}$$

ET has been calculated using the formula from Penman (1948). All necessary values were part of the Riviera data set. The factor 12 is used to process monthly values.

4.1.4. Magnitude-Frequency-Analysis

Magnitude-frequency-analyses are used to estimate the probability and the recurrence interval (RI) of events. Actually this approach comes from the field of hydrology but it is suitable for rainfall data as well (Chow 1953). The daily precipitation values are sorted in descending order, the highest value gets the rang one. The recurrence interval is calculated with equation 4.4 where N is the total of observations (including the days without precipitation) and r is the rang of the corresponding rainfall amount (Ahnert 1982).

$$RI = \frac{N + 1}{r} \tag{4.4}$$

4.2. Landuse & Landcover

The research area was sub-divided into more or less homogeneous areas according to physical and environmental properties (Flügel 1997; Yair & Kossovsky 2002). As far as possible for these (Hydrological) Response Units (RU) the parameters vegetation type, vegetation cover, stone cover and soil type were determined. If distinguishable the landuse was also mapped in classes of USGS Land Cover Classes (Table 4.2) (Vogelmann et al. 2001, USGS 2014). As not all parts of the Krom Antonies river catchment could be mapped on a sufficient scale some parts were mapped on the basis of aerial photographs obtained from National Geo-Spatial Information (NGI 2010a).

The class open water is without exception due to the existence of water storage dams and reservoirs. Low intensity residential stands for agglomeration of houses and farm buildings. Barren rock & sediments refers to any surface which is not sufficiently covered by vegetation.

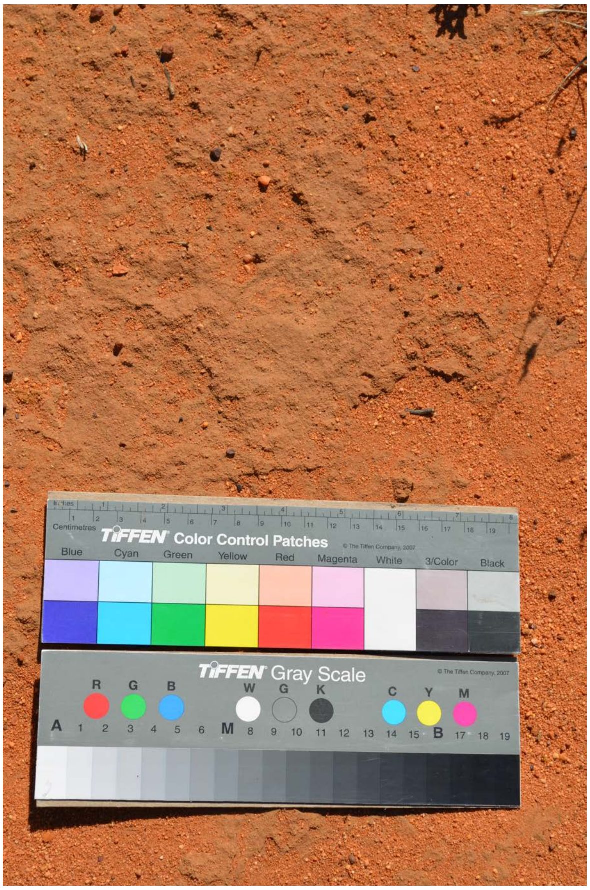


Figure 4.2: Example of a top soil picture and the corresponding color separation chart.

Shrubland is any non-agricultural land covered by shrubs whereas grassland stands for land without obvious agricultural usage but without shrubs. Orchards & vineyards are plantations, as pasture classified areas had to show traces of livestock farming. This includes horse breeding and related areas too. All agricultural used zones without obvious cultivated fruits were classified as fallow. Woody wetlands and herbaceous wetlands are mostly meadows or direct surroundings of water reservoirs.
Geomorphological mapping was neglected or conducted at very small scale considering the sheer size of the catchment. In addition certain areas were either not accessible due to steep slope and the result seemed not to be promising or the owners did not permit to enter.
Furthermore a picture was taken of most RUs showing topsoil and a color separation chart (example shown in Figure 4.2).

4.3. Soil Erosion & Deposition

The net erosion and deposition was modelled using the Unit Stream Power Erosion and Deposition (USPED) model (Mitas & Mitasova 1998). It is a 3-dimensional enhancement of the well established Universal Soil Loss Equation (USLE) after Wischmeier & Smith (1978) and follows in principle the same equation to derive the sediment transport capacity (T). The factors of equation 4.5 equal USLE and Water Erosion Prediction Project model (WEPP) (Flanagan et al. 2007) and were not developed for USPED exclusively (Mitasova & Mitas 2001).

$$\mathrm{T} = \mathrm{R} \times \mathrm{LS} \times \mathrm{K} \times \mathrm{C} \times \mathrm{P} \tag{4.5}$$

The transport capacity T is a product of rainfall and runoff erosivity (R), length and steepness of slope (LS), inherent soil erodibility (K), plant cover (C) and conservation support practice (P) (Warren et al. 2005).

$$\mathrm{ED} = \frac{\Delta(\mathrm{T} \times \cos\alpha)}{\mathrm{dx}} + \frac{\Delta(\mathrm{T} \times \sin\alpha)}{\mathrm{dy}} \tag{4.6}$$

The erosion and deposition is a change in the sediment transport capacity which is calculated as a directional derivative using equation 4.6 where ΔT is the change in transport capacity, α is the aspect angle and dx and dy are the first order partial derivatives from a Digital Elevation Model (DEM) in east-west and north-south extent respectively (Mitasova et al. 1996a). In the resulting raster map the positive values represent deposition whereas the negative values indicate erosion.

Table 4.2: Landcover Classes after USGS (2014). Classes in gray were not observed during mapping.

Level I	Level II	No.
Water	Open Water	11
	Perennial Ice/Snow	12
Developed	Low Intensity Residual	21
	High Intensity Residual	22
	Commercial/Industrial/Transportation	23
Barren	Barren Rock & Sediments	31
	Quarries/Strip Mines/Gravel Pits	32
	Transitional	33
Forested Upland	Deciduous Forest	41
	Evergreen Forest	42
	Mixed Forest	43
Shrubland	Shrubland	51
Non-Natural Woody	Orchards & Vineyards	61
Herbaceous Upland Natural/Semi-natural Vegetation	Grassland	71
Herbaceous Planted/Cultivated	Pasture	81
	Row Crops	82
	Small Grains	83
	Fallow	84
	Urban/Recreational Grasses	85
Wetlands	Woody Wetlands	91
	Herbaceous Wetlands	92

4.3.1. Rainfall-Runoff Erosivity

The estimation of rainfall-runoff erosivity (R) is an exceedingly difficult task. The original procedure needs the total storm energy (E) times the maximum 30–minutes intensity (I_{30}) to consider the volume of rainfall and runoff (E) as well as the rates of particle detachment and runoff (I_{30}) (Renard et al. 1997b).
Since this is not obtainable for most parts of the world several other ways must be taken into account to estimate R. As a good approximation to determine the rainfall erosivity the Modified Fournier Index (MFI – equation 4.7) was used (Arnoldus 1977) where p_i is the average rainfall

each month and P is the average annual rainfall. Precipitation data were obtained from the South African Ministry of Agriculture, Forestry and Fisheries. For further description see section 4.1.1.

$$\mathrm{MFI} = \sum_{\mathrm{i=1}}^{\mathrm{i=12}} \frac{\mathrm{p_i^2}}{\mathrm{P_a}} \tag{4.7}$$

To convert MFI to values of R equation 4.8 is suggested by Angulo-Martínez & Beguería (2009) where a and b are empirical coefficients with values of 21.56 and 0.927 respectively.

$$\mathrm{R} = \mathrm{a} \times \mathrm{MFI^b} \tag{4.8}$$

To simplify the approach and because for the other factors there are no data sets which reproduce the intra-annual variability the (annual) mean R value was taken for analysis.

4.3.2. Slope Length & Steepness Factor

Contour data were obtained from NGI (2010b) with various vertical resolutions. Resampling was conducted to get a set of contour lines at least common denominator of 20 m vertical distance. The contour lines were used to generate a Digital Elevation Model (DEM) applying a regularized spline with tension (RST) (Mitasova & Mitas 1993; Mitasova & Hofierka 1993).
This DEM is the basis for further computing the slope length and slope steepness factor (LS). Slope length using USPED can be calculated using equation 4.9

$$\mathrm{LS} = \mathrm{A^m}(\sin\beta)^{\mathrm{n}} \tag{4.9}$$

where m and n are empirical coefficients of 1.6 and 1.3 respectively depending on which form of erosional processes are considered (Mitasova & Mitas 2001; Pistocchi et al. 2002). In this specific case m has been set to the default of 1.6. β is the slope in degree [°] and A is the upslope contributing area (Mitasova et al. 1999).
The slope β is the first derivative (equation 4.11) of the DEM. Its gradient represents a vector field pointing in the direction in which maximal variation of the values occurs. Thus it reflects the maximal rate of change of elevation models (Olaya 2009).

$$\mathrm{v\bar{z}} = \sqrt{\frac{\delta \mathrm{z}}{\delta \mathrm{x}}^2 + \frac{\delta \mathrm{z}}{\delta \mathrm{y}}^2} \tag{4.10}$$

$$\mathrm{SLOPE}(\%) = \arctan(|\mathrm{v\bar{z}}|) \times 57.29578 \tag{4.11}$$

The upslope (contributing) area A is a planar area and not a three dimensional surface area. It describes the spatial extent of an area contributing to a single grid cell (GRUBER & PECKHAM 2009). It can be estimated using GRASS function `r.flow` when calculating the flowline density which equals the upslope contributed area per unit width, when multiplied by resolution (GRASS DEVELOPMENT TEAM et al. 2014a). In this case the horizontal resolution is 20 m as a result of the contour lines from the DEM generation.
A is the amount of mass that is accumulated in cell i when the sum of each neighboring cell's A is multiplied by the fraction r of each neighboring cell plus the input I in cell i itself. The receiving fraction r is the result of 1−d – the draining proportion (d) into its neighbors. In GRASS GIS a single flow algorithm (SFD) is used to compute the flow in the direction of the gradient down the slope. Since it is a SFD all flow is transported to a single cell downslope (MITASOVA et al. 1995). Further information on hydrological terrain attributes can be obtained at QUINN et al. (1991) and TARBOTON (1997).

$$\mathrm{A_i} = \sum_{j=0}^{8}(\mathrm{A_{NBj}} \times \mathrm{r_{NBj}}) + \mathrm{I_i} \tag{4.12}$$

4.3.3. Soil Erodibility Factor

For the calculation of the soil erodibility factor (K) the equation 4.13 from WISCHMEIER & SMITH (1978) was used where M is a fraction of grain size classes, OM is organic matter (%) and p is hydraulic conductivity. 0.137 is the conversion from imperial to SI units.

$$\mathrm{K} = 0.137 \times \frac{[2.1 \times 10^{-4} \times \mathrm{M}^{1.14}(12 - \mathrm{OM}) + 3.25(\mathrm{s} - 2) + 2.5(\mathrm{p} - 3)]}{100} \tag{4.13}$$

Soil Structure Class

The soil structure classes are defined according to WISCHMEIER & SMITH (1978). As a result of the field work's soil interpretation it was uniformly set to class 1 because of the ubiquitously found single grain fabric (USDA 2014).

Primary Particle Size Fractions

A total of 76 sediment samples were used for grain-size analyses using a LS 13 320 Laser Diffractometer of BECKMAN COULTER INC. (2009). After HCl-testing the samples no inorganic carbon was found so the ordinary pretreatment was neglected. The sediment fraction greater than 1 mm in diameter was dry sieved in order to filter most organic content.
The grain-size analysis is based on the general acceptance that particles of certain size or diameter diffract light at an unique angle (BEUSELINCK et al. 1999). This relationship is

inversely proportional. To measure this relation a suspension of sediment sample is routed through a steady beam of light (750 nm). The resulting scattering pattern is a function of light intensity and diffraction angle and is specific to each particle size. Detailed description can be found at de Boer et al. (1987), Loizeau et al. (1994), and Beuselinck et al. (1998) and Witt et al. (2012).
The primary particle size fractions (M) were calculated using the version of Wischmeier & Smith (1978). This includes the percentages of the silt (Si: 2µm – 50 µm), the sand (Sa: 50 – 100 µm) and the clay fraction (C: less than 2 µm) after USDA 2014.

$$\mathrm{M} = (\mathrm{Si} + \mathrm{Sa}) \times 100 - \mathrm{C} \tag{4.14}$$

Hydraulic Conductivity

The Kozeny-Carman formula (Kozeny 1927; Carman 1937; Carman 1956) after Odong (2007) was used to derive the hydraulic conductivity from grain-size analysis. In equation 4.15 g is the acceleration due to gravity 9.81 $\frac{\mathrm{m}}{\mathrm{s}^2}$, v is kinematic viscosity of water which was set to a standard 20 °C value of 1.002 according to Weast et al. (1988) and d_{10} represents the grain size diameter (mm) of the first decile.

$$\mathrm{K} = \mathrm{p} = \frac{\mathrm{g}}{\mathrm{v}} \times 10^{-3} \left[\frac{\mathrm{n}^3}{(1-\mathrm{n})^2}\right] \times \mathrm{d}_{10}^2 \tag{4.15}$$

n is the media's porosity which can be derived from empirical relationship (equation 4.16) using the coefficient of grain uniformity U calculated with the first and sixth decile (Vukovic & Soro 1992).

$$\mathrm{n} = 0.255(1 + 0.83^{\left[\frac{\mathrm{d}_{60}}{\mathrm{d}_{10}}\right]}) \tag{4.16}$$

After transforming the values from centimeters per second to inch per hour the permeability codes (see Table B 2) after Römkens et al. (1997) could be applied and the classified data was then attached to the spatial data.

Organic Matter

In order to calculate organic matter (OM) for equation 4.13 a representative picture (example in Figure 4.2) of most RU was taken showing top soil and a color separation chart. These pictures were read into R with rgdal (Bivand et al. 2014) and for selected coordinates within a region of obviously homogeneous soil the mean value (40 random measurements in a radius of ten pixels) of R, G and B were obtained using the function `over` from rgeos (Bivand & Rundel 2014).

The reference RGB values were acquired in the same way at the color separation chart without using mean values.

$$\begin{bmatrix} X \\ Y \\ Z \end{bmatrix} = \begin{bmatrix} 0.412453 & 0.357580 & 0.180423 \\ 0.212671 & 0.715160 & 0.072169 \\ 0.019334 & 0.119194 & 0.950227 \end{bmatrix} \times \begin{bmatrix} R \\ G \\ B \end{bmatrix} \tag{4.17}$$

The soil RGB values needed to be normalized. Since the pictures were stored in 8-bit integer JPG the measured values were divided by 255 to attain the reflectance in percent. Then it needed to be converted into another color space model. With the auxiliary color space CIE XYZ (CIE 1932) and the conversion factors in equation 4.17 (VISCARRA ROSSEL et al. 2006a) the RGB values were converted into CIE L a* b* color space (CIE 1978).
The standard illuminant D65 with $X_0 = 95.047$, $Y_0 = 100$ and $Z_0 = 108.883$ was conducted what is closest to normal daylight (POYNTON 2003).

$$\mathrm{L} = 116 \times \left(\frac{Y}{Y_0}\right)^{\frac{1}{3}} - 16 \text{ if } \frac{Y}{Y_0} > 0.008856 \tag{4.18}$$

$$= 903.3 \times \left(\frac{Y}{Y_0}\right) \text{ otherwise} \tag{4.19}$$

$$\mathrm{a}^* = 500 \times \left[\left(\frac{\mathrm{X}}{\mathrm{X_0}}\right)^{\frac{1}{3}} - \left(\frac{\mathrm{Y}}{\mathrm{Y_0}}\right)^{\frac{1}{3}}\right] \tag{4.20}$$

$$\mathrm{b}^* = 200 \times \left[\left(\frac{\mathrm{Y}}{\mathrm{Y_0}}\right)^{\frac{1}{3}} - \left(\frac{\mathrm{Z}}{\mathrm{Z_0}}\right)^{\frac{1}{3}}\right] \tag{4.21}$$

CIE L a* b* is a color space model using euclidian distances for color description. L stands for luminance and is valid between 0 (black) and 100 (white). a* and b* represent the chroma, where a* stands for the proportion of red (positive) and green (negative) and b* stands for blue (negative) and yellow (positive) (SCHANDA 2007; WARGALLA 2014).
To evaluate the top soil pictures a selection of 34 RU sediment samples was taken for further laboratory analysis. They were dried at 105 °C, dry sieved to less than 2 mm in order to extract coarse material like seeds, litter and pebbles and HCl-tested. To verify the absence of inorganic carbon in the samples the loss on ignition (LOI) was performened at 550 °C and 900 °C each for three hours (DEAN 1974).
The amount of total carbon (TC) was measured using the TruSpec CHN–S–Analyzer (LECO CORPORATION 2006). TC was assumed to represent total organic carbon (TOC) which equals

Table 4.3: Applied classification for RUSLE cover management factor (WISCHMEIER & SMITH 1978).

		Cover [%]			
		0	25	50	75
Canopy Type & Height	No appreciable canopy	0.45			
	Tall weeds or short brush with average drop fall height of 50.8 cm (20 inch)		0.36	0.26	0.17
	Appreciable brush or brushes, with average drop fall height of 198.12 cm (6.5 ft)		0.4	0.34	0.28
	Trees, but no appreciable low brush. Average drop fall height of 396.24 cm (13 ft)		0.42	0.39	0.36

C_{org} because of the lack of total inorganic carbon (TIC) from the LOI900. This could only be accepted because of the silicic parent rock material.
The C_{org} data were split into a calibration data set (21) and a testing data set (12) in order to develop a pedotransfer function (PTF) (VISCARRA ROSSEL et al. 2008).
This was done using a Partial Least Square Regression (PLSR) with the three components L, a^* and b^* and the leave-one-out cross-validated predictions (MEVIK & WEHRENS 2007). One sample had to be neglected because of the distorting effect to the other values. It was collected in a meadow and was almost completely humous. Organic carbon values where multiplied by 1.72 to appraise the surface organic matter (SCHWERTMANN et al. 1987). Finally the values for OM were written to the corresponding RUs.
Since that not all of the catchment was covered by reasonable RUs (either sketchy parts remained between RUs or the surrounding slopes were not accessible) some values needed to be interpolated. There are two different approaches that were used:
The hydraulic conductivity was derived from 76 samples in the agriculturally used part of the valley. But to cover the whole catchment some so-called dummy points had to be included. They were placed beyond the bordering outcrops that surround the horseshoe-shaped catchment.
The second approach had to be applied for the areas without attributes in between RUs. This was the case for organic matter, because not all RUs had a corresponding picture. The gaps were closed by randomly sampled points over a raster with calculated values. These points were used to process a bilinear spline.

4.3.4. Cover Management Factor & Support Practice Factor

The vegetation cover values from the RUs were reclassified to fit to the suggested cover classes by Wischmeier & Smith (1978) as in Table 4.3. The vegetation type was reclassified in equal measure. The dominant vegetation group was taken and used as determining factor.
Calculating the C factor brought the same issues as the K factor. The areas showing no sufficient values were treated as described above.
Due to the lack of data and the complexity of P it was set uniformly to one.

5. Results

5.1. Climatic Variability

5.1.1. Drought Index

3-month scale

The results of the Standardized Precipitation-Evapotranspiration Index (SPEI) calculation are plotted in Figure 5.1 and summarized in Table 5.1. The SPEI for the 64-year time series using a 3-month mean shows a strong seasonality. Altogether 85 months show SPEI values less than -1 and indicate moderate drought conditions. They occur throughout all twelve months of the year with absolute frequencies between six and ten, except for May with seven and October and November with no droughts. There are 15 severe droughts with SPEI values less than -2. October and November also do not hold droughts on this scale. Maximum counts of droughts are in March with three severe ones. All other months show one or two occurences.
Looking at the humid periods there are 91 months in the time series that show moderate SPEI values bigger than 1. The frequencies range from five to eleven. September shows no such events. The maximum comes up in November and May, the minimum of five wet spells is in July. 20 pronounced humid periods are observed in the data set. The most appear in April and July (3), September shows none again.
The whole time series holds periods of arid and humid conditions. In the first few years of the NCEP data set one event of severe drought conditions (greater than -2) can be observed. The 1960s and early 1970s as well as late 1990s and early 2000s appear as a phase of repetitive short-time droughts. Humid phases seem to be more distinct in the 1950s, mid 1970s, mid 1980s and mid 1990s.

6-month scale

On a 6-month scale the SPEI is less evenly seasonal than on the 3-month scale, but several distinct periods begin to evolve. For moderate drought conditions 91 events can be distinguished. Their range of absolute frequencies goes from no value for October and June and a minimum value of six in March to a maximum of twelve in September. No clear cumulation appears. Thirteen events of severe drought conditions can be stated for that climate records. June and October again do not show any drought events on that scale. The maximum value is in March (3), second most occurences are in July (2), all other months show one event.

