Rainfall variability and drought in the central and northern KwaZulu-Natal Drakensberg: 1955 – 2015

by

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DECLARATION

I, **Catherine Ann Smart,** declare that this dissertation, entitled **Rainfall variability and drought in the central and northern KwaZulu-Natal Drakensberg: 1955 - 2015,** which I hereby submit for the degree of **Master of Science (Environment and Society)** at the University of Pretoria, is my own work and has not been previously submitted by for a degree at this, or any other, tertiary institution.

______________________	6/12/2017
Signature	**Date**

ABSTRACT

This study assesses the temporal rainfall variability trends, the effects on river discharge and the identification of drought periods across the central and northern Drakensberg (1955 – 2015), as well as at a catchment scale, in the Sterkspruit Catchment. Records from five stations, covering the central and northern Drakensberg, indicate that there is a statistically significant decrease in rainfall over the recording period. The long-term trends in interannual variability show an increase in the variability of annual rainfall over the five stations. The mean annual rainfall in the Drakensberg is highly seasonal and an analysis of the monthly rainfall indicates an increase in the variability of the distribution of monthly rainfall and the strengthening of rainfall seasonality in the Drakensberg, which is shown by a statistically significant decrease in autumn rainfall. The El Nino Southern Oscillation (ENSO) influences summer rainfall variability in the Drakensberg and a strong correlation exists between the summer rainfall and the Southern Oscillation Index, for the preceding periods, suggesting that changes in the intensity and frequency of the ENSO should negatively affect the rainfall in the Drakensberg. A cyclicity of approximately 10 years between dry periods was found, from 1955-2015, and it is noted that the cycles are becoming increasingly variable over time. An analysis conducted across four stations in the Sterkspruit Catchment assessed valley-scale variability and how rainfall variability and drought conditions may affect agriculture in the area. The Sterkspruit Catchment has experienced an increase in rainfall variability and dry conditions in the 21st century. The assessment of meteorological droughts, at both study scales, using the Standard Precipitation Index, displayed a significant decreasing trend, indicating an increase in the number of dry years over time. Hydrological drought was assessed, using the Streamflow Drought Index for two rivers in the Drakensberg. Both drought indices found an increase in the number of dry years experienced over time, and they highlighted 1982, 1992, 1994, 2003, 2007 and 2015 as being the years with the lowest rainfall and experiencing drought conditions. Discharge in the Drakensberg reflects the rainfall and seasonal trends and the Little Tugela River, at the bottom of the Sterkspruit Catchment, shows decreasing discharge trends over time. Farmers in the Sterkspruit Catchment rely on the Bell Park farm dam water to irrigate crops throughout the year because the rainfall is unreliable, especially the early-seasonal and summer rain. However, in 2015, the dam ran dry for the first time since it was built. Thus, more insight into rainfall trends and cyclcity in the Drakensberg could help farmers to understand and plan for the periods with low rainfall conditions.

Key Words: Rainfall variability, seasonality, drought, discharge, rainfall cyclicity

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TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

1. INTRODUCTION

1.1 Background

Over recent decades, significant changes in the mean state of the climate have occurred in most parts of the world (Kruger and Nxumalo, 2017). In particular, trends in precipitation have become increasingly variable across the globe (Donat et al., 2013). Rainfall has been regarded as one of the most significant climate parameters affecting human activities and these variable rainfall trends are thus of interest to various socio-economic sectors, such as agriculture and water resource management (Kruger and Nxumalo, 2017). In addition to the trends in seasonal and annual rainfall totals, the trends in extreme precipitation deficits are of importance, in terms of assessing the occurrence and likelihood of drought conditions. Owing to rainfall being of upmost importance to society, the economy (agriculture in particular) and water resource management, the understanding and prediction of rainfall variability is regarded as being a priority (Ratna et al., 2014).

Southern Africa's geographic position, its steep orography, its contrasting oceanic settings and atmospheric dynamics enhance the incidence of extreme weather events and interannual variability of the hydrological cycle (Fauchereau et al., 2003). Rainfall in southern Africa has been found to be decreasing as a result of changing and variable climates (Hulme et al., 2001; Arnell et al., 2003; IPCC, 2007a; Mazvimavi, 2011; Kusangaya et al., 2014). In addition precipitation regimes have become more variable over the past decades, with longer dry periods and an increase in the number of extreme rainfall events (Smakhtina, 1998; Richard et al., 2001; Kruger, 2006; Koerner and Collins, 2014). It is important to understand rainfall variability over southern Africa and South Africa, because of the prominence of rain-fed and irrigated agriculture in the region, the arid to semi-arid nature of the climate and the high degree of interannual and interdecadal variability observed over the region (Todd et al., 2004). Thus, southern Africa is one of the regions that is most sensitive to precipitation shifts and variability (IPCC, 2007b) and where the future rainfall distribution, frequency and variation remain significant for research (Gordon et al., 2000).