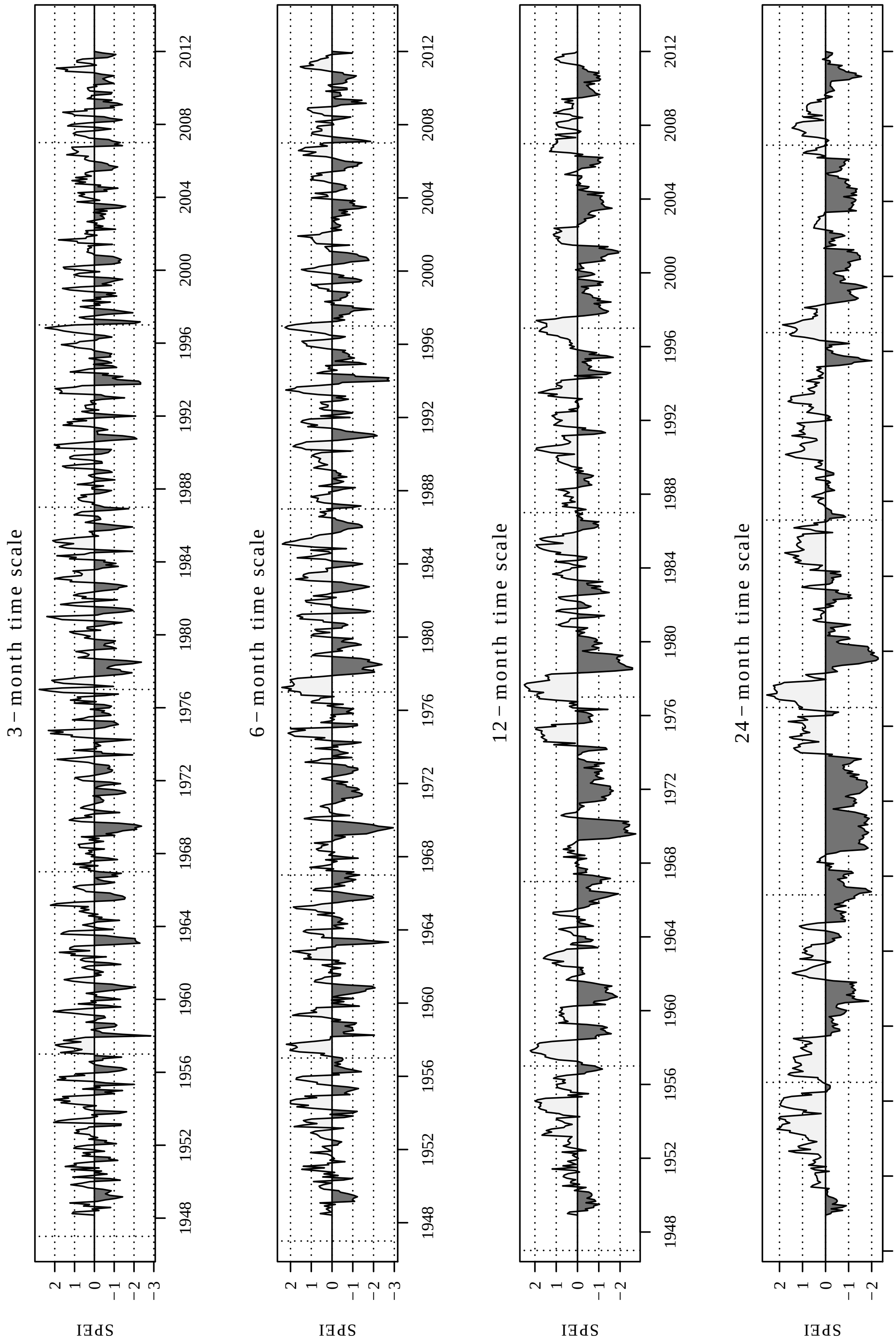


Figure 5.1: Results of the Standardized Precipitation-Evapotranspiration Index for 3-,6-,12- and 24-month scale.

99 moderate humid periods can be determined. The minimum occurence is in July with four moderate wet spells. Second lowest count is in September with six entries. The maximum occurs in May and December (11). There are two cumulations: the first from April to June, the second in November and December. The pronounced humid periods are distributed equally in the seasonal cycle. Most months bear a value of one or two. Maximum values are in February (3) and July (4).
Altogether the curve of the plotted SPEI values (Figure 5.1) shows an evolving trend. The 1950s appear as an above normal humid phase, as well as the mid 1970s and to a lesser extent the mid 1980s and early 1990s. The mid 1990s are marked by a short-time wet spell. On the contrary the 1960s and early 1970s as well as late 1970s are characterized by prolonged drought phases. The late 1990s and most of the 2000s show drought phases on a 6-month time scale.

12-month scale

The SPEI shows 130 values less than -1 on a 12-month time scale. They range from eight to thirteen occurences in all months of the year with most droughty conditions in December and January and least in March. Only five severe droughts were detected on that time scale. They cumulate in the dry season (January, February, April and October each one) with one additional drought in August.
There are altogether 101 moderate humid periods. The minimum value is four in October and June has a maximum of twelve occurences. A cumulation of wet spell events can be stated for the austral winter months. 15 occurences of pronounced humid periods can be observed. February, October and November each have two wet spells, all other months show only one.
The general trend of the SPEI value curve starts similar to the 3- and 6-month scale. The beginning of the data set shows a light drought in the late 1940s. The mid to late 1950s are a phase of above normal humid conditions. The early 1960s are moderately dry. The time from the mid 1960s to the early 1970s is affected by the absence of distinct humid periods and an alternating dominance of moderate and severe drought conditions. The mid 1970s are extraordinarily humid followed by a phase of drought in the end of that decade. The 1980s and early 1990s tend to above normal conditions, but an extraordinarily wet period can not be stated. The curve is fluctuating with some minor humid phases. After a relatively humid period in the end 1990s a ten year moderately drought can be seen. Around 2010 must be considered as a drought period.

24-month scale

The calculation of the SPEI on a 24-month scale brought a frequency of 145 moderate and no severe droughts on the 64-year time series. The moderate droughts on that time scale occur

Table 5.1: Absolute values of the frequencies of humid and drought periods using the Standardardized Precipitation-Evapotranspiration Index on various time scales.

	Scale	3-months		6-months		12-months		24-months	
		> 1	**> 2**	**> 1**	**> 2**	**> 1**	**> 2**	**> 1**	**> 2**
SPEI Humid Periods	January	9	2	7	2	9	1	7	2
	February	7	2	7	3	7	2	6	3
	March	7	2	9	2	8	1	9	2
	April	7	1	10	1	9	1	12	0
	May	11	1	11	1	11	1	11	0
	June	8	1	10	2	12	1	10	0
	July	5	2	4	4	9	1	13	0
	August	9	1	7	2	10	1	13	0
	September	0	0	6	2	8	1	9	1
	October	8	1	7	2	4	2	8	1
	November	11	1	10	1	7	2	6	1
	December	9	1	11	1	7	1	7	1
		< -1	**< -2**	**< -1**	**< -2**	**< -1**	**< -2**	**< -1**	**< -2**
SPEI Drought Periods	January	7	2	10	1	13	1	12	0
	February	10	1	11	1	11	1	13	0
	March	7	3	6	3	8	0	13	0
	April	9	1	9	1	10	1	12	0
	May	6	1	9	1	10	0	12	0
	June	10	1	0	0	11	0	12	0
	July	7	2	8	2	11	0	14	0
	August	10	1	9	1	11	1	12	0
	September	10	1	12	1	10	0	13	0
	October	0	0	0	0	10	1	11	0
	November	0	0	9	1	12	0	10	0
	December	9	2	8	1	13	0	11	0

throughout all months of the year with frequency ranges between ten (in November) and 14 (in July). No obvious cumulation of drought events can be stated.

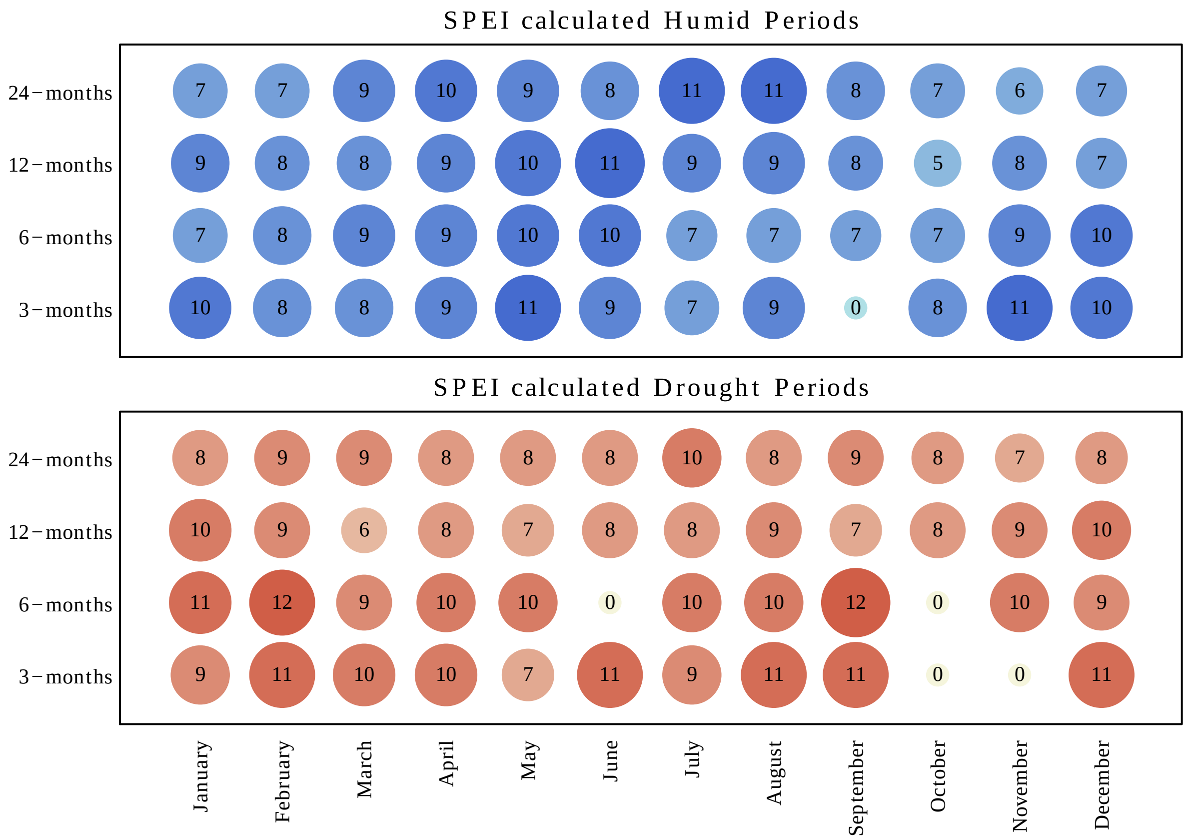


Figure 5.2: Comparison of the different time scales and its relative frequencies (%) of the Standardized Precipitation-Evapotranspiration Index. Values are rounded to integers. All absolute frequencies can be found in Table 5.1. Deviation from unity is due to rounding.

The wet spells show 111 minor and eleven major events. Values less than 2 range from six occurences (February and November) to thirteen (July and August). SPEI values bigger than 2 seem to cumulate in the dry season. The maximum occurs in February (3). The winter months do not show such values.

The SPEI values curve depicts varying humid conditions for the 1950s followed by a period of predominantly below normal conditions till the end 1960s before the period of drought prolongs until the mid 1970s. This phase ends in a humid phase in the mid and late 1970s. The 1980s start with a severe drought before the mainly slightly humid phase sets in till the mid 1990s. In the late 1990s a decennial droughty period starts succeeded by a short time of near to normal conditions. Around 2010 is marked by a moderate drought.

Dry & Wet Spells in the Course of the Year

A comparison of all four time scales is presented in Figure 5.2. Looking at the humid periods it can be seen that a 3-month time scale bears almost no alteration in the curse of a year. All indicated humid periods have a share of 7 % to 11 %. No humid period can be stated for September using SPEI on a 3-months time scale for this data set. The least share of 7 % can be stated for July. No obvious cumulation occurs.
The 6-months time scale shows not much more variability for the humid periods. Least values (7 %) can be stated for July till October. A relative concentration occurs in early rain and late dry season (March to June). All months have a share of humid phases between 7 % and 11 %.
The 12-months time scale shows the same trend as the above-mentioned. There is only a little variability throughout the year with a slight concentration in the austral winter months. All values range between 7 % and 11 % – except for October. It holds only 5 % of the detected wet spells on that time scale.
A 2-year scale reveals the similar pattern. A little less humid periods can be stated for the summer months whereas in the winter months the chance to experience a wet spell is much greater. The minimum is 6 % in November, the maximum of 11 % is reached in July and August. Remarkably there is some sort of double peak in the rain season. June is marked by only 8 % while months of the season show little higher values.
If the distribution of the humid periods is considered across the time scales the following occurs: September always occurs with minor shares (0 % – 8 %) and it is the only month with that shows the absence of a wet spell at one time scale. So there is only 6 % variability across all scales through the year. On scales lesser than twelve months the summer months tend to higher percentages than on larger time scales. The winter months mostly show the highest shares.
Drought periods on a 3-month time scale are not evenly distributed through the year. In October and November there was no dry spell detected. The remaining values range from 7 % to 11 % with the minimum in May. There is no tendency of cumulation.
The 6-month time scale ranges from 9 % to 12 %. In June and October no dry spells occur. The minimum occurence of drought is in December and March. The highest shares are in February and September.
The SPEI on larger time scales differs a little bit from the above-mentioned. The values appear more evenly distributed. The 12-month minimum is 6 % in March and the maximum is 10 % in January and December. All other values range from 9 % to 7 %. The 24-month SPEI has a minimum of 7 % in November and its maximum in July (10 %). The remaining months hold either 8 % or 9 %.
Across the time scales no obvious trends can be stated. On scales lesser than twelve months one might see a tendency towards extreme values in the second half of the year. Generally on that

scales the shares are higher since on that scales there are four months with no droughts at all. The temporal pattern of drought is characterized by a lack of variability.

5.1.2. Magnitude-Frequency-Analyses

The descriptive statistics of the three weather stations can be found in Table 5.2. Looking at Riviera the station covers minimum ten years ($a_{complete}$). In addition the total of measured days (d_{total}) is displayed since this station shows big gaps and to relate the days with rainfall (d_{rain}). The total of rainfall days indicates a strong seasonality of the precipitation with a rainy season setting in in April and lasting till August. The climax of the rainy season is reached in May and June with a share of almost 50 % of days of rain.
The values of the arithmetic mean do not differ that much from the other months. Considering a weighted mean (x_w) the gap between the humid and arid months increases. It also shows a maximum mean value of bigger than 30 mm in May and still values bigger than 20 mm in April and June. The standard deviation is for all months of the rain season increased with a maximum of 13.04 in May. The maximum value of rainfall of all three data sets is part of the Riviera time series. 126 mm were recorded on the 16th of May 2006.
The months of February, March and September appear to be the transition period taking into account the percentage of rainy days. Arithmetic and weighted mean differ from the rain season, but not from the dry season in the second half of the year. It lasts from October to at least January. January is the only month that stands out at first glance. It shows the lowest amounts of rain – either arithmetic or weighted mean – it has a very small standard deviation and the overall maximum amount of rain for that month is only at 11.18 mm.
The Kromvlei station covers at least eleven years without gaps in data. The seasonality is not that distinctive as the Riviera data, but a rain and a dry season with a transition phase can still be distinguished. Days with rain only represent around 20 % of the rain season in the Kromvlei data. The rain season (arithmetic) mean values are remarkably higher than for Riviera. The weighted mean also differs. It is higher for all months of the winter rain so as the maximum values are. Generally May has the most rainy days, rainfall amount and highest variability as in Riviera. The extreme amounts of rain from 16th of May 2006 can be found again in May with 110 mm as the second highest value ever in the three data sets.
January again is the month with the least rainfall, least variability and the lowest maximum amount of rain. During that period of the year the mean values (both) are also higher. In some cases the values even doubled like it is the case in November which means this month is extraordinarily humid in Kromvlei. This data set shows the same trend as it is in Riviera for the first three months of the year: After the very arid January some more rain falls in February and then March reveals more rainy days but less rain amount than January again.

Table 5.2: Monthly Analyses of Daily Rainfall Data from Riviera, Kromvlei and Piketberg.

Station		$a_{complete}$	d_{total}	d_{rain}	d_{rain} [%]	$\bar{x}$	$\bar{x}_w$	σ	maximum
Riviera	Jan	12	372	28	8	3.13	5.54	2.79	11.18
	Feb	11	324	55	17	5.37	15.66	7.56	31.24
	Mar	10	331	63	19	3.03	10.09	4.72	18.80
	Apr	11	330	132	40	5.81	21.30	9.56	40.20
	May	11	369	175	47	7.06	30.98	13.04	126.00
	Jun	12	368	171	46	6.14	24.10	10.53	54.20
	Jul	13	403	170	42	5.62	19.56	8.88	58.00
	Aug	12	393	138	35	5.09	14.21	6.83	31.50
	Sep	12	389	66	17	3.59	10.55	5.02	28.90
	Oct	12	388	27	7	4.39	13.19	6.27	40.80
	Nov	12	371	31	8	4.70	14.84	6.97	31.24
	Dec	11	356	32	9	4.92	16.50	7.68	29.70
Kromvlei	Jan	12	372	24	6	6.82	9.46	4.40	18.00
	Feb	11	311	28	9	8.56	16.85	8.69	32.00
	Mar	11	341	46	13	6.69	14.57	7.50	23.00
	Apr	11	330	59	18	12.63	26.96	13.63	50.00
	May	11	341	94	28	16.92	41.00	20.33	110.00
	Jun	11	330	64	19	19.14	37.47	18.85	80.00
	Jul	11	341	80	23	17.32	29.81	14.82	78.00
	Aug	11	341	72	21	13.73	24.00	11.94	53.00
	Sep	11	330	38	12	9.81	17.37	8.69	40.00
	Oct	11	341	16	5	7.51	14.18	7.16	42.00
	Nov	11	330	17	5	10.68	27.88	13.80	65.00
	Dec	11	341	14	4	8.00	16.54	8.45	36.00
Piketberg	Jan	7	217	21	10	4.41	12.28	6.13	22.61
	Feb	6	183	33	18	2.56	7.48	3.65	13.20
	Mar	6	186	39	21	2.47	4.09	2.07	6.10
	Apr	6	180	86	48	2.67	8.01	3.83	16.76
	May	6	186	110	59	3.28	10.93	5.04	28.19
	Jun	5	174	93	53	3.37	12.84	5.67	29.97
	Jul	5	168	113	67	3.12	13.54	5.73	31.10
	Aug	6	186	79	42	3.00	9.27	4.36	20.57
	Sep	6	180	37	21	2.09	6.68	3.11	15.24
	Oct	6	214	16	7	2.68	7.84	3.77	14.48
	Nov	6	191	18	9	2.72	9.48	4.36	21.34
	Dec	7	217	13	6	3.77	13.76	6.29	20.70

A third data set was analyzed from a station outside the valley beyond the mountain range of the Piketberg Mountains (see Figure 5.3). This data set covers minimum five mostly six complete years. It also reveals a distinct seasonality of the rainfall. Just like in the valley the months April to August show the highest share of days with rainfall. The other months have shares bigger 40 % and 50 % respectively. The maximum is in July with approximately two thirds of the month with some kind of rainfall event. The mean amount of rain is generally lower than for the other two stations and the variability is more or less equal throughout the year. Even the maxima are lower with the highest amount of rainfall of 31.10 mm in July.
The estimated transition phase between dry and rain season is also distinguishable but only due to the percentage of days with a rainfall event. The arithmetic mean values are generally low, so as the weighted mean is. Interestingly December and January show higher mean values (arithmetic and weighted) and higher standard deviations.
When the three stations are aligned they form a transect across the Piketberg Mountains from northwest to southeast. Along this transect the magnitude-frequency-analysis of rainfall was conducted (see Figure 5.3).
Subfigure A shows the station of Riviera. It takes statistically 4.03 days for an event of rainfall to come. But the magnitude of these events is rather low with 0.2 mm. Since the rainfall is not distributed all over the year equally and it occurs mainly in the time of the winter rain the RI values lesser than 365 must be seen as frequency per year. Thus it can be stated that rainfalls with 0.2 mm per day and a RI of 4.03 occur theoretically 90 times a year. The recurrence of rainfall with a magnitude of 10 mm is more than three weeks (24 days) or 15 times a year. A rainfall event with a recurrence interval of one year results in a downpour of 38.10 mm. All but one value fit into the linear model which is given in the Figure 5.3 A. That is the 16th of May event with its 126 mm of rain. It holds a recurrence interval of more than twelve years. Rang two holds 58 mm and is only a six-year-event. The estimation function for that station has positive slope of 19.38 using a logarithm to the base of ten. The intercept is -14.85.
The Kromvlei station is depicted in Subfigure B of Figure 5.3. Statistically it lasts more than a week (7.34 days) until any kind of precipitation is measured at the Kromvlei station. Remarkably that is the only value measured with 0.5 mm of rain, but since the neighboring rang has the value of one and it takes 7.35 days, it can be omitted. In other words an event of 1 mm can be expected almost 50 (49.66) times a year. A ten millimeter rainfall has a recurrence interval of 15.75 meaning that it takes statistically more than two weeks until a rainfall events reaches that amount. In theory such rainfalls can be expected 23 times in a year. Statistically a rainfall event of approximately 60 mm can be estimated once a year (368 days). The highest two rainfall values do not fit properly into the linear model. That is a 110 mm event with a RI of 11.02 years and a 99 mm event with a RI of 5.5 years. All other agree resonable. The generally higher

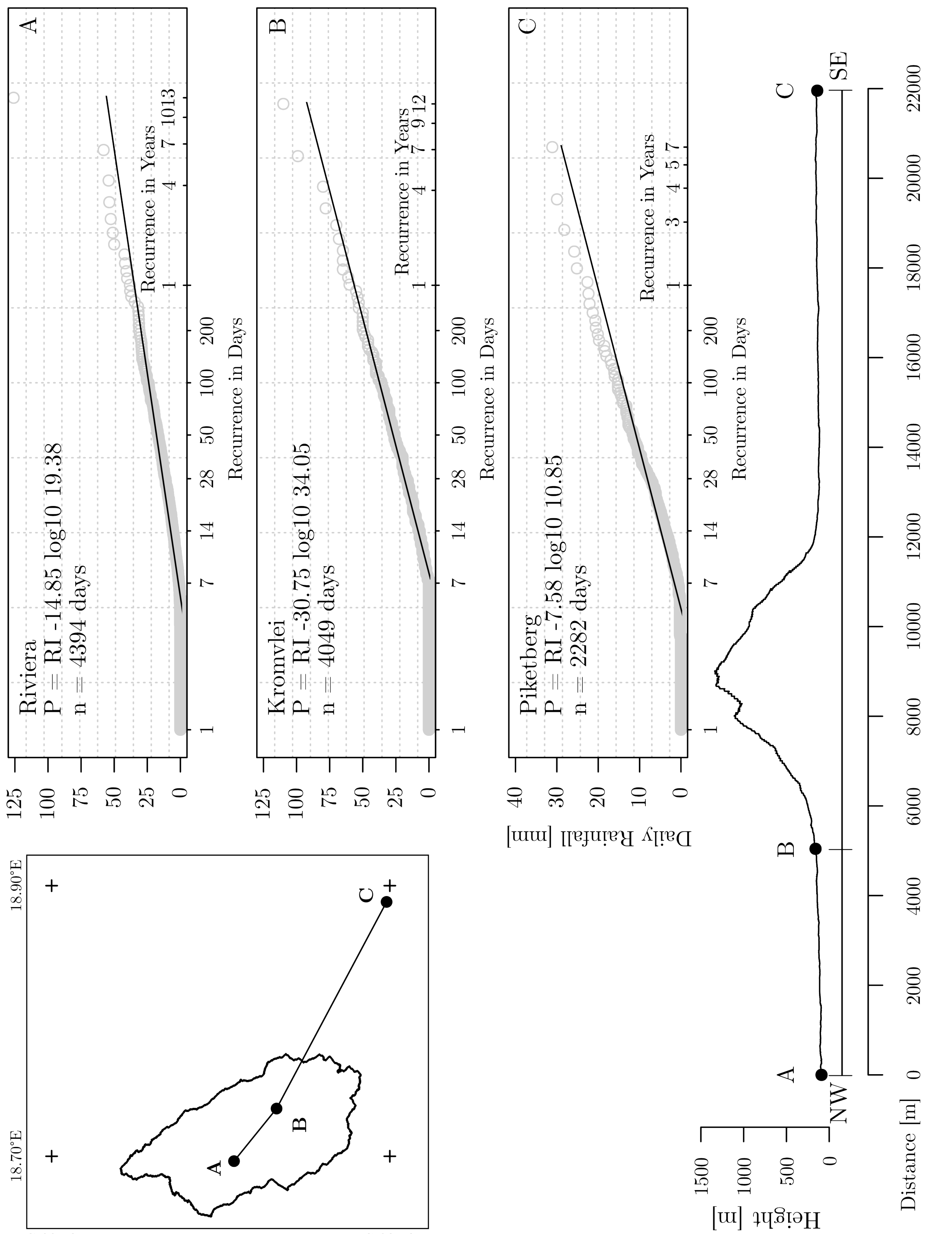


Figure 5.3: Magnitude-frequency-analyses of rainfall data for the weather stations of Riviera, Kromvlei and Piketberg. Kromvlei data is semiofficial. The transect crosses the Piketberg Mountains. Note the different scaling at A, B and C.

rainfall values of that station result in a high slope of 34.05 with a common logarithm and an intercept of −30.75.
The MFA of the weather station close to Piketberg is shown in Subfigure C. At that station a rainfall of 0.2 mm can be expected every 3.46 days and a ten millimeter rainfall recurs statistically every 39.36 days. That is 0.2 mm 105 times a year and ten millimeter nine times. A fair amount of 20 mm rainfall is likely to happen almost two times (1.92) a year with a RI of 190 days. The highest storm event at that station has an amount of 31.10 mm which represents a RI of more than six year (2283 days). The second highest one recurs every three years (1141 days) with an amount of not even 30 mm (29.97 mm) of rain. Piketberg shows the highest intercept (−7.58) and the moderate slope of 10.85 to the base of ten.
If one compares the three MFA from Figure 5.3 it can be stated that there is a gradient on the one hand in the valley of the Krom Antonies with more rainfall/larger quantities at Kromvlei than at Riviera, with a difference in the values of a specific rainfall event and a strongly differing estimation function of the linear model. On the other hand the two stations in the valley can be well distinguished from the Piketberg station. The rainfall amounts are smaller outside the valley, all the values agree properly (and somewhat better) to the model and the curve is less steep. A Wilcoxon rank sum test at 0.05 significance level indicated that all three data sets differ significantly.

5.1.3. Indices of Aridity

Table 5.3 shows the results of the calculation of I_m and I_e for the Riviera weather station. The Aridity Index after de Martonne indicates a seasonality. The months of September to April hold low and sometimes very low I_m values. Generally the first four months of the year show lower values than the last four months but since this is one contiguous period it is just a manifestation of a prolonged dry season. Remarkable is the circumstance that in between of the two months with the lowest values of I_m in January and March the February is in strong contrast. But nevertheless the whole eight-months period is classified as dry. The months of May to August show I_m values greater 10 and up to 15.62 as the maximum in June. The time of the winter rains is classified as semidry.
The Aridity Index after UNEP is less oversimplified. It shows a much finer distinction of the aridity in the course of the year. I_e values are lowest in January and March (0.02) and December (0.03). Again between the two lowest months there is a higher value in February (0.04). This time that causes the month of February to be classified as arid instead of hyperarid as the other three months. Aridly classified are also April and September to November. February is lowest followed by November (0.05). The maximum value for I_e of the arid months is 0.18 in September.

Table 5.3: Aridity indices after DE MARTONNE 1926 and UNEP (1992) for Riviera data. Climate classification according to MALIVA & MISSIMER 2012 and BALTAS 2007.

	de Martonne	Climate	UNEP	Climate
January	0.95	dry	0.02	hyperarid
February	1.8	dry	0.04	arid
March	0.91	dry	0.02	hyperarid
April	4.86	dry	0.16	arid
May	12.76	semidry	0.57	dry subhumid
June	15.62	semidry	0.81	humid
July	13.27	semidry	0.6	dry subhumid
August	12.44	semidry	0.48	semiarid
September	6.07	dry	0.18	arid
October	3.11	dry	0.07	arid
November	2.76	dry	0.05	arid
December	1.41	dry	0.03	hyperarid

From May to August the values are much higher than the rest of the year. August is the only month with semiarid climate (0.48). May and July have a dry subhumid climate with an I_e of 0.57 and 0.6 respectively. June holds the highest I_e of 0.81 which makes it uniquely classified as humid.
Both indices show a period of less severe aridity in the months of May to August. Furthermore they show a higher value for February with less distinct aridity in the arid season of the year.

5.2. Landuse & Landcover

The USGS landcover classes differ greatly in size. Table 5.4 shows that the Open Water class totals 35.8 ha which is 0.3 % of the 119 km^2 (11972.1 ha) catchment. This class contains water reservoirs exclusively and is to be found often in the lower course of streamless small valleys (see Figure 5.4).
In close proximity of open water the class herbaceous wetlands is observed which is like a transition zone to woody wetlands. Both wetland classes are mostly related to the Krom Antonies river and its tributaries. Woody wetlands are closer to the river beds while the other class is like a buffer around the woody zones. Where trees and shrubs are absent from the riverine vegetation the herbaceous vegetation class is dominant along the river course. Herbaceous wetlands and woody wetlands cover 50.6 ha and 162.9 ha of the catchment what equals 0.4 % and 1.4 % respectively.

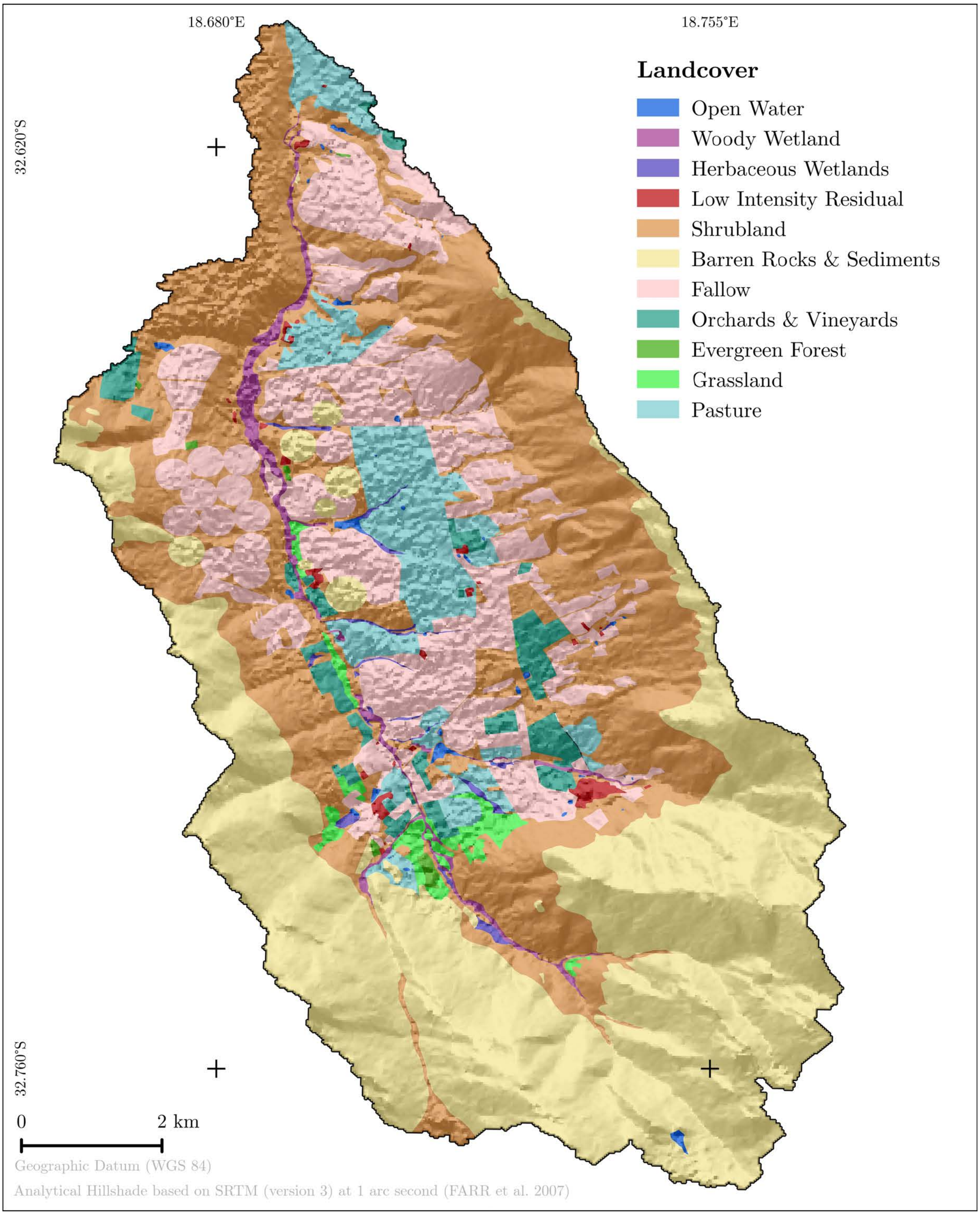


Figure 5.4: Map of landuse and landcover classification after USGS (2014).

Table 5.4: Landuse and landcover classes of the Krom Antonies catchment according to USGS (2014) and the corresponding share in the catchment.

	USGS Class	Area [ha]	Percentage
Open Water	11	35.8	0.3
Low Intensity Residential	21	54.2	0.5
Barren Rocks & Sediments	31	4439.4	37.1
Evergreen Forest	42	19.3	0.2
Shrubland	51	3822.7	31.9
Orchards & Vineyards	61	315.4	2.6
Grassland	71	141.0	1.2
Pasture	81	794.9	6.6
Fallow	84	2135.9	17.8
Woody Wetlands	91	162.9	1.4
Herbaceous Wetlands	92	50.6	0.4
		11972.1	100.0

Another minor class is low intensity residual which covers 54.2 ha (0.5 %). This class is located mostly near the river within an agglomeration of reservoirs. Remarkable examples are the settlements of Kromvlei, Moutonshoek and Karookop.

The smallest landcover class is evergreen forest with only 19.3 ha (0.2 %). This class is not really forested but more groups of trees. For most of this class the form of planting (in rows) and vicinage to settlements indicate that it is not natural vegetation but planted.

The biggest class is barren rocks & sediments. It forms a virtual border of the horseshoe-like catchment either by solid rock or loose sediment. Some of the agricultural fields within the catchment had to be classified as barren too. This class occupies with 4439.4 ha more than one third (37.1 %) of the valley.

The second biggest class is shrubland. It is located mostly on the foot slopes of the surrounding mountains or between agriculturally classified areas. A large portion of that class is found at the left bank of the Krom Antonies' lower course. Altogether this class covers 3822.7 ha what is equivalent to 31.9 %.

The class fallow comprises every surface which showed obvious agricultural traces like plowing or mulching but no active agricultural use at the time of mapping. This class is mostly found in the central part of the catchment on plane ground. Most of the pivots in the north-western part of the valley were classified as fallow. 17.8 % (2135.9 ha) of the valley were fallow.

6.6 % of the valley are mapped as pastures. Besides the extensive cow pastures mostly in the center part of the catchment there are some horse breeding ranches – namely Wilgerbosdrif

downstream and near Moutonshoek more upstream. The flood plains of the Krom Antonies were partly used as pasture too but were classified as woody wetlands thus the class pasture should actually cover more than the calculated 794.9 ha.
Another agriculturally used class is orchards & vineyards. this means either plantation of vine or citrus. They are located close to the river or near settlements and occupy 315.4 ha and 2.6 % of the valley. A relatively new plantation is found on the foot slopes in the north-western part of the valley. Hydrologically some parts of the plantation does not belong to the Krom Antonies river catchment.
The last group is grassland which is to be found near the river and/or near the settlement of Moutonshoek. It showed no signs of pasture and shrubs were absent too. With 141 ha (1.2 %) it is a relatively small class.

5.3. Soil Erosion & Deposition

5.3.1. Rainfall-Runoff Erosivity

The rainfall-runoff Erosivity (R factor) was estimated using daily rainfall data from Riviera weather station. The results of the R factor and the intermediate step of the MFI for each month are shown in Table 5.5. The course of the values starting in January to December implies

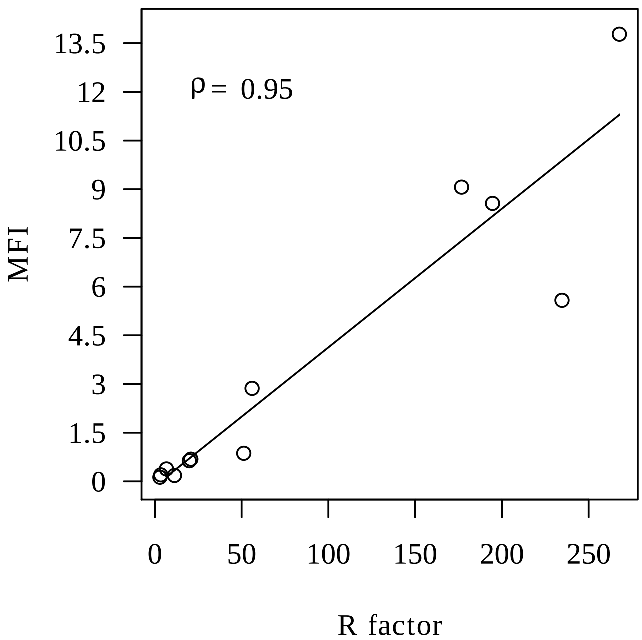


Figure 5.5: Correlation of monthly MFI and R factor using Spearman's ρ (Spearman 1904).

Table 5.5: Monthly MFI and R factor for Riviera Station.

	MFI	R factor
Jan	0.20	3.41
Feb	0.18	11.30
Mar	0.13	2.91
Apr	0.87	51.22
May	5.58	234.63
Jun	13.78	267.70
Jul	8.57	194.59
Aug	9.07	176.73
Sep	2.87	56.01
Oct	0.69	20.72
Nov	0.64	19.86
Dec	0.38	6.66

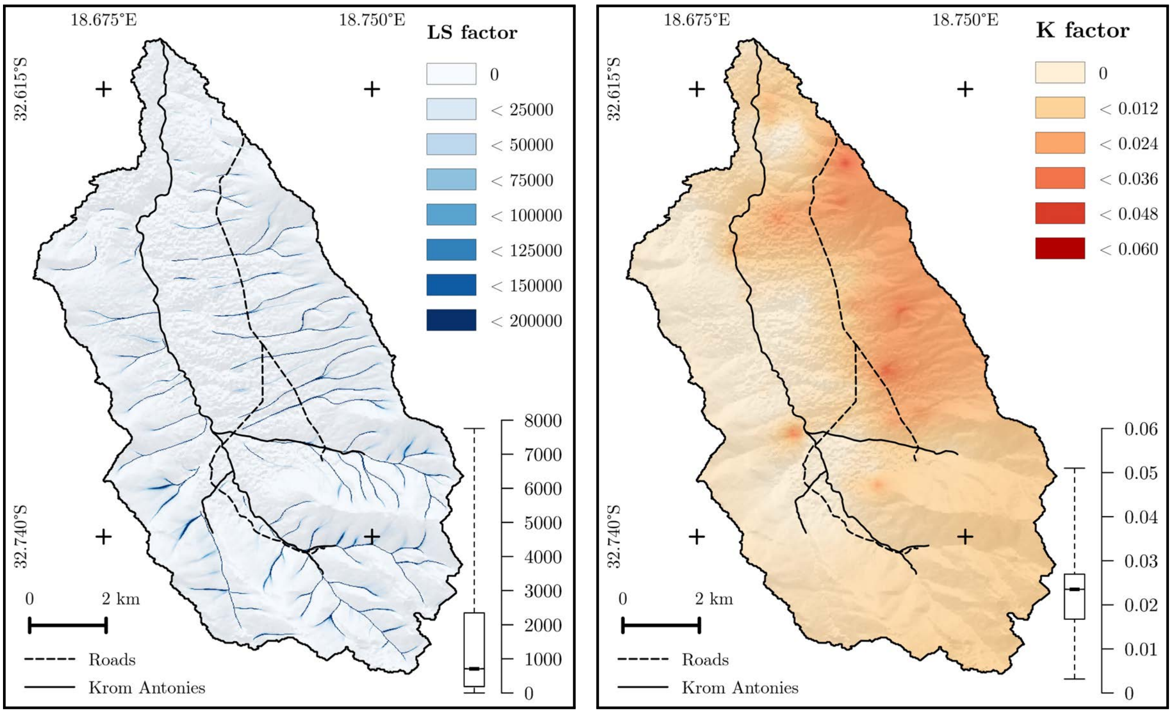
a seasonality with highest values in the months of May until August. June holds the maximum of 13.78. February and March share the lowest values with 0.18 and 0.13 respectively. The R factor follows the MFI trend. It also shows a strong seasonality with the maximum of 267.70 in June. May to August show generally increased R values greater than 100. The lowest value holds March with 2.91.
MFI ($W = 0.7737$, p-value $= 0.004776$) and R ($W = 0.7886$, p-value $= 0.006974$) are not normally distributed as a Shapiro-Wilk-Test (Shapiro & Wilk 1965) revealed. Thus the correlation between MFI and R was calculated with Spearman's ρ which indicates a strong correlation of 0.95.
It is remarkable that the low values agree very well with the linear model in Figure 5.5 but as the values get higher the deviation from the model seems to increase. This trend is not homogenously. While the month of July with a MFI of 8.57 and a R of 194.59 fits very well to the model the May values (MFI: 5.58 and R: 234.63) strongly deviates with a negative slope and the July with the maximum MFI of 13.78 and a R of 267.70 diverges from the model with a positive slope.

5.3.2. Slope Length & Steepness Factor

The map in Figure 5.6a shows the LS factor. It can be seen that slope length and steepness follow the direction of the drain. While in the headwater areas of the mountainous southern part the LS (if present) is spread dispersely all over the upper slopes it narrows and forms stream-like features downslopes. These channels hold exceptionally high values. On the footslopes this process stops rapidly. While the surrounding mountain ranges show extraordinarily high values of the LS factor the valley floor shows enhanced values only in very restricted areas. Near the outlet there are relatively less areas with increased LS while in the southern part these areas are concentrated. The eastern part shows higher concentrations of LS than the western part does. The distribution of the LS values reminds on χ^2 distributed values. This indicates that only a minority of the pixels bears an exceedingly high LS factor while the absolute majority holds reasonable values. The median is at 708 and 25 % of the raster cells do not exceed values greater than 2349. Nevertheless the mean value of 80254 ± 1889993 is really high.

5.3.3. Soil Erodibility Factor

The calculated K factor is shown on the map in Figure 5.6b. The values are exceedingly low since they were converted to metric units. The raster cells' mean values is 0.02303 t ha h/ha MJ mm with a standard deviation of 0.0077 t ha h/ha MJ mm. The minimum values of 0.003 t ha h/ha MJ mm are pretty minor. There are only very restricted regions close to the junction of the two major tributaries of the Krom Antonies near Moutonshoek that incorporate such



(a) LS factor **(b)** K factor

Figure 5.6: The resulting slope length and slope steepness (LS factor) with m = 1.6 and the soil erodibility (K factor).

values. The southern mountain ridges hold the majority of values lesser than 0.02 while the slopes west of Krom Antonies are mostly below 0.01 t ha h/ha MJ mm. The northern part downstream looks rather patchy. A small portion of the lowest K factor alternates with regions of relatively high values. Thus no general trend occurs in that region. The eastern slopes and mountain ranges are generally of higher K factor. As a result of the spline interpolation there are remaining relicts of points with values that differ strikingly from there surrounding samples. The course of the river has no obvious effect on the K factor.

Product of the Primary Particle Size Fractions

The grain-size distribution of the samples after USDA can be seen in Figure 5.7, the raw data is part of the addendum (Table B 1) and the spatial distribution of them is depicted on the overview map in Figure 5.8. The clay fraction is generally small but at least one sample (S004) shows a clay content of more than 18 %. The mean clay content is 2.7±2.4 %. It can be seen that those samples that are relatively rich in clay (bigger than 5 %) are mostly located near or directly on the footslopes. There are two samples (S004 and S005) that deviate from that pattern. They were collected on a citrus plantation relatively close to the river. Those samples with very low clay content (less than 1 %) are mostly in the central part of the valley but not directly in the meadow. The zone between the large more or less plain valley bottom and the surrounding slopes shows moderate clay content between 2 % and 5 %. Generally the western part of the valley shows a lesser amount of clay than the eastern part.

The silt fraction (less than 5 µm) indicates a large variability. The sample with the least silt content holds 1.7 % the most percent of silt is almost 50 %. The mean silt in the samples is 14.2 ± 11 % with a median of 10.8 %. The silt distribution in the valley is similar to the clay distribution. The slopes hold more silt (except for the western slopes) while the valley bottom does not. Samples collected directly in or very close to the meadow also hold a higher silt content. And again the samples from the citrus plantation have a higher content of fine grained material but they do not hold the maximum. S039 from a slope in the northern part of the valley holds the maximum of 49.2 % silt. Samples from the western part show a lesser silt concentration than the eastern part in general.

The fraction of very fine sand (less than 10 µm) shows less variability. It holds a minimum of 0.7 % and a maximum of 15.5 % silt. The mean very fine sand content is 7.7 ± 3.4 % with a median of 8 %. The spatial distribution of that grain-size fraction is less distinct. Still the slopes hold the majority that fraction but there are also locations on the valley bottom that hold high portions of it. The highest amount of the very fine sand fraction holds I018 (15.5 %) which was collected on a table grape plantation. The lowest concentration (0.7 %) of very fine sand was measured for S050 on a footslope.

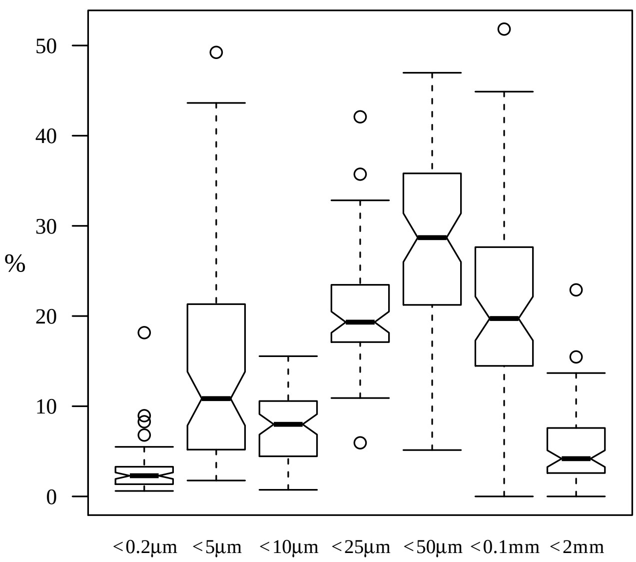


Figure 5.7: Variability of the samples. Grain-size classes after USDA (1987).

The samples have a mean content of the fine sand fraction (less than 25 µm) of 20.7 % with a standard deviation of 6.1 %. It is also sample S050 which holds the minimum content of that friction with 5.9 %. The maximum of 42.1 % has sample S009 which was sampled on the western side of the Krom Antonies in a heavily vegetated barranco-like valley. Generally there is no obvious spatial distribution pattern of that fraction. The samples that bear the most clay and silt (S024, S039 and I006) in the mid- and upper slope positions have lesser portions of that fraction. No obvious distinction can be made between the western and eastern side of the valley regarding to the fine sand fraction.

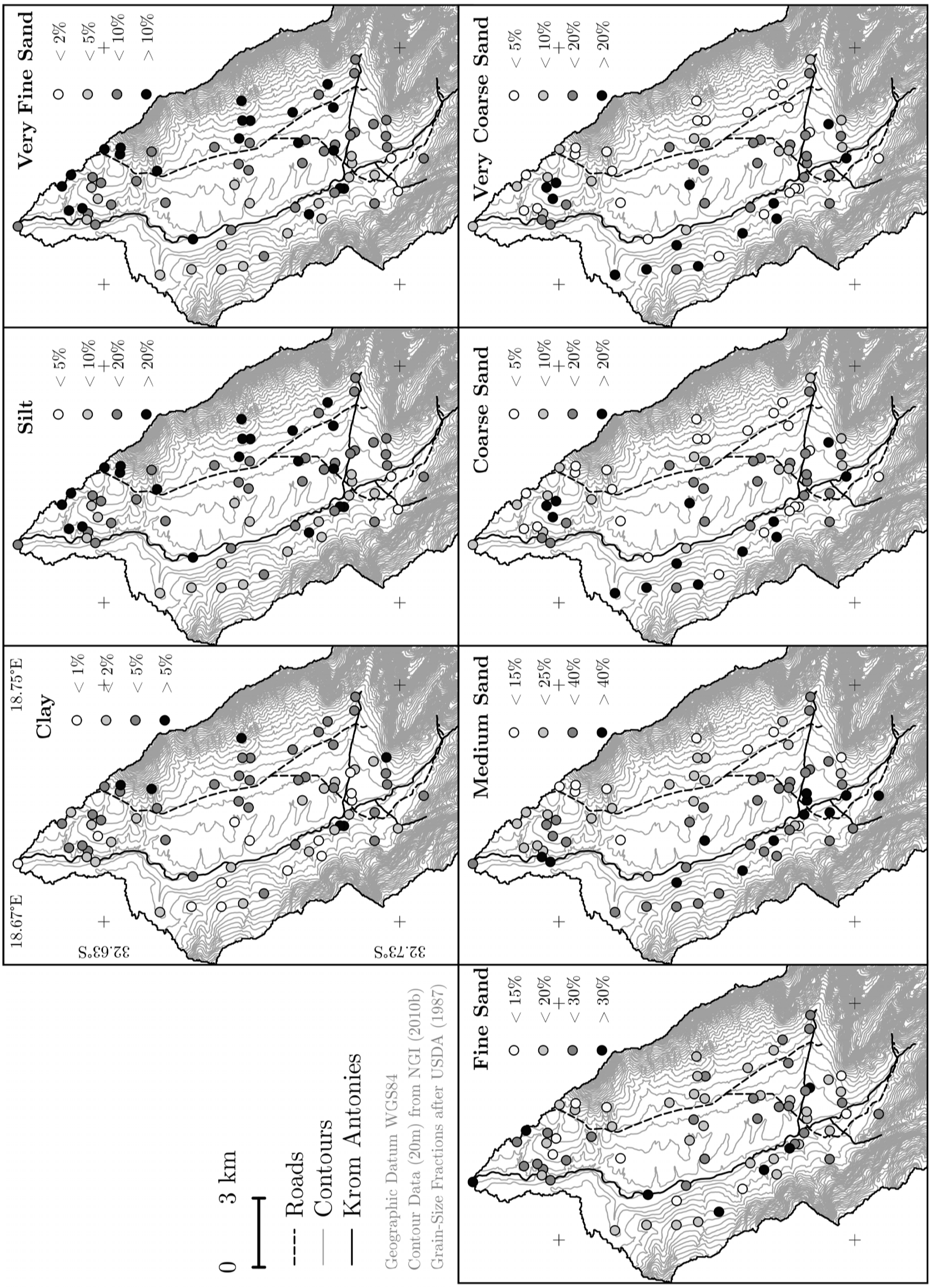


Figure 5.8: Distribution of the sediment samples and its grain-size classes after USDA (2014). Class breaks are rounded to quartile values.

Looking at the medium sand fraction (less than 50 µm) reveals that this is the major grain-size fraction in the catchment. The mean content is 28 ± 10.6 % and the median is 28.7 %. It shows a wide variability since the maximum of 46.9 % is in sample S011 that was collected in the vicinity of the junction of the two main tributaries of the Krom Antonies and the minimum is 5.1 % in sample S004. S004 also had the highest clay content. It clearly can be stated that the western part of the valley holds higher content of the medium sand grain-size fraction. Also the upper part of the valley and the region near the outlet are rich in medium sand. The eastern part holds also high share of that fraction but generally lower than the west. The samples collected on the mid and upper slopes show the least content.
A less clear picture can be drawn for the coarse sand fraction (less than 0.1 mm). It shows the highest variability with a range of 51.8 pp. The mean coarse sand is 21.3 ± 9.8 %. The median value is 19.7 %. In this grain-size fraction it is again S004 and S050 that share the minimum and maximum values. The minimum content is 0 % and the maximum is 51.8 %. This is the first fraction where samples exist that do not have any share of. The spatial distribution is less clearly as the medium sand. There are more high values in the western part than in the east but not all samples in the western part have high contents of the coarse sand fraction. The majority of the samples holds a content of 15 % to 30 % coarse sand. Interestingly the samples from the mid and upper slopes do not show the least concentration.
The share of very coarse sands (less 2 mm) is generally low in the samples. The mean is 5.4 ± 4.2 % with a median of 4.2 %. The maximum share of that fraction holds I014 from a shrubland sample in the eastern part near Kromvlei. The slopes do not show coherently high amounts of coarse sands. Some show high shares up to 15 %, others have less than 5 % of that fraction. Generally the northern part downstream has a higher share of coarse grained sands. It cannot be stated if the western or the eastern part of the valley holds more coarse content since there is no distinct spatial pattern.

Hydraulic Conductivity

The sediment samples' hydraulic conductivity is depicted in Figure 5.9. Most of the values are rather low with the majority of the values between 0.7 cm min^{-1} and 3.1 cm min^{-1}. It has a mean value of 3.2 ± 7.5 cm min^{-1} and a median of 1.8 cm min^{-1}. In Figure 5.9 values bigger than 8 cm min^{-1} are not considered while plotting but for statistics. Following BAHRENBERG et al. (1985) values till 59.2 are not an outlier. Thus 62.4 and 85.7 cm min^{-1} remained unaccounted for further analyses.

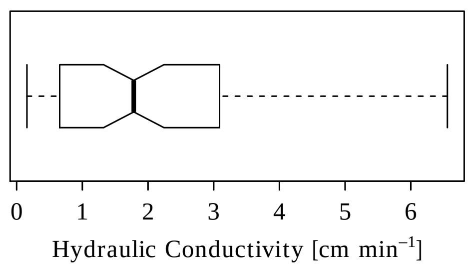


Figure 5.9: Boxplot of the hydraulic conductivity. Four outlying values are not covered by this plot.

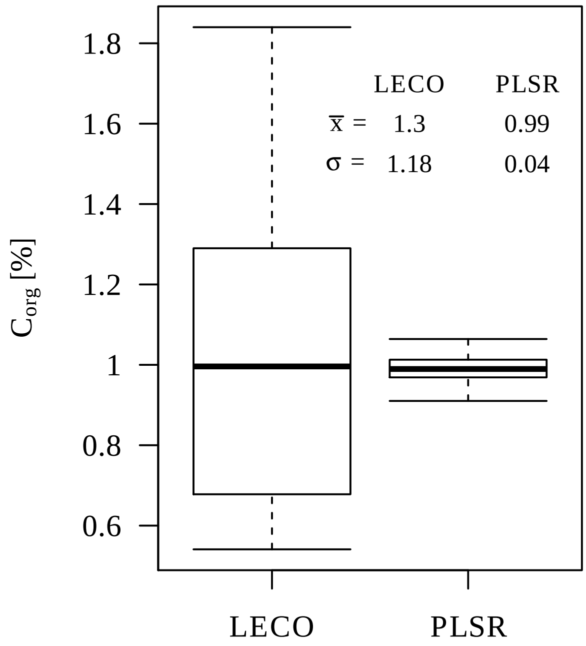


Figure 5.10: Comparison of the organic carbon estimated in laboratory via TruSpec CHN–S–Analyzer (LECO) and estimated using a regression model (PLSR).

Table 5.6: C_{org} estimated with TruSpec CHN–S–Analyzer in the laboratory and predicted with PLSR (testing data).

	LECO	Predicted Response
S007	1.040	0.988
S008	0.580	1.029
S009	1.150	1.004
S010	0.771	0.997
S011	0.802	0.991
S012	0.642	0.984
S013	1.240	1.048
S014	1.150	1.049
S020	1.290	1.004
S021	0.635	0.986
S022	1.130	0.978
S023	1.060	0.977
S024	2.250	0.925

Table 5.7: RMS error of the calibration and testing data set from the PTF.

	intercept	L	a*	b*
Calibration Data RMSEP	0.2663	0.2641	0.2636	0.2634
Testing Data RMSEP	0.4233	0.4281	0.4323	0.4357

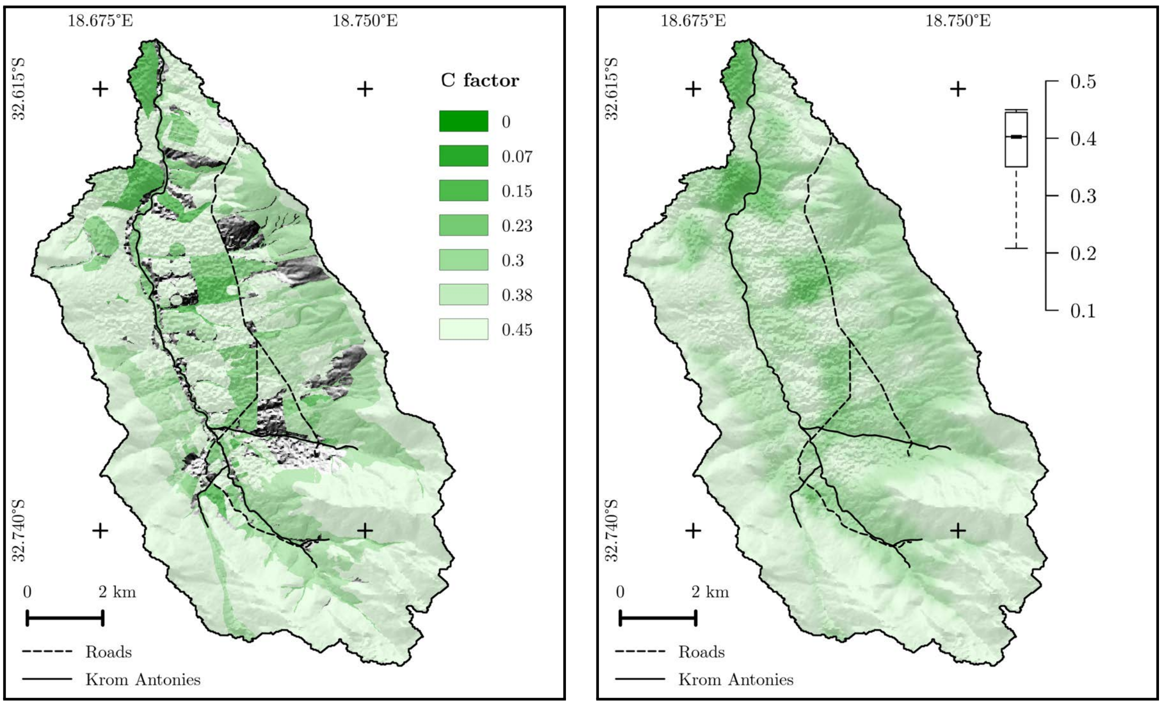
Organic Matter

The laboratory analysis brought an average content of total carbon of 1.3 % with a standard deviation of 1.2pp. The samples have a narrow range with a minimum of 0.5 % and a maximum of 6.8 % which must be considered an outlier after Bahrenberg et al. (1985). Thus altogether two samples were excluded. The second highest sample has an organic carbon content of 3.6 % which is a range of 3.1 pp.

Altogether 33 samples were analyzed. 21 of them were part of the calibration data set. 12 samples were used as a testing data set for the PLSR using the CIE L a* b* color space as derived from the top soil images as the predicting variable and C_{org} from the LECO as response variable.

The model has a very low performance. L has an explained variance of 36.8 %, a* has 40.2 % and b* has an explained variance of 23 %. Thus all components were taken into consideration for further calculation. Even different samples in the testing and calibration data sets brought no reasonable result and the cross-validated estimate of the subsets still differed strongly as the example in Table 5.6 shows. Table 5.7 shows the root mean squared error of the PLSR for the testing and calibration data sets. Some values agree very well such as S007 with 1.04 % measured with the LECO and 0.988 % estimated with the PLSR and S023 (LECO: 1.07 and PLSR: 0.97) with a deviation of less than 0.1 pp. On the other side there are samples that totally do not agree like S024 which holds a measured C_{org} which is more than the double of the predicted response.

Figure 5.10 shows the very diverging results of the two techniques. The plot does not consider values that exceed the one and a half times of the interquartile range for better presentability. While measured C_{org} is not normally distributed the predicted C_{org} is. It has a tiny standard deviation of 0.04 with a mean of almost 1 %. Both medians are identical (0.99 %).



(a) Mapped **(b)** Spline interpolation

Figure 5.11: The cover management factor (C factor) as a result of reclassification from the RUs attributes (Figure 5.11a) and the spatial interpolation for USPED (Figure 5.11b).

5.3.4. Cover Management Factor

The resulting map from the field mapping shows reasonable coverage of the catchment with RUs (Figure 5.11a). The surrounding mountain ranges show no appreciable canopy and mostly bare sediments or rocks thus they were classified to the highest C factor of 0.45. Much of the agriculturally used part of the valley is also classified to such high C values.
The transition zone from the mountain ranges to the valley bottom indicates slightly reduced C values since the slopes hold more shrubs and bushes. The meadow of the Krom Antonies is also vegetated – in some parts heavily vegetated – with shrubs and woody vegetation so that all of the river course holds C values around 0.3. Only a few units show a very low C of 0.17 which is the minimum raster value. They are restricted to the central and northern parts of the valley except one spot near the settlement of Moutonshoek at the upper course of the Krom Antonies at the foot slopes.
The values of the Cover factor range from 0.17 to 0.45 with a median of 0.4. The mean factor is 0.39. That indicates that most areas in the valley hold a significantly high factor and the distribution is very skewed.

5.3.5. Erosion & Deposition Rates

Figure 5.12 shows the resulting map of the calculation of the USPED model. Statistical indicators might be void in this place because the model tends to extreme values where the flow accumulation exceeds a specific threshold. Nevertheless the mean budget in the catchment is negative with a mean erosion of 0.02 t ha^{-1} yr^{-1}. The standard deviation of 44268.44 t ha^{-1} yr^{-1} is comparatively huge although 50 % (IQR) of the cell values range between -6.03 t ha^{-1} yr^{-1} and 2.48 t ha^{-1} yr^{-1}. The median is -0.24 t ha^{-1} yr^{-1}. The boxplot in Figure 5.12 shows no outliers with a value of 2.5 times the value of the IQR. The maximum value is 4733916 t ha^{-1} yr^{-1} and the minimum is -4733437 t ha^{-1} yr^{-1}.
It can be seen that the processes of erosion and deposition are highly correlated to the topograpy and the drainage system in particular. The mountainous region in the south shows most of the extreme severe erosional surfaces. Partly it is not only restricted to rill erosion and erosion along a drainage, in the headwater areas it forms extensive areas of denudation. Furthermore a little downstream and in parallel to the erosional surfaces there are also extensive areas of accumulation. The ridges itself hold very low values either of erosion or deposition.
If one compares the western and eastern slopes of the valley there is a distinction to be drawn. It is not only that the western slopes and barranco-like features are of less length they also show less intense erosion. In addition the eastern side is characterized by very closely connected little valleys that show extreme severe erosion rates all across the glacis and pediment. This is most intense in the upslope region of Ou Muur and Karookop and near the southern flanks of the

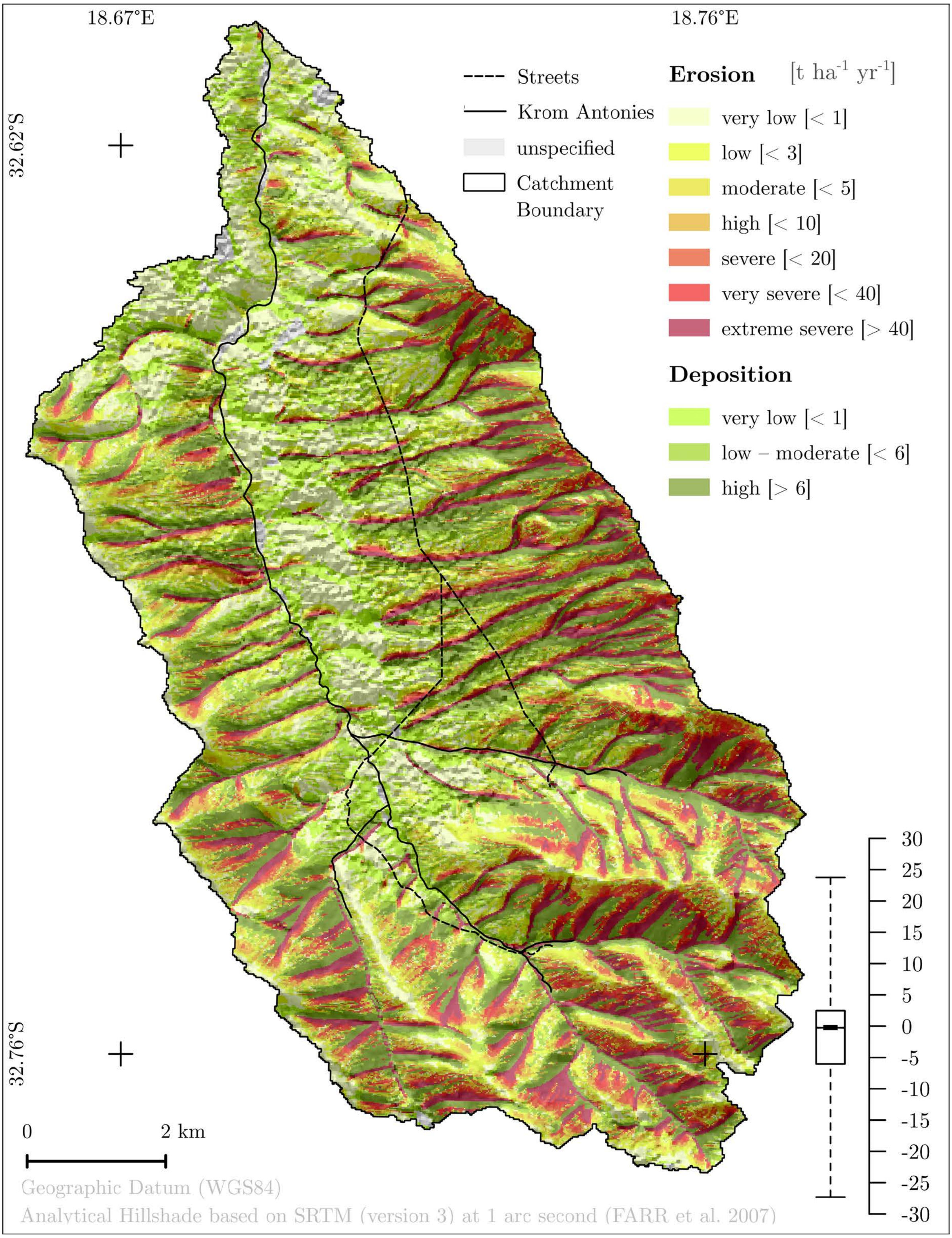


Figure 5.12: Map of the estimated erosion and deposition in the Krom Antonies catchment. The values were classified after VAN DER KNIJFF et al. 2002 and BORRELLI 2011.

Olifantskop. On the western slopes the linear erosion features end at the river bed of the Krom Antonies. The linear features from the east do not reach the river. They cross the main road of the valley without a significant reduction in intensity and taper off in the valley floor. The northern and central part is characterized by very little morphodynamic. Most of the areas that could not be specified are located in that part of the catchment.

Table 5.8 summarizes the size and portion of each class. It shows that solely 40 % of the catchment is determined by depositional processes. Almost 20 % of the surfaces show deposition rates bigger than 6 t ha^{-1} yr^{-1} which is the most prominent class with more than 2300 ha.

The second highest deposition class (bigger than 1 t ha^{-1} yr^{-1} and less than 6 t ha^{-1} yr^{-1}) still occupies more than 10 % of the valley and has a surface of 1244.5 ha. Surfaces with values less than 1 t ha^{-1} yr^{-1} deposition cover 1142.9 ha in size which is equivalent to 9.6 %.

Remarkably the erosion differs a little bit. Most cells are also in the highest class with a share of 17.7 % and a little more than 2100 ha. The second highest class has also the second highest rank with 10.8 % and almost 1300 ha but than the adjacent classes have all a more or less minor share of less than 7 % with various parts of the surface. But the class of less than 1 t ha^{-1} yr^{-1} has a bigger share again. This class covers approximately 10 % of the valley which equals to 1215 ha.

So together more than one third of the valley's surface suffers from either high or extremely high erosion rates of more than 10 t ha^{-1} yr^{-1}. Together with the deposition rates that makes more than the half of the valley that is influenced by high or very high morphodynamic. In contrast by summing up the two lowest classes it is only about 20 % which is morphologically relatively stable. Finally the unspecified areas which hold 0.8 % cover 92.6 what is the least class.

So in the end approximately 58.5 % of the valley are mainly characterized by erosive processes and 40.7 % are primarily deposition areas.

Table 5.8: Area and share of erosion and deposition classes.

	Area [ha]	**Share [%]**
Erosion		
extreme severe	2108.8	17.7
very severe	1292.0	10.8
severe	515.6	4.3
high	807.1	6.8
moderate	696.4	5.8
low	473.3	4.0
very low	1215.0	10.2
Deposition		
very low	1142.9	9.6
low – moderate	1244.5	10.4
high	2348.8	19.7
unspecified	92.6	0.8

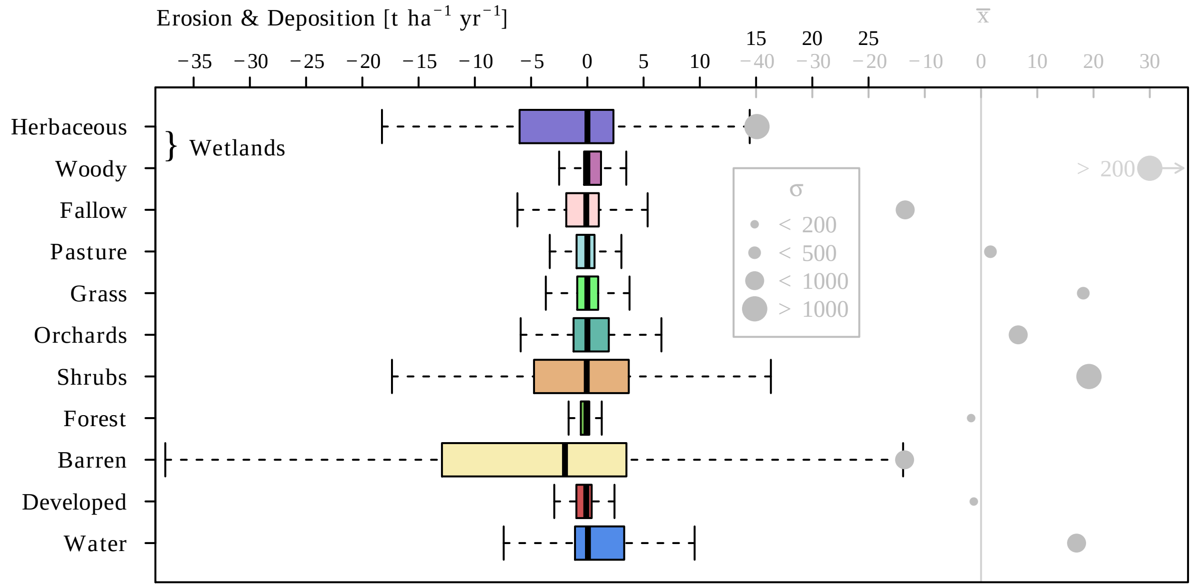


Figure 5.13: Landuse classes and the resulting USPED values. The boxplots do not cover values that exceed the IQR by one and a half times. Note that the mean value of the class woody wetlands does not fit the scale and for the plotting purpose the landuse class names differ. Absolute values can be found in Table B 3 in the addendum

5.3.6. Landuse, Erosion & Deposition

If one looks at the different landuse classes (see Section 5.2) and their properties according to erosion and deposition it is a mutable picture. The Figure 5.13 shows this relation statistically. The sole values can be found in Table 5.9. There are three to four classes that show a partly tremendous variability. These are the classes herbaceous wetlands, shrublands, barren rock & sediment and to a lesser extent open water.

Herbaceous wetlands show a high variability. It has a mean erosion rate of 1159.24 ± 5261.89 t ha^{-1} yr^{-1} and a mean deposition of 907.59 ± 4854.98 t ha^{-1} yr^{-1}. Both standard deviations are excessively high. This class is ranked as the second highest calculated. Shrublands are also highly variable but do not show extremely high values. They show a mean deposition of 325.20 ± 1392.93 t ha^{-1} yr^{-1} and erosion of 237.50 ± 1125.94 t ha^{-1} yr^{-1}.

Barren rocks & sediments and open water do not have such a very high standard deviation in comparison to the other highly variable classes. It is less than 1000 t ha^{-1} yr^{-1}. They show relatively high values for erosion of 138.32 t ha^{-1} yr^{-1} and 162.27 t ha^{-1} yr^{-1} respectively. There is on average more deposition on a barren surface (215.45 t ha^{-1} yr^{-1}) than on open water surfaces (155.2 t ha^{-1} yr^{-1}).

A reasonable variability is shown by the classes orchards & vineyards and fallow. The first shows high values for both. There is 111.82 ± 589.00 t ha^{-1} yr^{-1} deposition and 103.65 ±

Table 5.9: Mean erosion and deposition values including standard deviation for the different landuse and landcover classes.

	Deposition [t ha^{-1} yr^{-1}]	**Erosion [t ha^{-1} yr^{-1}]**
Herbaceous wetlands	907.59 ± 4854.98	1159.24 ± 5261.89
woody wetlands	2334.80 ± 12046.40	2022.47 ± 11570.74
Fallow	84.28 ± 489.95	83.65 ± 559.65
Pasture	47.20 ± 230.45	33.74 ± 171.51
Grassland	64.08 ± 327.64	24.04 ± 124.77
Orchards & Vineyards	111.82 ± 589.00	103.65 ± 554.38
Shrubland	325.20 ± 1392.93	237.50 ± 1125.94
Evergreen Forest	5.60 ± 19.76	7.06 ± 20.35
Barren Rocks & Sediments	215.45 ± 941.36	138.32 ± 759.76
Low Intensity Residential	12.49 ± 44.01	10.68 ± 39.22
Open Water	155.20 ± 746.55	162.27 ± 863.66

554.38 t ha^{-1} yr^{-1} erosion. Although it has a very large standard deviation it can be stated that the ratio of both processes seems to be stable with a tendency to deposition for the whole class. The same trend applies to fallow landcover but at lesser mean values. The mean erosion is 83.65 ± 559.65 t ha^{-1} yr^{-1} and the mean deposition reads 84.28 ± 489.95 t ha^{-1} yr^{-1}.
Woody wetlands have a moderate variability but the mean values for erosion and deposition are huge. These are the highest calculated values in the whole data set. Deposition has a mean rate of 2334.80 t ha^{-1} yr^{-1} with an excessively high standard deviation of 12046.40 t ha^{-1} yr^{-1}. It is almost the same for erosion. The mean is 2022.47 ± 11570.74 t ha^{-1} yr^{-1}. Deposition on average is lesser in the pasture class. There is a comparatively moderate erosion rate of 33.74 ± 171.51 t ha^{-1} yr^{-1}. Deposition is also relatively low with a mean of 47.20 t ha^{-1} yr^{-1} and a standard deviation of 230.45 t ha^{-1} yr^{-1}.
Grassland tends more to deposition with a mean of 64.08 ± 327.64 t ha^{-1} yr^{-1}. The mean erosion is much lesser with 24.04 t ha^{-1} yr^{-1}and a standard deviation of 124.77 t ha^{-1} yr^{-1}. In comparison to the following two classes it shows a moderate variability. The two classes low intensity residential and evergreen forest have a narrow range of values. The first has the second lowest value ever with a mean erosion of 10.68 ± 39.22 t ha^{-1} yr^{-1} and an average deposition rate of 12.49 ± 44.01 t ha^{-1} yr^{-1}. This is still relatively high but in relation to the other classes this is morphologically stable. The latter and in terms of area second smallest class holds a mean of 7.06 ± 20.35 t ha^{-1} yr^{-1} for erosion and 5.60 ± 19.76 t ha^{-1} yr^{-1} for deposition respectively. These are the lowest ever calculated mean values for a class with the used settings for USPED.

6. Discussion

6.1. Climatic Variability

6.1.1. Wet & Dry Spells from 1948 – 2012

Drought periods and oscillations in climate are not only known since direct measurement for South Africa. Ballard (1986) reports on a period of unusually severe drought and abnormally low rainfalls in eastern and southern Africa in the first three decades of the 19th century. Dean & Macdonald (1994) show a decline in stocking rates since the beginning of the last century as a result of loss of productivity. They cannot conclude if this is due to climate change, overexploitation, market forcings or a combination of all.

The 1950s wet spell period agrees with findings of Dai et al. (1998) who state that a shift of El Niño Southern Oscillation (ENSO) activity towards warm phase set in in the 1970s. Before that time ENSO-induced anomalies in ENSO-sensitive areas were less severe. It is likely that this normal and slightly above normal (meaning wetter) phase reflects the ENSO period reported by Richard et al. (2000) ranging from 1946 to 1969 with relatively wet "decaying ENSO" phases. There seems to be disagreement and steady progress in science about ENSO and southern African climate and rainfall in particular. Reason & Rouault (2002) found "ENSO-like" patterns in seasurface temperatures, but they showed no coherent relationship to the winter rainfall region of South Africa. Wet and dry phases shall be merely related to positive and negative phases of the Antarcic Oscillation (AAO) (Reason & Rouault 2005). In 1981 there was found no relation of the winter rain regions' wet and dry spells in South Africa to any oscillation (Tyson 1981), while Lindesay (1988) found negative correlation between rainfall and Southern Oscillation Index in the winter rain South Africa. The 1950s wet phase does also agree with a global dataset from van der Schrier et al. (2013) that indicates wetter conditions in that decade for Southern Africa.

The droughty period ranging from the mid 1960s until 1974 slightly deviates from the findings of Tyson (1984) but the general trend remains identical. He associates this periods with prolonged blocking action in the southwest of the Indian Ocean. That results in above normal rainfall in central South Africa (Lindesay & Jury 1991), but the western part remains dry. The mid 1970s constant wet spell agrees with results from Miron & Tyson (1984). This time was characterized by an eastward displacement of the Indian Ocean high so that the tropical easterlies gained more influence which resulted in a remarkable moist period (Tyson 1984). It only agrees partly with Dyer & Tyson (1977) who report on a wet phase until the early 1980s. The neighboring Eastern Cape experienced a phase of ongoing drought from the 1940s until the

early 1970s. This was superseded by an excessively wet decade in the 1970s (Jury & Levey 1993a; Jury & Levey 1993b).
The severe drought of 1981 to 1984 (Walker et al. 1987) is also covered by the SPEI, but to a lesser extent than in other parts of South Africa and it is not as long lasting as proposed by the authors. The 1980s and early 1990s relatively wet period strongly deviates from the rest (Mason & Jury 1997; Tyson et al. 1997; Milton & Dean 2000; Richard et al. 2001). The 1992 severe drought in the country just shows up as a period of relative dryness in a phase of above normal rainfalls in the research area. Also the above (and below) normal rainy seasons as stated by Reason et al. (2006) do not agree very well. The 1996 proposed wet season (Crimp & Mason 1999) is covered reasonably – it occurs more like the transition from drought to above normal rainfall –, but the 2001 one can be determined using SPEI on the four used time scales. Additionally the 1998 dry spell after Reason et al. (2006) can not be found on all time scales sufficiently. A negative SPEI on longer time scales around 1995 could be associated with a relative drought and additional decline in wheat production in whole southern Africa (Leichenko & O'Brien 2002).
An anomalously wet year 2000 with devastating floods in eastern South Africa (Kane 2009) might result in a relatively moist period in the drought phase from 1998 to 2001. If at all these rainfalls can be seen on the shorter time scales in Figure 5.1 – which is unlikely. But it is possible that the heavy rainfalls in southwestern South Africa in August 2002 and in March 2003 (Singleton & Reason 2007) influence the SPEI across all time scales. The 3-month time scale shows peak values that indicate wet conditions in 2002 as well as the 6-month SPEI does. The 12- and 24-month SPEI shows also relatively wet conditions in a prolonged dry period. This might be also influenced by the austral winter 2003 as described by Kadomura (2005). While summer rain areas suffered from severe drought the winter rainfall areas in Western Cape experienced a very dry July followed by an abnormally wet August including heavy rainfalls, storms and serious flooding in wide areas. In contrast it is reported of a drought in Western Cape in 2003 (WMO 2006)[1] which is not covered by the SPEI at all. The relatively wet period from 2007 to 2009 might be the result of increased precipitation in combination with severe flooding of the West Coast municipalities in two successive years (Holloway et al. 2010). The drought from 2009 to 2011 is covered by the SPEI on various time scales, but it does not appear as severe as reported by Holloway et al. (2012).
Generally the SPEI values do not fit into a proposed dry period from the late 1970s to recent times across the southern hemisphere as stated by Verdon-Kidd & Kiem (2014). On contrast on larger time scales it shows some ten years long phases of predominant conditions. Very roughly speaking the 1950s and most of the 1980s were wet and mid 1960s to mid 1970s and mid 1990s to mid 2000s were dry. This pattern has no agreement in any publication. Kane (2009)

[1]See Figure A 1

associates some biennial and triennial oscillations with ENSO. Reason & Rouault (2002) showed a decadal variability of South African rainfall and its connection to ENSO, but this is not explanatory for the unsteady pattern in the late 1970s and early 1980s. Mason & Tyson (1992) showed that the southwestern Cape Province is merely affected by seasurface temperatures in the South Atlantic Ocean and its anomalies during rainfall season. A possible explanation is a sunspot cycle related periodicy. Sunspot cycles largely affect southern hemisphere westerlies with a quasi-decadal variability (Thresher 2002). Vines (1980) found a 10 – 12 year periodicity in rainfall in South Africa (besides 16 – 20 and 6 – 7 year cycles) which might be explained by sunspots as well.
So finally it can be stated that all types of droughts are inherent part of the climate in the Krom Antonies valley, whereas the exact causes are manifold. At least the climatological and to a lesser extent the hydrological causes are immutable and are a question of good management practice.

6.1.2. Seasonality, Frequency & Variability

The Western Cape province is affected by highly variable rainfall. Predominantly this is winter rainfall, meaning most time of the year is distinguished by scarcity of rain. The Cape largely differs from the rest of South Africa in that point. The Aridity Index after de Martonne (1926) (I_m) and the Aridity Index after UNEP 1992 (I_e) depict that clearly. I_m only shows the existence of a winter rain climate. I_e reveals the pattern of rainfalls.
According to the I_e most of the rainfalls occur from May to July. During that time the aquifers and storage dams need to be filled up for the remaining nine months of the year. Mzezewa et al. (2010) show that in South African summer rainfall area it is likely to receive less than 50 % of the mean amount of rainfall. Goodess & Jones (2002) present similar results for Iberian winter rainfall areas. That means these rains are likely to fail or to overshoot the mark with weather extremes occuring in the time of the rainy season as stated by Holloway et al. (2010). Additionally that implies if the farmers in the Krom Antonies valley would rely on rainfed agriculture solely the growing period would be shorten significantly – mostly because of the temperature induced excessive evaporation. The dry season is dominantly classified as arid and hyperarid. That reduces the length of the growing period to less than 16 % of the given time what is not sufficient for agriculture. Natural potential is barely enough for livestock farming after Bot et al. (2000) since hyperarid areas have no agricultural potential (Koohafkan & Stewart 2008).
The interannual variability of droughts on different time scales (Figure 5.2) showed no conclusive pattern. Maybe the approach of SPEI is not adequate – it is the deviation of standardized variable from the median – to determine seasonal patterns or some other kind of frequencies,

but Wu et al. (2007) showed the usage of standardized indices in arid environments and when working with seasonal rainfalls and concluded that when used with caution and calibration such indices are applicable.
There is a slight cumulation of wet periods across the time scales in winter months, but it takes no wonder that there are relatively more wet spells because seasonal rainfalls occur in winter. On 3-month scale there are up to 10 % of wet spells in the dry season months. It is most likely that this is affected by the nature of the SPEI. Even light rainfalls within a dry season are indicated by SPEI as a positive deviation from mean. This results in partly very erroneous values when using SPEI's predecessor SPI – especially in mountainous areas (Vicente-Serrano & López-Moreno 2005). The only month with no wet spell on a 3-month time scale is September. This can be explained by the preceding rain season and the declining rainfall amounts, so that coming from a high niveau in a month with relatively low niveau SPEI can not detect a wet spell. The relatively low frequencies of wet spells in the dry season can be explained similarily and increased frequencies are interpreted to be the result of interannual variability of the onset of the rain season.
The drought frequencies are similar hard to interpret as the wet spells. They are highly variable and more or less evenly distributed throughout the year. The frequecies of 11 % in August and September show impressively that shortcomings of water are possible and likely in the rain season. The peak value of 12 % in September on 6-month scale is interpreted to origin from a failing or below normal rain season. 0 % in June shows that in May and June the rainfalls set in and that the median does not fall below the preceding 6 months. On longer time scales no clear peaks occur. The very slightly increased frequencies might be a result of seasonality, but it is a sheer speculative statement.
The table of number of rainy days and mean rainfall amounts (Table 5.2) indicates relatively homogeneous mean values of rainfall throughout the year for each station, but the number of rainy days differ significantly. On the one hand in winter the westerlies gain more influence on the atmospherical circulation above Western Cape. On the other hand the similar mean values of rainfall are affected by tremendous evapotranspiration rates during dry season. Gash et al. (1991) showed the vast evaporation values of a Sahelian fallow shortly after the onset of dry season which were close to potential evaporation. The values were still high after several weeks of dry season. Comparable ecosystems in Australia show the remarkable seasonality of transpiration values of forest-like savanna which is up to 90 % of annual transpiration in the dry season (O'Grady et al. 1999).
The table shows also an inherent difference between the three stations in use. Riviera and Kromvlei are relatively close (approximately 5 km) but the number of rainy days differs significantly. Multiple factors are possible explanations: Kromvlei stores in integer while official Riviera station also covers floating numbers so that also very little rainfall amounts are measured

as well. That means that a relatively big part (varying between 25 and 60 %) of the rainfalls in the course of the year occurs as drizzling rain or as nebular moisture – most likely in the morning hours – that somehow precipitates on the measuring instrument at Riviera. Additionally dryland rains are usually temporal erratic but also spatially highly variable (e.g. Goodrich et al. 1995; Lázaro et al. 2001; Modarres & de Paulo Rodrigues da Silva 2007). Another explanation is that Kromvlei station is not an official ARC station so that the accuracy of the measurement is poor. This is not supported by the fact that Kromvlei is part the Karsten Group (Karsten Group 2015) – an agricultural company that does business in a dryland where all attention is paid to the scarce resource of water.

Nevertheless Riviera is akin to Kromvlei and both show a clear distinction to the Piketberg station. Considering the westerlies as main source of moisture in that region, the lack of moisture sources in the continental South Africa, the topography of the Great Escarpment and the related katabatic bergwinds (van Rooy 1936) and the elevation up to 1500 m a.s.l. of the Piketberg Mountains most likely the majority or at least much of the rainfalls in the Krom Antonies valley are associated with orographic rainfalls while Piketberg is in a topographic and climatic unfavorable position. Piketberg indeed shows a higher share of rainy days, but it also shows much lesser rainfall amounts (arithmetic and weighted). This station's values are also stored in floating point number, thus it is most likely that most of the rainfall events are rather small events.

That relation is also supported by the magnitude-frequency-analyses (Figure 5.3). Riviera and Kromvlei indicate similar recurrence intervals whereas Piketberg strongly deviates. The time series for that station is much shorter than for the two in the Krom Antonies valley, but according to Ahnert (1982) five years of data should be sufficient. Additionally the log-linear relationship of the rainfall events and the line of best fit agree very well. So Piketberg seems to belong to a different hydrologic and climatic "regime" respectively.

It is problematic that Kromvlei and Riviera show completely different intercepts and slope values. In the Kromvlei plot all rainfall events still follow the log-linear relationship. A small deviation like the two highest values are slightly overestimated by that function, but they still fit into a range of probability (Ahnert 1982). Bergkamp et al. (1999) show for different climates in Spain that even largely varying rainfall amounts still follow this relation. Both for Riviera the highest rainfall event strongly deviates. For Kromvlei and Riviera the peak values belong to the same event in May 2006. That event is not covered by the Piketberg station, because the period of measurement sets in in October 2007. The difference from Riviera to Kromvlei is 16 mm (126 mm and 110 mm), which is unlikely to be explained orographically since there is only a difference in altitude of about 20 m between the stations. Maybe it is related to operational failure in one or both of the stations – which is unlikely. Another cause could be the spatially highly variable rainfall as described by Seuffert (2002). This is supported by the second highest Kromvlei value of 98 mm what has no counterpart at Riviera.

6.1.3. Critique

Critique on the Datasets

The Riviera dataset is from official ARC station, nevertheless it lacks a lot of daily measurements and partly it does not cover several month. ARC did not deliver any data about (their) data handling and instrumentation, so that the data has been processed "as is". Some systematic errors must be considered:

- a systematic wind field deformation above the gauge orifice (own experience),
- most likely high evaporation loss in the collector and
- random error in the values.

Points one and two can not be verified, but obviously faulty values (like −15 °C) have been modified by hand or have been erased from the dataset (e.g. months with no valid entries). Additionally the used data sets are deviating from climatic normal (1961 – 1990).

The Global Historical Climatology Network version 2 (GHCN) (Peterson & Vose 1997) and the Climate Anomaly Monitoring System (CAMS) (Ropelewski et al. 1984) are compiled together to form the gauge-based precipitation data set NCEP (Chen et al. 2002). Xie et al. (1996) mention that the CAMS suffers from several limitations in sparsely populated areas. On the one hand such areas usually do not have a dense network of gauging stations and on the other hand the precipitation there is mostly highly variable. In this context the drawbacks of CAMS have already been mentioned in 1986 as insufficient homogeneity (partly no common base period) and critical spatial interpolation (Janowiak et al. 1986). Furthermore gridding tends to averaging data (Sun et al. 2006). An overview of possible errors during the creation of a global dataset from multiple sources is given by Lawrimore et al. (2011).

A general problem is depicted in Figure 6.1. The evapotranspiration and precipitation values for the overlapping approximately eleven years are plotted on another. The actual evapotranspiration (AET) rates were obtained too just to check on the agreement with the data from Riviera. The rainfall values agree reasonable, but AET rates deviate up to 95 %.

Deviations in the precipitation values might be explained with partly lacking data at Riviera. An almost four months lasting period of no measurements is indicated by the grey vertical bar. Other, shorter periods were not highlighted because they do not span more than a complete month. Presumably other periods of 100 % deviation between the datasets can be explained with the lack on one side. Another approach to explain it, is the averaging trend of gridded rainfall data (Sun et al. 2006). Spatially highly variable rainfalls are overestimated for single point measurements like Riviera. Additionally spatial interpolation techniques largely affect the

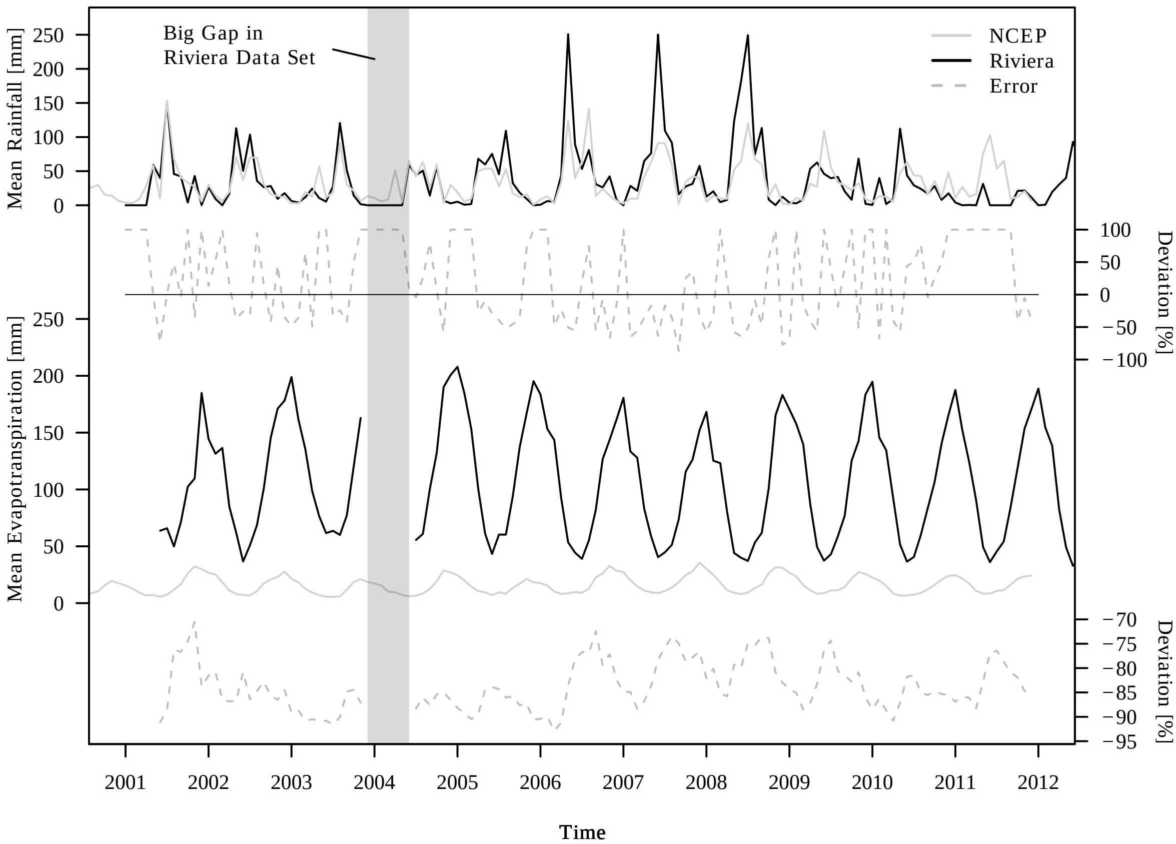


Figure 6.1: Comparison of evapotranspiration and precipitation from NCEP and Riviera station.

magnitude of extreme values – especially in regions where measurements are scarce (Haylock et al. 2008). Seuffert (2002) reports on the comparison of radar estimated and measured rainfall data and concludes that they are barely comparable. Additionally the topography must be considered when drawing conclusions about the deviations. The NCEP values are derived from a 0.5° × 0.5° grid. At the longitude of the study area that is almost 50 km × 50 km, which covers several mountain ranges of the Piketberg Mountains. So rainfalls of a totally differing climatic regime like stated for the climate station of Piketberg are possibly contributing to such a grid cell. That agrees fine with Austin (1987) who states that most deviations from the best fitting relation can be described in terms of physical factors.
To interpret the deviation between the evaporation data bears some problems. Since the NCEP data is not documented properly a little confusion occured if evaporation data is stated in inch. But NCEP agrees with the guidelines of World Meteorological Organization thus all units are SI units and evaporation is given in millimeter (WMO 2008). And even if NCEP values are converted with a factor of 2.54 (from inch to millimeter) the values are disputable. Some method inherent error is the averaging effect of the Piketberg Mountains on the gridded data

because PET and AET mostly decrease with elevation (Bean et al. 1994). Liuzzo et al. (2014) show the impact of wind speed and temperature on evapotranspiration in the Mediterranean. There is no final and sufficient explanation why the values from Riviera and NCEP deviate that much.
The potential evapotranspiration was derived from NCEP temperature data using the approach of Thornthwaite (1948). Since no sufficient data for radiation, soil moisture, vegetation cover, wind speed, etc. are obtainable for the respective time this simple approach was the only possible one. Thus the radiation – the more important part of calculating evapotranspiration – was neglected (Chang 1959). Nevertheless Thornthwaite's formula shows reasonable results, even if it tends to underestimation (Lu et al. 2005).

Critique on the Methods

Several studies showed that SPEI detects droughts sufficiently and that PET is a factor that can not be neglected when working in Mediterranean environments (Lorenzo-Lacruz et al. 2010; Lopez-Bustins et al. 2013). Wolf (2012) found SPEI preferable to other indices as a proxy for streamflow in the western United States. Paulo et al. (2012) found SPEI less reliable than PDSI whereas Vicente-Serrano et al. (2010a) found SPEI superior, because it detected more droughts than SPI and sc-PDSI.
Vicente-Serrano & López-Moreno (2005) found for a Mediterranean mountain basin a lack of correlation of hydrological responses and SPI values on scales greater than twelve months. Since SPEI is a standardized index too, this might be considered when working in mountainous areas with drought indices. Additionally New et al. (2006) showed that there are no consistent and statistically significant trends for small scale rainfall patterns in western and southern Africa which in return means that precipitation and potential evapotranspiration alone might be not completely sufficient in order to determine hydrological and meteorological droughts in that area. That agrees with Wu et al. (2007) who mention that the statistical product of input data is not necessarily related to the physical functioning of the Earth system.
The MFA is not really a standard procedure when working with rainfall data, but several studies showed its usefulness in various climate regimes (e.g. Ayoade 1976; Hershfield 1981). Figure 5.3 shows a linear function of rainfall and recurrence interval. Krause (2013) mentioned in his work that linear relations between rainfall amount and recurrence interval indicate BWh climates after Köppen (1923) in Northern Africa. Since the Riviera data indicate a Csb climate it can be stated that a linear function is valid for C climates after Köppen too.

6.2. Landuse & Landcover

The applied landcover classification was originally established for the derivation of remotely sensed data only. Hence the classes are rather coarse and they possibly do not agree totally with the specific environment because they were established to fit the needs of the continental United States (Homer et al. 2007). The color pattern differs from the suggested original (USGS 2014) since the majority of the classes would result in an overall yellowish-brownish occurence of the landuse map. A selection of exemplary photographs for each class can be found in the addendum in Figure A 2.

In the Krom Antonies valley the landcover classes related to water or dependent from the presence of water are underrepresented. Open water, woody wetlands and herbaceaous wetlands together occupy 2.1 % of the valley. Woody wetlands can be interpreted as natural gallery forest since this class is almost solely related to the course of the river and its meadows. Nevertheless along the stream lots of alien trees took up residence, what is of huge environmental relevance since their consumption of water is not adapted to drylands and they cause massive loss of water which means economic loss as well (e.g. Le Maitre et al. 2000; Le Maitre et al. 2002; Richardson & van Wilgen 2004). Most of the herbaceous wetland surfaces are of anthropogenic origin. There is a minority that is interpreted as natural – especially in the vicinity of the headwater areas – but the majority is enclosing the artificial ponds and reservoirs. Except for the Krom Antonies river and its seasonal and episodic tributaries there would be no waterbodies in the catchment.

Lakes and ponds are widely absent in that area and South Africa altogether shows only 0.55 % (DEAT 2006) or 0.38 % (Fairbanks et al. 2000) of them. Mostly they are beyond the Great Escarpment or at the eastern coast (DEAT 2006). The relatively high share of 0.53 % for Western Cape after Fairbanks et al. (2000) results from the vleis which have a huge impact on the zonal statistics. For the time series from 1986 to 2007 Stuckenberg (2012) showed in the neighboring Berg river catchment that waterbodies in that area are exclusively anthropogenic and their spatial variation is dramatically seasonal. This anthropogenic impact is not necessarily negative because small dams surely reduce river discharge rates but they have a positive impact on the water quality as well (Mantel et al. 2010). Moreover farm ponds function as a biodiversity hot spot on a small scale because of the waterbird assemblages (Froneman et al. 2001; Giliomee 2006).

Forests are not very common in the research area. They just hold 0.2 % and can be neglected. The natural vegetation is a forest-grassland mosaic which has nothing to do with the nowadays sporadically planted forests (Geldenhuys 1994). Most of the grassland class most likely can be neglected too. The planes have been interpreted like that because of the lacking traces of pasture like manure, hoofprints, watering places, etc.. Additionally they were partly located in

atypical locations – on the footslopes – and the ubiquitous fencing was missing. The plant cover was denser too and less patchy than in the pasture class. Thus it is interpreted as either formerly used as pasture where the natural vegetation has not yet succeeded or as very infrequently used pasture. Especially the grassland areas near the Krom Antonies or in its meadow are surely now and then used as pasture.
The location of as pastures classified areas is interpreted as the result of a maximizing effect. It is located in plane to gently sloping areas so it is accessible for the cattle. The distance to the stream is relatively long, because cattle need less water than agriculture. Vast areas of pasture are located downstream where generally less water is available either due to evapotranspiration losses or due to abstraction. These mapping results agree with the generally occurrences of pastures in areas unfavorable for agriculture (Snyman 1998; Snyman 1999). Very large areas of pasture are located outside the hydrological catchment but inside the boundaries of the Water User Association. The cumulation of pastures in the vicinity of Moutonshoek which is actually a favored place for agriculture results from the horse breeding and its related farmstead.
The low intensity residual class occupies only 0.5 % of the catchment. Most of it is either Kromvlei or Moutonshoek – each located close to one of the main tributaries of the Krom Antonies. All settlements are either located in the vicinity of the stream or on the footslope of the mountain chains. Thus their location is directly coupled to the all-season availability of water. On a larger scale Ferreira et al. (2007) indicate that water is one of the fundamental restrictions of settlements in the Western Cape.
The catchment is located on the edge of classical wine regions (Bargmann 2003) and in a classical potato farming area (Archer et al. 2009). Hence the 2.6 % of vineyards & orchards class – which only covers plantations of grapes and citrus – is considerably small but of highest economical importance. Conradie et al. (2009) showed that the shares of fruits and vegetables and wine in particular increased between 1952 and 2002 significantly in the district of Piketberg[2].
The high share (17.8 %) of fallow land is surely influenced by the time of the mapping. In January potatoes and several cereals are off season. So for instance various fallow pivots can be distinguished easily in the landcover map (Figure 5.4). This class can be found mainly to the east of the Krom Antonies in the center of the valley surely because of the greater distance from the footslopes to the stream and its low agricultural potential. Moreover presumably several RUs were misinterpreted and they are originally pasture – either infrequently used or remains of a former usage as pasture. On both sides in the middle course of the stream a lot of agriculturally used plains have been mapped as fallow. This region shows that there is also a smooth transition from fallow to barren surfaces. Some RUs – especially pivots – were so sparsely covered by litter or vegetation that it was imperative to classify them as barren (see Figure A 2a).

[2]Piketberg is one of 31 districts in Western Cape.

More than one third of the catchment (37.1 %) was classified as barren either rock or sediments. The horseshoe-like valley shape strongly influences the large proportion because the vast majority of that class follows from the bedrock of the surrounding mountains and its talus material. The barren agricultural plains are only of minor importance compared to the rest but they are considered to be temporally highly variable. Additionally it is difficult to distinguish between natural and artificial bare surfaces in that region (Stuckenberg 2012).
Only a little less than one third (31.9 %) was classified as shrubland. The ubiquitous seminatural vegetation is concentrated on the slopes, partly merges with fynbos and is interpreted as botanical succession in formerly or infrequently used agricultural plains in some areas or in between several landcover classes. This class also is likely to be misinterpreted because in some areas low shrubs could hardly be distinguished from some pasture that had grown out of control. Shrubs from the talus cones can be differentiated from the shrubs in the plains. On the cones it appeared more patchy and more diverse. In the plains the shrubs seemed less diverse, with as remains of agriculture interpreted plants[3] and a generally lower growth.
Altogether 86 % of the valley are either barren, fallow or covered with shrubs whereas approximately 3 % are used intensively for agricultural purposes. That means the water storage dams and the plantations together cover some 3.5 km^2 of a 119.7 km^2 catchment. This numbers game excludes extensive farming and neglects the practiced 3-years crop rotation (DAFF 2013), but it emphasizes the extraordinary favorable structure of the catchment. There is a large contributing area that favors a small agricultural heavily used part of the valley. Surely this statement looses weight if one considers

- the large amount of groundwater that is pumped from several aquifers,
- the crops that are grown there and their water consumption (Hoekstra & Hung 2002)
- and the tremendous loss of water that follows from the 35.8 ha of water storage dams and the excessive evaporation rates in austral summer (Jovanovic et al. 2011).

According to Hoekstra & Hung (2002) potatoes consume 4680 m^3/ha, grapes 9030 m^3/ha and water melons 5750 m^3/ha per growing period in South Africa. The water to irrigate these crops comes either from the limited rainfalls or from aquifers that are

- not completely understood (Conrad et al. 2005),
- already stressed and partly damaged (Cape Nature et al. 2006),
- will be stressed more against the background of increasing temperatures, decreasing rainfalls (Greene et al. 2012) and increasing rainfall anomalies (Archer et al. 2009) and
- managed in a rigid and mostly insufficient way (Seward et al. 2006).

To address this issues a new management plan is being established (WCDA 2012).

[3]One of such are water melons that are used as food for the cattle.

6.3. Soil Erosion & Deposition

6.3.1. Partial Results

Rainfall Erosivity (R)

There are plenty of ways to calculate the R factor (da Silva 2004), but in this study it was derived from the Modified Fournier Index (MFI) which is according to Ferro et al. (1999) presumably the best way if the excessive basis data is not accessible and when working in Mediterranean areas with distinct seasonal rains. That totally agrees with Smithen & Schulze (1982) who found best correlation for MFI to derive R in Western Cape. There is also the way to obtain rainfall erosivity from spaceborne radar data, what has not been tested in the study area, lacks calibration and validation data and shows partially poor results in other African study areas (Vrieling et al. 2010).
There are huge fluctuations in terms of daily precipitation amounts (Figure 6.2) throughout all seasons with highest peaks in the winter months. This variability is typical for a Mediterranean rainfall regime. That is supposed to be accelerated by climate change thus rainfall-runoff erosivity is likely to increase too (Munka et al. 2008; de Luis et al. 2010; Bonilla & Vidal 2011). Several restrictions arise from the input data. In all likelihood the annual trend of rainfall-runoff erosivity is correct, but the Riviera data has so much gaps that the values presumably are not 100 % correct. Additional the single point measurement in the valley neglects the

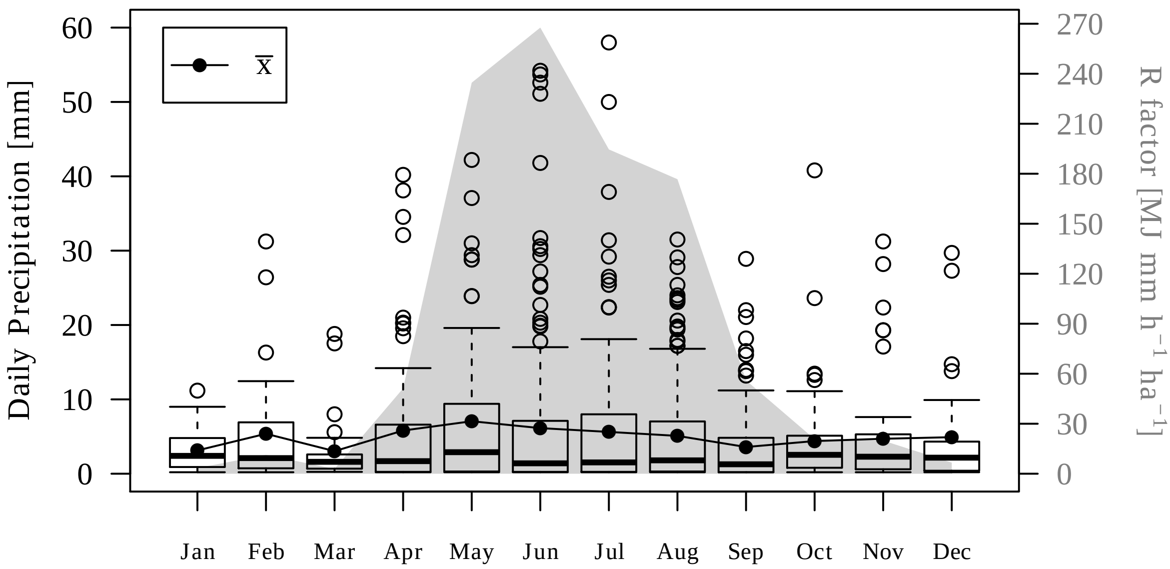


Figure 6.2: (R factor throughout the year for Riviera weather station. Note the points considered outliers exceed the limits of one and a half times the IQR.

topography and its influence on spatial pattern of rainfall intensity (Bacchi & Kottegoda 1995; Peñarrocha et al. 2002).
The mean annual value of 87 MJ mm h^{-1} ha^{-1} was used, despite the distinct seasonality of rainfalls which of course has a striking effect on the R factor (see Table 5.5). Following a classification of Oduro-Afriyie (1996) the four month of the rain season has extremely severe indices (greater than 100 MJ mm h^{-1} ha^{-1}), the two transition months April and September have moderate indices (40 MJ mm h^{-1} ha^{-1} – 60mm h^{-1} ha^{-1}), while the rest of the year is considered low (less than 40 MJ mm h^{-1} ha^{-1}). So the erosive rainfalls occur in winter only and it is likely that this pattern is levelled out by the scaling effect (Fraedrich & Larnder 1993).

Slope Length & Slope Steepness (LS)

The DEM was derived from contour lines using RST which is the most suitable technique for spatial interpolation since it has least RMS errors (Hofierka et al. 2002b). Nevertheless RST suffers from several limitations too, like

- segments and harsh segment breaks,
- overshooting,
- presence of non-realistic artificial surface features, and
- incorrect representation of artificial features (Cebecauer et al. 2002).

It is suggested to add supplementary geodata to avoid over- and undersampling (GRASS Development Team et al. 2014b), but in the case of the Krom Antonies valley no additional data was available. Hence the representation of the flat areas is considered to have a higher error.
Calculating LS is discussed in a complex way with various emphases. There are plenty of ways to calculate LS (e.g. Flacke et al. 1990; Zhang et al. 2013). Mostly the restricted approach of original USLE to estimate LS based on standard plot sizes is replaced by using the upslope contributing area (A) so that one-dimensional hillslope representation turns two-dimensional in a raster GIS (Fernandez et al. 2003). Using A has the advantage to incorporate the slope angle and terrain shape via derivates of the DEM as well as the water flow convergence and divergence (Mitasova et al. 1995). So not only rill and interrill erosion is accounted, but also ephemeral gully erosion when it results from flow convergence (van Rompaey et al. 2001). And generally several studies have proven that using the upslope area results in better (e.g. Hickey 2000; Garcia Rodriguez & Gimenez Suarez 2012) or at least comparable surface representation (van Remortel et al. 2001). The upslope area was calculated using a single flow direction (SFD) algorithm. Despite it lacks a little in precision the pathways – especially those of ephemeral gullies – have a better representation when using a SFD (Desmet & Govers

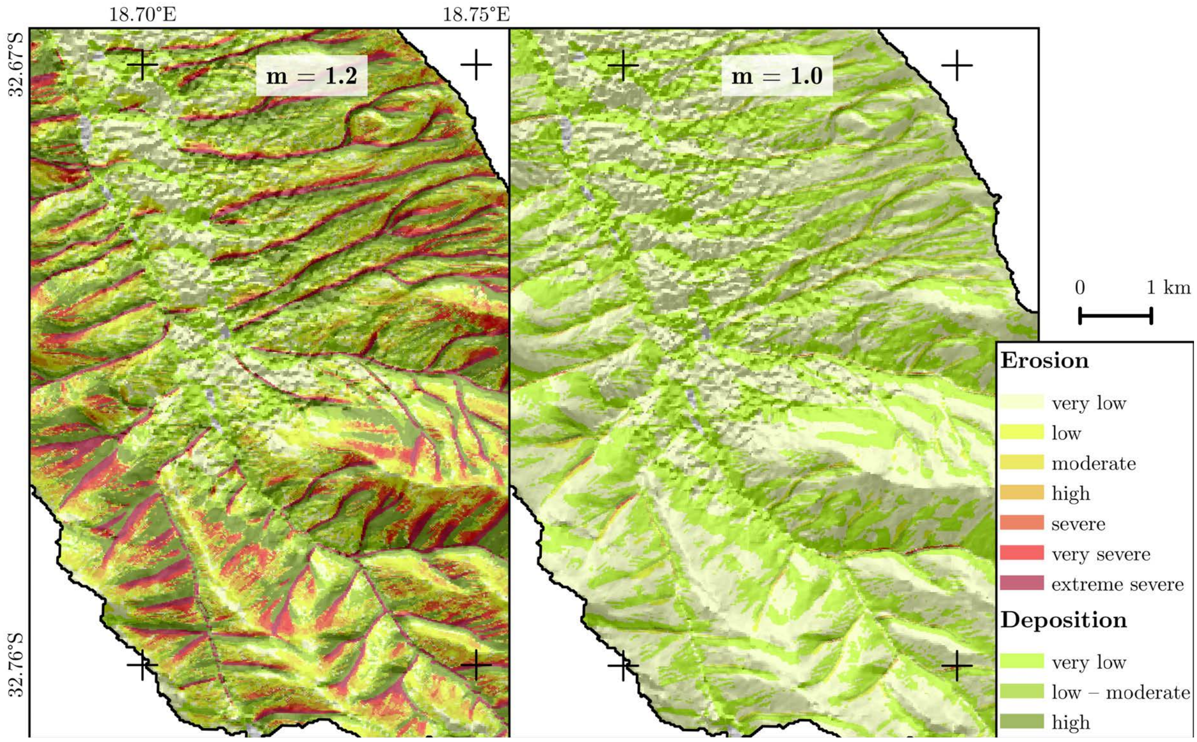


Figure 6.3: The left map shows erosion and deposition classes of USPED with LS factor m = 1.2 while the right segment of the map depicts the same model with m = 1.0 in the LS factor. In the study at hand the default m = 1.6 was used.

1996). Nonetheless the overall drawbacks of SFD are still valid and for hydrological purposes a multiple flow direction surely performs better.

The variables m and n determine the relation of rill to interrill erosion and they appear the most influencing factors when calculating LS. LIU et al. (2001) concluded that this relationship is likely to work on the base of theoretical approximation, but mostly it fails in complex topography. Per default m = 1.6 and n = 1.3 for prevailing rill erosion, but sometimes the outcome of cautious calibration is a range from 1 to 1.6 for m (MITASOVA & MITAS 2001). WARREN et al. (2004) showed the striking influence of spatial scale and slope computation on the accuracy when calculating erosion rates in southern Germany. Several authors avoid statements what factors they used for m and n (e.g. PISTOCCHI et al. 2002; SAAVEDRA & MANNAERTS 2005; LIU et al. 2007; DAMIAN et al. 2014) To point up the striking influence of m Figure 6.3 shows the results of USPED with different factors for m.

The distribution of high LS (Figure 5.6a) values in the headwater areas is coherent. The excessive LS values in such valleys and streams are way too high for later erosion and deposition estimation, but they indicate concentrated water flow (MITASOVA et al. 1999; MITASOVA et al. 2001) which is not unusual in drylands (ALEXANDROV et al. 2003).

Soil Erodibility Subfactors

Because the calculation of soil erodibility is such a data (and time) demanding procedure several factors, techniques and partial results must be considered:

Laser Diffractometry Restrictions arise from the laser diffractometry (LD) itself: According to Konert & Vandenberghe (1997) fractions greater than 63 µm tend to be overestimated because of the non-spherical shape of natural sand grains, while Loizeau et al. (1994) report nothing from the sand fractions but find an underestimation of the clay content. This agrees with the findings of Buurman et al. (2001) who discuss the platy structure of clays (less than 5 µm) as the main problem. Keck & Müller (2008) showed the improvements of LD to overcome these shortcomings, but also point out the necessity of validation and calibration instead of blind obedience to the output data. Despite the improvements LD has experienced over the last 20 years still doubts occur. Kelly & Etzler (2013) name five major pitfalls that can not be allayed completely.
One of such is the inability to validate the distribution calculations which can bypassed by processing raw data like in the study at hand. The other shortcomings of LD may be the case for the processed samples, but the clay content was interpreted from field survey as generally low to nonexistent. According to Keck & Müller (2008) only the submicron particle measurements are extremely sensitive to faulty instrumentation. That is a grain-size fraction mostly not met in the Krom Antonies.

Spatial Distribution of Grain-Size Fractions The overall distribution of the respective grain-size fractions goes along with a geomorphological interpretation. The generally low to very low clay fraction is found mainly on the (foot)slopes and to a lesser extent on agricultural fields. On the slopes this is interpreted as a weathering product, on the fields it might be seen as a sparse sign of pedogenesis. This interpretation is underlined by the fact that the highest clay content was found on a citrus plantation with a lot of time, some litter and good water availability. At first an increased organic carbon content was thought to be the cause, but the particular sample (S004) only shows 1.8 % according to the LECO.
The silt fraction is comparable big but it follows the same principles. It clearly can be seen in Figure 5.7 that the eastern slopes, the basin near the junction of the two major streams and the meadow show a higher share of silts. That might be due to the deposition in the alluvial sediments but it is also related to the geological setup. According to the geological map (Figure 2.2) the easterly mountains are some magnitudes older because they are of Malmesbury origin. In turn that means the rocks of the eastern mountain ranges in the Krom Antonies valley have been eroded for longer time what results in a higher content of fine grained material. This agrees with the findings in the region from several authors (Demlie et al. 2011; Eze &

Meadows 2014). That also explains the higher share of medium and coarse sand on the western side of the stream. It is younger, less weathered and not facing such intensive morphodynamics.

Hydraulic Conductivity The hydraulic conductivity was calculated using the Kozeny-Carman formula (Kozeny 1927; Carman 1937; Carman 1956) because the traditional approach after Hazen (1892) is limited to very homogeneous sands which does not cover the requirements of natural sediments (Carrier 2003).
Except for two outlying data points the values look reasonable. Martin & Moody (2001) found similar hourly rates in New Mexico and Colorado in post fire forests. Dunkerley (2002) found infiltration rates for Australian forest savanna that are heavily reduced in comparison to the ones used in this study. The high impermeability is related to the specific vegetation. Another strong influence on the infiltration rates is the stone cover either in topsoil or underground (Cerdà 2001). Brakensiek & Rawls (1994) showed that removal of rock fragments reduces infiltration rates by half.
Despite the large variability of values when applying the classification in order to obtain the permeability code after Römkens et al. (1997) all 76 samples fall into the same class. This was not the expected outcome but from field observation and interpretation it can be agreed that all samples have a rapid infiltration – even when they are highly variable. Especially at the slope where a higher content of silts were determined the effect of the silt was presumably foiled by the increased stone content and chasms in the parent rock material.

Organic Matter Most PTF heading for organic carbon (besides several minerals or nutrients) are not restricted to the visual spectrum of light. Viscarra Rossel et al. (2006b) give an excessive overview on spectral issues and the assessment of soil properties. Walvoort & McBratney (2001) used the spectral range from 250 nm – 2450 nm (UV-VNIR-SWIR) to detect carbon in soils. They also applied a PLSR using six factors and they report of the marvelous RMS error of 0.06, while others only got $R^2 = 0.76$ in the same spectrum (Islam et al. 2003). A Partial Least Square Regression with leaving-one-outlier-out performs better than comparable statistical methods after Muñoz & Kravchenko (2011).
Without explaining their technique in particular Daniel et al. (2003) report on $R^2 = 0.86$ when estimating organic matter with the V/NIR spectrum. Shepherd & Walsh (2002) show a successful PTF using portable spectrometer measurements from 400 nm to 2500 nm to estimate organic carbon with a range of R^2 from 0.55 to 0.81 for several African soils.
Some other studies used spaceborne data to derive information on soil organic carbon. Hill & Schütt (2000) successfully used Landsat 5 images to detect soil organic matter in Spain that was calibrated via field and laboratory spectrometry data in the range from 350 nm – 2500 nm. Gomez et al. (2008) compared field data with estimations drawn from Hyperion EO-1 data and

found strongly deviating results for Australian soils. They blame either the insufficient signal to noise ratio of the satellite or the spatial resolution of 30 m for their unsatisfactory results.

It can be stated that almost all studies dealing with kinds of carbon and/or organic matter used different spectra. Mostly they especially do not cover the visible range and focus on the NIR and SWIR (Viscarra Rossel et al. 2006b). The used photographs to derive soil color naturally roughly represent the range from 400 nm to 700 nm (VIS) what seems to be not sufficient to derive soil properties. There are various explanations.

Table 6.1: Selection of Electromagnetic Spectra after D'Amico et al. 2009.

λ [nm]	Abbr.		Name
< 400	UV		Ultraviolet
400 – 700	VIS	V/NIR	Visible
700 – 1400	NIR		Near Infrared
1400 – 3000	SWIR		Short wavelength Infrared

Viscarra Rossel et al. (2006a) compare soil color properties of dry and wet soils and conclude that results are less faulty when wetting the samples. That statement later found agreement when Viscarra Rossel et al. (2008) established a PTF for wet soils. Additionally they used standard laboratory conditions like diffuse light sources, steady lighting conditions, etc.. That is something the study at hand was definitely not able to apply. The lighting intensity could be levelled out by using the relative reflectance and the standard soil color chart, but the sunlight was direct at varying angles through the day and the soils were dust-dry. Furthermore the RGB-sampling could be erroneous due to remaining litter or mulch remains, but the pixel coordinates were chosen carefully by visual control and possible outliers in the sampling should be eliminated by multiple samples in the surroundings and averaging of these samples.

Additional explanations are interactions due to masking effects and spectral overlaps in V/NIR so that the establishment of a PTF is highly dependent on the specific sample and general assumptions get void in part (Vohland et al. 2011). So calibration is a necessity, but as Chang et al. (2001) show several calibration techniques can have massive effects on the predicted responses. Another possible approach to explain the failing relationship of soil organic carbon and surface reflectance is the colinearity problem reported by Summers et al. (2011). There are so much variables likely to influence a spectrometer that even under laboratory conditions using a high-resolution spectrometer the prediction of soil organic carbon was weakest among other variables. This could be also related to a shifting absorption feature from longer to shorter wavelengths as a result of soil typical impurities as reported by Ben-Dor et al. (1999).

Soil Erodibility (K)

For the KwaZulu Natal (eastern part of South Africa) a study of erosion factors (Lentsoane 2006) found surprisingly low K-values for either communal rangeland and dryland maize production. All the calculations were done plot based so it is a rather coarse estimation, but nonetheless the maximum value of 0.043 t ha h/ha MJ mm and the minimum of 0.016 t ha h/ha MJ mm is very low too,[4] especially if one considers the reverse soil permeability classes LENTSOANE (2006) states. HARTMANN et al. (1989) compared the nomograph technique with algebraic estimation of K for a time-series of several South African testing sites. They found values within a range of 0.0013 t ha h/ha MJ mm and 0.0754 t ha h/ha MJ mm.[5] BREETZKE et al. (2013) found only values lesser than 0.02 t ha h/ha MJ mm on various landuse classes. The occasionally occurring deviation of the algebraic equation and the nomograph is at least worrisome and must be taken into account when interpreting the results. PLATFORD (1982) also worked on testing sites with several rainfall simulators and also found discrepancy between the nomograph and the equation.

So there are several shortcomings in this study that likely influence the estimation of the soil erodibility factor. It does not necessarily mean the values are wrong but nonetheless it must be act with caution when working with the results because:

- The K factor is highly sensitive to the percent organic matter (LARSEN & MACDONALD 2007) – what presumably is not estimated sufficiently since the PTF somehow failed.
- TORRI et al. (1997) showed that organic matter and the clay fraction not necessarily yield a significant relationship to explain K using a regression. K was estimated on testing plots and thus could be verified easily.
- The hydraulic conductivity was derived from grain-sizes and not measured directly. It neglects knowledge about thickness of the sediment cover and does not comprise soil moisture and possible aggregates – even though aggregates are unlikely in the Sandveld.

Cover Management Factor (C)

The estimation of the cover management factor is a challenging issue. Actually it is a series of subfactors that is expressed as the deviation from a clean-tilled continuous-fallow. Ergo the higher the values the more protected is a surface by vegetation (YODER et al. 1997). In the study at hand it is simplified to a reclassification of mapped values of vegetation cover and vegetation type per RU to fit the needs of USLE. This was done via a decision tree that originally was established for permanent pasture, range and idle land, which assumes that vegetation and mulch are distributed randomly across the area (WISCHMEIER & SMITH 1978). That agrees well with the most time patchy vegetation in the research area and the dominant landuse. It would have been possible to estimate C values remotely. For example LU et al. (2001) successfull derived it from the Normalized Difference Vegetation Index (NDVI) using AVHRR satellite

[4] Originally it was processed in imperial units. Thus the maximum is 0.31 t ac h/hundreds ac ft tonf in and the minimum is 0.12 t ac h/hundreds ac ft tonf in.

[5] In imperial units 0.01 t ac h/hundreds ac ft tonf inand 0.55 t ac h/hundreds ac ft tonf in.

data. Schönbrodt et al. (2010) were successful with a similar technique using Landsat 5 data but suggest to apply to a macroscale only because landuse dynamics are likely to interfere the spectral signals on a microscale. So the mapping results from the RUs were preferred, despite it also suffers several weaknesses. USLE C values are derived from empirical data gathered between the 1930s and 1960s in the United States. Since then cultivation techniques and practices have changed so that the RUSLE procedure might fit better todays management practices. On the contrary USLE database was used to calibrate RUSLE C values (Gabriels et al. 2003).
The resulting C values are interpreted as generally low and the classification scheme is rather coarse. For example the prominent pivots on the western benches of the Krom Antonies largely differ according to vegetation type (from fallow and pioneer plants to cereals) and vegetation cover classes (100 % – 5 %), but nevertheless they all got the same C factor of 0.45. According to Wischmeier & Smith (1978) that is the maximum value which is at least disputable when 'no appreciable canopy' results in a C factor of 0.45. Breetzke et al. (2013) compared USLE and SLEMSA (Elwell 1978) subfactors for eastern South Africa and found distressing differences. The C_{USLE} holds a value of 0.008 for shrubland and thicket, while C_{SLEMSA} is 0.05. In this work a comparable RU holds 0.34. The same applies for managed grasslands: C_{USLE} is 0.008, C_{SLEMSA} is 0.06 and a similar RU holds a C value of 0.26 and 0.17 respectively. The most severe disagreement shows the class temporary commercial dryland.[6] C_{USLE} is 0.421 and C_{SLEMSA} is 0.06. In this study values of either 0.38 and 0.42 or 0.45 were classified. So the estimated values are closer to USLE values than to the region specific erosion model SLEMSA.
Le Roux et al. (2008) explicate the problem of the cover factor in Western Cape fynbos. They used a NDVI approach to obtain the C values and concluded that C and their corresponding erosion rates are grossly overestimated. Their conclusion charges the senescent fynbos to cause a NDVI decrease which induces a higher C value. In Eastern Cape values ranging from 0 to 0.87 were detected also by NDVI and for similar landuse (Mhangara et al. 2011). No information on the reliability of their derived C values was given by Smith et al. (2000), they just conclude that remote sensing data combined with field observation produces results. They do not report how these results look like.
No final conclusion about C values can be drawn. They seem too low and too high at once.

6.3.2. Soil Erosion & Deposition in the Krom Antonies Catchment

Soil Erosion & Deposition in various Landuse Classes

Open water which means the water storage dams in the Krom Antonies catchment are sediment sinks. Foster et al. (2007) show that these sinks are excellent and partly high resolution

[6] "TCD: Large, uniform, mechanised, well-managed field units under temporary crops with lack of major irrigation schemes" (Breetzke et al. 2013).

geoarchives in the central Karoo. From the field work it can be stated as well that the reservoirs show stratified layers which are likely to be caused either by sedimentation or by evaporation losses of stored water. Since Open Water is a rather small landuse class this finding is of almost no consequences. But nevertheless from this class serious threat can arise too. FOSTER et al. (2014) show the impact on the downstream (fluvial) connectivity and the potential impact from breaching dams when landscapes are altered in Eastern Cape. It was shown for Romanian dams that size correlates negatively with silting rates (RÃDOANE & RÃDOANE 2005). In southern Africa (Zimbabwe) CHIKATI (2007) found an increase of siltation rates the smaller the dams are. On that account it can be concluded that most reservoirs might have a rather high siltation rate since their small scale. On the contrary the erosive rainfall events are rare. Thus no clear trend is obvious. Further research is needed to clarify these relationship.
Both wetland classes fit properly to the geomorphological interpretation. The Herbaceous Wetlands are usually located further upslope so the transport capacity normally is higher there. That is reflected by a huge mean erosion rate of 39.84 t ha^{-1} yr^{-1} – the highest rate of all classes. Further down the sedimentation cascade there is the class of Woody Wetlands. The majority of that class is formed by the meadow of the Krom Antonies and its tributaries. It is a tremendous sediment sink. The mean deposition rate in that – spatially speaking – small class is greater than 200 t ha^{-1} yr^{-1}. But this value surely is trigged by some erroneous calculated cells since the variability is reasonable and is showing an altogether positive trend. TOOTH et al. (2009) show for the summer rainfall dominated east of South Africa that river courses can be remarkably stable in that area when certain circumstances are met. This is not likely in the Krom Antonies because there is no proposed (geological) barrier near the outlet and the sandstone – especially from the Malmesbury formations – is highly erodible (ALBHAISI 2012). The classes Low Intensity Residential and Evergreen Forest are either very small in spatial extent or it is barely possible to estimate reasonable values for the USLE subfactors in a heavily disturbed, non-agricultural and heterogenous area like residential. Nevertheless the very small mean erosion rate of 1.75 t ha^{-1} yr^{-1} for the Evergreen Forest landuse class seems to be realistic. SCOTT (2000) found generally low erosion rates under native forests in South Africa. The values for Low Intensity Residential can not be verified because there are no erosion models for residential environments. Additionally it does not make sense to apply agricultural factors on such surfaces. Thus no conclusions will be drawn from this class.
All the barren surfaces are eroding but this class shows a huge variability. Several problems occur:

- Barren rocks are not likely to be eroded by overland flowing water on a morphological short time scale (e.g. one year) and USPED (and (R)USLE) are designed to determine soil and sediment loss respectively. Thus the class name is deceptive.

- Barren rock's planes contribute to the upslope area of the barren sediments further downslope. So the flux of water and energy is higher downslopes because barely no infiltration occured upslope and protective vegetation is mostly absent or very sparse.
- Barren rocks & sediments is more a collective landuse class than a coherent and distinct surface attribute.

Therefore the variability is enormous ranging from steep slopes with vast quantities of rocks or rock fragments to nearly even agricultural fields that seemed abandoned. So the mean surface process is erosion but a high portion of that class is determined by sedimentation. It has to be mentioned that this is the only class where the median obviously deviates from close to zero (Figure 5.13) so erosion must be considered the primary process.

Vetter et al. (2006) report that shrubs can invade rangelands when grass biomass is reduced by grazing. In theory they benefit from locally increased water and nutrition availability respectively. Shrubs can cause further water and nutrition losses for other resources (Allsopp 1999), which in turn results in a further depletion of resources by loss of grass cover what perpetuates additional soil loss and bare patches. But most interpretations get void since Shrubland is a collective landuse class too.

Fallow areas are generally eroding surfaces in the study at hand (13.49 t ha^{-1} yr^{-1}). This is a little bit surprising because most of the fallows are located in the centerpart of the valley which is geomorphologically interpreted more as a deposition area. One explanation is that the higher K factor on the slopes of the western mountain range affects the spatial query, because a lot of fallow RUs were mapped there. In return that means that the western slopes are more erodible.

The overall trend of the Pasture class is a slight trend of deposition with very little variability. The mean trend of deposition agrees well with the geomorphological interpretation. Because most pastures are located at the valley bottom and in the lower course of the river this is a classical area for sedimentation. This agrees reasonable with a study from Australia. The mid-catchment is described as a zone of throughput or transfer. The sediment storage there is low, the transport capacity is moderate to high (Fryirs & Brierley 2001). Kakembo & Rowntree (2003) report on considerably little erosion on grazing lands in the Eastern Cape.

Milton et al. (1994) describe the degradation of rangeland in five steps. According to the description of the symptoms most pastures show step number 2 to 4 which is inter alia,

- plant and animal losses, reduced secondary productivity,
- perennial biomass reduced (short-lived plants and instability increase),
- bare ground, erosion, aridification.

Especially reduced biomass was mapped frequently, but there is still the possibility that the time of the mapping was unfavorable, since it was dry season. Nevertheless the vegetation was rather patchy or absent and under the current climatological conditions a dramatic shift in the vegetation cover is rather unexpected.

Generally orchards & vineyards is a small landuse class. But it clearly shows a trend towards deposition with 6.61 t ha^{-1} yr^{-1}. That is interpreted to be the likely result of the topography and its factor LS on the footslopes, since it is usually reported that monocropping increases erosion rates and degradation of soil structure (e.g. Pagliai et al. 2004; Lal et al. 2004; Chen et al. 2009). Scott et al. (1998) show that the replacement of natural fynbos communities by man-made plantation increases erosion rates markedly. This is dramatically intensified when wildfires come into play. Then water repellency in the soils is increased which causes higher frequencies and higher magnitude of overland flows. The suspended sediments amongst others hold pesticides. Such input occurs frequently in the region at least once per season (Schulz 2001). So it is likely that despite small in size the impact in one or another way is large.

Grassland is a primarily depositional class. This class does not suit a comparison to common landuse classes in the region. The term grassland in this study deviates from the grassland biome which in prominent in the east of South Africa (O'Connor & Kuyler 2009) and comprises most of the veld-type biomes. So no reasonable conclusion can be drawn. But the mean value of 18.19 t ha^{-1} yr^{-1} of deposition makes sense because most of the small landuse class is close to the foothills which makes deposition probable.

Soil Erosion & Deposition on Catchment Scale

When looking at the overall erosion rates and its spatial pattern it might be concluded that soil erosion is the most severe threat to the valley. But that is the erroneous way interpreting the map in Figure 5.12. Keeping in mind that it only shows a potential of surface processes – especially for a little instant of time – it is not as bad as seems. Platford (1982) found that some expected high soil losses for sandy soils did not materialize and expected moderate soil losses on fallows turned up to be much more severe than expected. In comparison to the rest of South Africa the erosion rates for Western Cape are low (Morel 1998). That goes along with the statement of Hoffman & Todd (2000) who report that soil degradation (what includes soil erosion) is not the foremost environmental issue in Western Cape.

From the field work it was recognized that in some areas of the catchment "ordinary" soil erosion is not the primary erosive process. In large areas close to the headwaters and sporadically further downstream gully incision is the main threat for current and future landuses. Contributing area, geology, landtype and soil erodibility are the most prominent factors for the development of gullies (Le Roux & Sumner 2012), but unfortunately USPED overshoots highly concentrated streamflow that leads to gully incision (Mitasova et al. 1999).

These gullies occur in abandoned or abandoned-like areas where the threshold for gullying is lowest (Lyons et al. 2013). Abandoned-like in this context means several frequently used off-road terrain courses that make use of the not agriculturally used slopes far upstream. In combination with partly patchy and partly shrub vegetation and with a generally high slope length and slope steepness factor that are ideal conditions for gullying and other linear erosive processes. That agrees with findings of Nortjé et al. (2012) from the Krüger National Park where repeatedly off-road driving leads to sub-soil compaction, limited infiltration and root penetration, less vegetation cover and an overall reduced resilience against several stressors. It is presumed to be overall the same: at beaches in more humid climates (Priskin 2003), in desert environments (Vollmer et al. 1977) and in mountainous environments – where more sediment load is produced by horses and hikers than by motorbikes (Wilson & Seney 1994). Another shortcoming of an USLE-derived soil erosion model is that it was applied numerous times in South Africa, but they never really have been comprehensively tested and calibrated (Breetzke et al. 2013). Additionally many of the input variables are subjective, because they are charged with alot of assumptions and derived from limited field data. So alternative but also valid estimations can lead to valid conclusions too. Thus errors propagate and accumulate through the spatial erosion model and can lead to erroneous results (Jetten et al. 1999). In this specific study there are several subfactors that can not be verified independently. Especially soil erodibility and the cover management factor appear to make sense whereas the ways of deriving them is not free of criticism. So the USPED model in this context must be seen through the eyes of George Box: "Essentially, all models are wrong, but some are useful." (Box & Draper 1987: p. 424)

7. Conclusions

The study at hand shows that the Standardized Precipitation-Evapotranspiration Index is capable of detecting dry and wet spells from a global data set for Mediterranean conditions in South Africa reliably on a mesoscale. According to this droughts are inherent part of the Western Cape climate and they seem to be recurrent on various cycles. The study shows that since 1948 moderate drought on all time scales is virtually normal condition. The reasons are manifold and partly not yet understood entirely. Man is still at the mercy of weather and climate and so meteorological and hydrological droughts are still something to cope with in the long run, but agricultural droughts get foiled by the excessive irrigation with groundwater and socio-economic droughts get void in a valley with less than 1000 permanent inhabitants. There is no obvious temporal pattern of the occurence of dry spells, wet spells seem to be more frequent in the austral winter.

Most rainfall is barely enough to be effective. Rainfalls with a significant amount are rare, the mean amount is triggered by single events. The precipitation inside the valley is spatially highly variable, clearly seasonal and since 1948 nonlinear slightly decreasing. The valley seems to be located in an extremely favored position, because rainfall variability, frequency and quantity of rain is clearly distinctive in comparison to a weather station from outside the valley.

The pattern of landuse and landcover follows the topography and thus the availability of (surface) water. The spatial pattern can be outlined to: the more distant from the headwaters the less water intensive the use. Only a very small portion of the valley is used heavily for agricultural purposes that are extremely water intensive, which is market-oriented production. Vast areas of fallow agricultural lands are indistinguishable from barren sediments. No natural forest can be found. As forest classified areas as well as open waters are entirely man-made. Wetlands are always related to the streams of the Krom Antonies and to the surroundings of the water storage dams. The landuse and landcover results are most likely distorted by a three-years crop rotation and indeterminable RUs during field mapping.

The Unit Stream Power Erosion Deposition model shows satisfactory trends but unsatisfactory values. The general trend can be agreed from the field work. But the approach of the model and the approach to derive its subfactors suffer from several shortcomings, such as the missing capability of the model to process highly concentrated streamflow what makes an accurate quantification impossible due to overestimation of soil loss, thus only reclassified values can be used. Additionally the rill-interrill factor (m) affects the final result massively and in the literature only vague recommendations instead of empirically derived values are given. Moreover the conversion of mapped field results to fit empirical USLE values does not agree well with data in the literature, especially not for the regional model SLEMSA. The derivation of organic carbon from surface photographs using a Partial Least Square Regression failed, thus the mean value of C_{org} was taken, so that spatial variability is not represented appropriately. The rainfall-runoff erosivity is as seasonal as the rainfall itself, thus the annual mean value for R neglects that approximately eight months a year the detachment capacity of rains is low and four months the erosivity of the rainfalls is quite high.

But despite the values are overshooting the resulting maps reveal a pattern that agrees with geomorphological interpretations. Especially the slopes are very prone to erosion. The valley bottom tends to morphological stability or at least its sediments do not get eroded by flowing water. Gullying in particular is an issue. In the Krom Antonies valley it is related to either improper landuse like the off-road track in the upslope area or the clearance of natural shrub vegetation. Mostly erosion or deposition is not explicitly related to a specific landuse or landcover, but rather to geomorphology. If at all, the landuse is a product of the catchment's geomorphology and hydrology.

Several topics are interesting, but are unfortunately not covered by this study. Further research might relate to:

- Vast areas of the valley are fallow or fallow-like most time of the year and from the field work the issue of soil loss by winds is recognized: a quantification of soil loss by eolian processes would surely reveal valuable results.
- Large quantities of groundwater are abstracted for waterintensive agriculture, but water use seems to be a political issue and gathering data about quality and quantity seems to be a political act as well: a procedure to measure all amounts of pumped water should be implemented to answer the question how much do the aquifers get stressed?
- Large amounts of water are wasted either due to inefficient irrigation, leaking pipes and wells or evaporation losses: the amount of "natural" water loss by evaporation from the dams or from the fields irrigated during the day should be quantified. In a follow-up step countermeasures should be contrived and implemented to save the scarce resource.
- The implementation of the National Water Act seems to have failed (Schreiner 2013): the water governance structures should be evaluated on a microscale to resolve the probing question *who takes when how much from where for what purpose?*

In summary: the valley is blessed with a favored location, but suffers from spatial and temporal highly variable rainfalls and frequent droughts. Only a small fraction of the valley is used for intensive agriculture, but the mean rainfall amounts are not sufficient for rainfed agriculture, thus groundwater is abstracted. Vast areas of the valley are left most time of the year "as is", what makes it prone to erosion and gullying in particular.

From this study the rainfalls in the valley are considered to be the most sensitive natural factor in a per se fragile geosystem. Rainfalls trigger aquifer recharge, agricultural productivity (and hence income) and biodiversity to a certain degree. An increase of rainfall variability is common sense against the background of a changing climate (Karl & Trenberth 2003) and it entails more, longer and more severe dry and wet spells. This also means a shift in farming practices and cropping calendars or even the disappearance of agriculture in some regions (Gbetibouo & Hassan 2005). Also it implicates an increased erosivity of rainfalls what in return means more loss of valuable land by either erosion or deposition, possibly accelerated siltation of the water storage dams and an overall increase of costs to maintain and to adapt infrastructure to mitigate respectively.

This fragile geosystem is also confronted with an increase of agro-industrial landuse. In Europe this kind of landuse the last 60 years was successful to meet the challenges of food security at the high costs of considerable degradation of soil and water resources (Boardman et al. 2003).

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Addendum

Table B 1: Grain-Size Classes [%] according to USDA 1987.

ID	<0.2 µm	<5 µm	<10 µm	<25 µm	<50 µm	<0.1 mm	<2 mm
I001	3.1	12.6	10.8	25.9	23.0	15.3	9.4
I002	1.9	12.1	12.9	35.7	19.4	14.8	3.1
I003	2.4	20.2	12.1	27.6	21.2	8.9	7.6
I005	1.0	4.1	5.1	30.9	40.0	15.7	3.3
I006	6.8	36.2	9.6	14.8	10.5	10.6	11.5
I007	3.6	27.7	10.9	19.5	15.3	10.8	12.1
I008	0.7	1.8	3.9	25.2	45.5	20.9	2.1
I009	0.9	3.2	4.5	18.3	34.1	34.6	4.5
I011	2.6	10.1	7.4	24.2	31.3	19.6	4.9
I012	2.6	13.5	7.6	20.8	29.7	22.7	3.1
I013	3.0	24.8	12.1	23.4	23.1	13.1	0.4
I014	3.3	21.9	9.9	18.6	14.2	9.2	22.9
I015	2.2	23.4	12.9	27.7	22.1	8.4	3.2
I017	3.3	18.1	10.8	22.6	21.9	18.2	5.0
I018	3.5	35.4	15.5	19.4	11.5	10.7	3.9
I019	2.0	8.2	7.2	18.7	34.9	26.8	2.1
I021	2.2	10.6	8.0	18.2	22.6	25.1	13.4
I025	3.7	27.5	10.7	32.3	17.5	8.4	0.0
I027	2.9	8.0	5.8	25.6	42.2	14.2	1.3
I032	3.4	23.7	10.9	23.0	18.4	10.5	10.1
I033	5.5	37.8	12.1	17.7	9.8	14.5	2.7
S001	3.5	22.5	10.8	18.7	20.5	14.7	9.4
S002	3.8	24.2	11.2	18.6	19.5	16.3	6.4
S003	3.5	12.1	9.3	17.1	23.3	23.6	11.0
S004	18.2	43.6	10.6	17.3	5.1	0.0	5.2
S005	8.3	22.2	11.4	26.4	18.7	10.1	2.8
S006	1.4	12.1	9.7	30.4	28.7	15.0	2.7

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ID	<0.2 µm	5 µm	<10 µm	<25 µm	<50 µm	<0.1 mm	<2 mm
S007	0.8	3.1	2.8	15.7	43.2	30.7	3.7
S008	0.8	2.3	3.4	15.1	37.3	41.0	0.1
S009	1.9	9.3	10.3	42.1	25.1	8.8	2.5
S010	2.6	10.1	6.4	19.3	31.9	26.9	2.8
S011	0.7	2.7	2.3	15.7	47.0	29.1	2.6
S012	0.7	2.9	4.0	17.6	41.7	29.5	3.6
S013	1.9	10.8	5.8	19.3	32.6	23.3	6.3
S014	1.6	13.0	10.0	20.5	26.8	19.4	8.7
S015	3.1	14.6	9.7	22.2	21.9	20.6	8.0
S016	3.0	9.8	7.5	18.0	28.6	28.9	4.2
S017	3.5	15.6	10.6	20.6	24.6	20.2	4.9
S018	1.5	7.3	6.3	21.0	32.8	27.8	3.4
S020	1.4	6.7	4.0	20.2	40.7	25.0	1.9
S021	0.6	2.1	1.3	13.8	42.1	36.5	3.7
S022	2.7	13.6	9.4	23.9	29.4	17.0	4.0
S023	2.5	16.7	7.8	18.7	22.3	21.4	10.6
S024	5.1	34.9	8.5	15.0	14.2	15.9	6.5
S025	1.6	8.2	5.0	15.9	34.1	32.0	3.3
S026	1.9	11.0	8.4	32.8	31.9	12.4	1.6
S027	0.8	4.7	5.4	24.3	39.3	24.1	1.4
S028	2.8	18.2	10.6	23.5	27.9	15.5	1.5
S029	2.8	6.1	4.4	16.5	40.8	27.6	1.8
S030	2.9	18.3	9.8	22.7	21.6	18.6	6.0
S031	1.6	8.0	5.5	22.2	40.3	22.3	0.2
S032	1.7	8.5	3.0	18.1	43.5	23.4	1.8
S033	0.6	1.9	2.6	11.0	31.8	44.9	7.2
S034	2.9	21.3	10.8	22.7	20.1	13.9	8.3
S035	1.4	5.2	3.5	17.0	35.8	30.2	6.9
S036	3.0	19.2	9.5	20.2	21.8	16.8	9.5
S037	2.3	9.7	5.3	14.5	26.5	31.2	10.6
S038	4.3	28.9	11.3	18.7	11.2	11.9	13.7
S039	9.0	49.2	10.3	11.6	5.7	7.0	7.1

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ID	<0.2 µm	5 µm	<10 µm	<25 µm	<50 µm	<0.1 mm	<2 mm
S040	4.5	26.5	11.6	18.1	11.4	15.9	12.0
S041	0.7	2.1	2.7	10.9	43.2	39.5	0.8
S042	2.3	8.8	7.6	31.6	34.5	14.7	0.6
S043	1.2	3.8	4.2	15.2	31.9	37.2	6.5
S044	1.0	4.0	4.3	19.1	36.6	27.4	7.6
S045	0.8	2.7	2.7	13.2	45.6	32.2	2.7
S046	1.9	13.2	9.2	22.9	33.9	18.7	0.0
S047	4.8	32.1	9.2	13.9	10.4	14.1	15.5
S048	1.4	10.6	9.7	27.7	24.2	19.7	6.7
S049	1.3	4.6	6.2	24.7	33.7	24.9	4.7
S050	1.2	2.3	0.7	5.9	34.0	51.8	4.0
S051	2.3	7.5	6.0	21.4	32.7	27.6	2.6
S052	0.7	2.9	3.3	17.0	37.4	32.6	6.0
S053	1.0	3.5	3.7	17.5	37.9	32.0	4.3

Table B 2: Soil Permeability Classes after RÖMKENS et al. 1997 and GIOVANNINI et al. 2001.

Texture	Description	Class	Saturated Hydraulic Conductivity [in hr^{-1}]
Silt clay, clay	very slow	6	<0.04
Silt clay loam, sand clay	slow	5	0.04 – 0.08
Sandy clay loam, clay loam	slow – moderate	4	0.08 – 0.2
Loam, silt loam	moderate	3	0.2 – 0.8
Loamy sand, sandy loam	rapid – moderate	2	0.8 – 2.4
Sand	rapid	1	>2.4

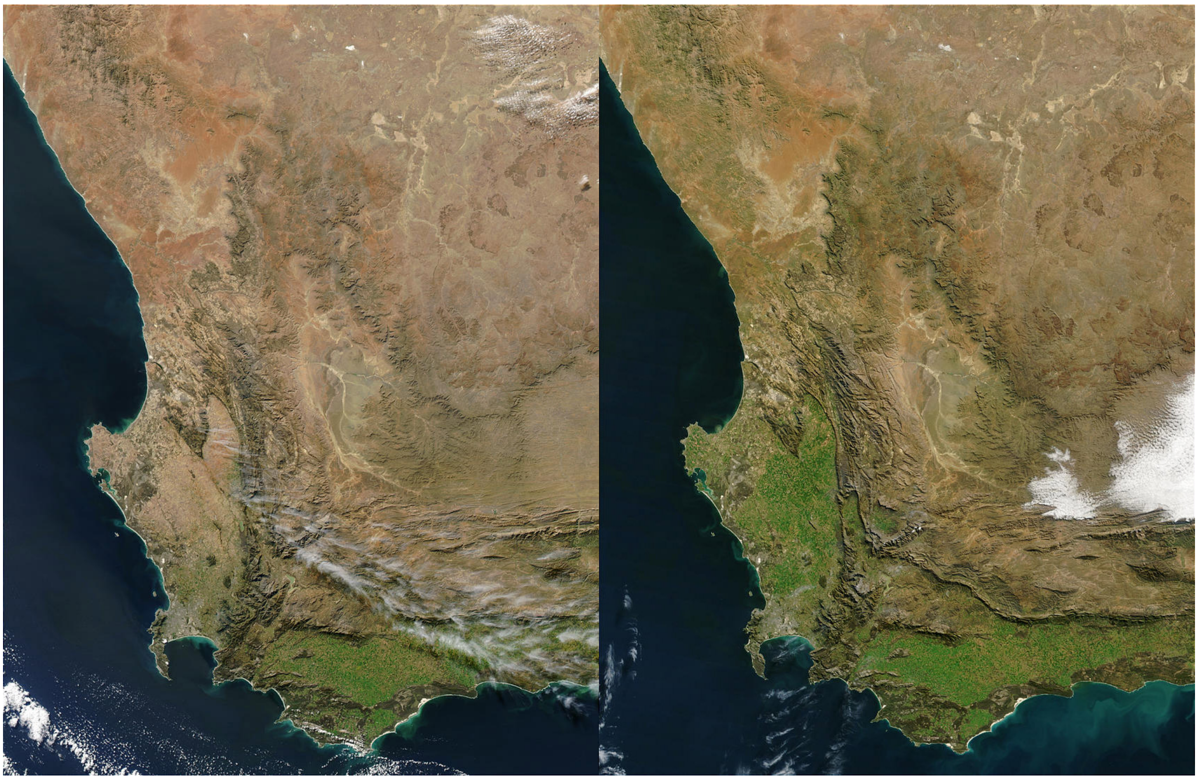

Figure A 1: Modis scene from 2003 drought in Western Cape. Original quote from NASA Earth Observatory (2003): The current drought in South Africa's Western Cape province is withering and stunting vegetation across this crop-producing region. The pair of images above shows southwestern Africa on July 21, 2003 (left), and July 21, 2002 (right). Vegetation health (greenness) is dramatically reduced in the left hand image, particularly in the region between Cape Columbine (left center edge) and southward to the Cape of Good Hope (narrow strip of land on the outside of the horseshoe-shaped False Bay). Most of South Africa's wheat is produced in the Western Cape, with the areas in the southwest corner of the province being among the most important wheat-producing regions in South Africa. Erratic rainfall in the Western Cape often produces wide variations in wheat yields and quality, with the country having a surplus of wheat only during very good rainfall years and shortages during the majority of years. Low rainfall and soil moisture conditions in the Western Cape will reduce this year's wheat and barley crops. The high-resolution image provided above is 500 meters per pixel.

Table B 3: Statistics of the Landcover Classes and the corresponding Erosion and Deposition values from USPED model. Columns 8 to 12 show the statistics for the central 98 % as an approach to get rid of the extreme values caused by some USLE factors. All units except Area are in t ha^{-1} yr^{-1}.

		Complete Data Set					1st to 99th Percentile				
Class	**Area [ha]**	$\bar{x}$	**s**	$\sum$	**Minima**	**Maxima**	$\bar{x}$	**s**	$\sum$	**Minima**	**Maxima**
Open Water	35.82	16.98	14974.72	-204036.56	-388143.56	154803.27	16.98	814.54	14979.65	-9367.30	6832.48
Low Intensity Residual	54.19	-1.28	407.78	-22604.06	-10251.29	1660.45	-1.28	42.76	-1710.62	-394.05	379.97
Barren Rocks & Sediments	4439.36	-13.59	34727.29	-21493411.78	-3412822.50	3373392.00	-13.59	845.41	-1444970.82	-11090.86	10553.65
Evergreen Forest	19.26	-1.75	31.06	-773.95	-169.37	177.32	-1.75	21.03	-815.62	-137.22	140.16
Shrubland	3822.73	19.19	66248.69	26046553.12	-4733437.00	4733915.50	19.19	1285.71	1780644.69	-12082.14	13483.32
Orchards & Vineyards	315.43	6.61	12818.48	-1332707.65	-321556.06	323830.50	6.61	582.37	50221.99	-7363.05	8024.68
Grassland	141.01	18.19	30248.43	6920210.93	-195106.73	736880.69	18.19	247.93	63011.12	-1655.41	4240.53
Pasture	794.85	1.67	4060.73	-291921.29	-194772.98	103589.15	1.67	203.45	32284.01	-2414.16	2921.66
Fallow	2135.93	-13.49	15405.59	-3740758.57	-743020.12	743043.75	-13.49	538.05	-703779.16	-7729.04	6133.91
Woody Wetlands	162.95	200.43	60055.28	-5874881.40	-1697399.38	480704.19	200.43	12013.48	799115.19	-115355.14	107968.58
Herbaceous Wetlands	50.57	-39.84	15615.71	-11764.98	-143756.25	157354.27	-39.84	5147.68	-49804.34	-53730.11	57061.52

(a) Barren Sediment (former field)

(b) Barren Sediment (natural)

(c) Barren Rock

(d) Orchards & Vineyards

(e) Fallow

(f) Shrubland

(g) Pasture (near outlet)

(h) Pasture (further upstream)

Figure A 2: Examples of landuse and landcover classes (low intensity residual is missing).

(i) Grassland with stabilized gully incision

(j) Evergreen Forest (planted oaks)

(k) Herbaceous Wetlands

(l) Woody Wetland (upper course)

(m) Woody Wetland (middle course)

(n) Woody Wetland (lower course)

(o) Open Water (downstream)

(p) Open Water (upstream)

Figure A 2: . . . continued

Supplementary Data (DVD)

- 00_R-SCRIPTS
- 01_DATA
  - Climate Data (official: Riviera, Piketberg, Vaandrigsdrift, inofficial: Kromvlei)
  - Attributes of Response Units
  - NCEP data
  - SOIL_PICTURES (folder contains top soil photographs)
- 02_SPATIAL_DATA (R = Raster; V = Vector)
  - Aerial Photographs (R)
  - Contour Data (V)
  - Response Units (V)
  - Sample Locations from GPS (V)
  - Lithology (V)
  - DEMs (R)
  - Weather Stations (V)
- 03_LABORATORY_DATA
  - BECKMAN_DATA (folder containing raw data from laser diffractometer)
  - Sample Weights
  - LECO Data
- 04_KROM_ANTONIES_R_WORKSPACES
- 05_KROM_ANTONIES_GRASS_LOCATIONS
- 06_TABLES
- 07_LITERATURE
- KROM_ANTONIES_BIBLIOGRAPHY.BIB
- KROM_ANTONIES_THESIS_NORBERT_ANSELM.PDF
- LICENSE.TXT
- README.TXT (documentation)

Erklärung

Hiermit erkläre ich, dass ich die Masterarbeit *Assessment of Landscape Sensitivity in the semiarid Krom Antonies River Catchment, Western Cape, South Africa* selbständig angefertigt und keine anderen als die zitierten Quellen, Hilfsmittel und Codes verwendet habe.
Des Weiteren erkläre ich, dass ich diese Arbeit nicht in anderen Prüfungsverfahren eingereicht habe.

Berlin, 27$^{\text{th}}$ February 2015

(Norbert Anselm)