South Africa is classified as a water-stressed country, with scarce water resources, as the average rainfall of 600 mm year is well below the world average of 860 mm (RSA, 2002; Otieno and Ochieng, 2004). The natural availability of water across the country is also uneven and this is compounded by the strong seasonality of rainfall, resulting in large spatio-temporal rainfall variability (Ratna et al., 2014). The rainfall experiences strong interannual fluctuations, with recurrent wet and dry periods, in response to the coupled ocean-atmosphere modes of variability over the Pacific, Indian, and Atlantic Oceans (Dieppois et al., 2016). This includes factors such as; the penetration of moisture from sources in the southwest Indian Ocean, which are associated with movements of the Inter Tropical Convergence Zone, the moisture penetration from the south-east tropical Atlantic (Reason, 2001; Cook et al., 2004) and the topographic effects (Thomas et al., 2007). Rainfall-producing systems include tropical temperature troughs, westerly troughs, mid-latitude dynamics, cut-off-low pressure systems and thunderstorms (Ratna et al., 2014).

1.2 Rainfall Variability

The important variables to be considered when assessing rainfall variability are reflected in the characteristics of rainfall distribution throughout the year. These include the timing of the onset and end of the rainy season (seasonality), the wet and dry spell duration periods (cyclicity) and the occurrence of high and low rainfall events (MacKellar et al., 2014). Substantial research has been focused on identifying rainfall patterns over time and more particular attention has been on rainfall cycles at continental, regional and local scales (Nicholson, 2000; Nel, 2009). Much of the research on the summer rainfall area of South Africa found a quasi-20-year rainfall oscillation pattern (Dyer and Tyson, 1977; Vines, 1980; Jury and Levey, 1993; Tyson et al., 2002; Rouault and Richard 2003; 2005; Kane, 2009; du Toit and O'Connor, 2014; Malherbe et al., 2015). Recently, however, Jury (2015) found that rainfall across South Africa exhibited a 12-year oscillation over the period 1960-2010. A range of short period cycles, in the order of 2-7 years (Vines, 1980; Jury and Levey, 1993; Kane, 2009) and 10-12 years (Tyson et al., 1975; Vines, 1980; Malherbe et al., 2015) have also been identified. Previous research in the KwaZulu-Natal Drakensberg indicates rainfall oscillations of 16-20 years (Tyson et al., 1975) and 10-20 year oscillations (Nel, 2009).

Studies have shown that there has been a tendency towards a significant decrease in the number of rain days over time, across almost all of South Africa (DEA, 2013), as well as an increase in the intensity of rainfall events and an increased dry spell duration (Smakhtina, 1998; Mason et al., 1999; Richard et al., 2001; Kruger, 2006). A decrease in rainfall over the eastern and north-eastern regions of the country is evident (Richard et al., 2001; Kruger, 2006; MacKellar et al., 2014; Kruger and Nxumalo, 2017) and this decreasing rainfall in the east is more pronounced in the work of MacKellar et al. (2014), who saw statistically significant decreases in the rainfall and the number of rain days over the central and north-eastern parts of the country, in the autumn months.

Nel (2009) demonstrated a shift in seasonality at stations across the KwaZulu-Natal Drakensberg for 1955–2000, where the mean annual precipitation showed no significant trend, but an increase in summer rainfall was accompanied by decreased autumn and winter rainfall, resulting in a shorter wet season and a more pronounced seasonal cycle. MacKellar et al. (2014) found an increase in spring rainfall in the southern KwaZulu-Natal Drakensberg over time, which suggests an early seasonal onset of rainfall. This is an indication that decreases in rainfall in these particular areas have become prominent during recent decades, especially in autumn, and that this contributes to the increased rainfall seasonality (Kruger and Nxumalo, 2017). Such findings are consistent with the results from Thomas et al. (2007) for north-west KwaZulu-Natal that showed an increase in early-season rainfall, with a decrease in late-season rainfall between 1950 and 2000. Similarly, seasonal shifts were also observed in Limpopo over the same period, where there has been a tendency for a later seasonal rainfall onset, accompanied by increased dry spells and fewer rain days (Tadross et al., 2005; Thomas et al., 2007).

Recent droughts in southern Africa and South Africa have highlighted the importance of understanding interannual rainfall variability in the region. In 2015, the annual rainfall total in the summer rainfall season of South Africa was 403 mm, which was below the 608 mm average calculated over a 112-year period (SAWS, 2016). It was established that 2015 was the year with the lowest annual total rainfall on record for South Africa, causing widespread water shortages and drought conditions across various parts of the country (SAWS, 2016).

1.3 Drought

South Africa is known to be subject to the occurrence and impacts of droughts and floods, because the climate is characterised by a high degree of interannual rainfall variability (Dent et al., 1987; Richard et al., 2001; Rouault and Richard, 2003; Thomas et al., 2007; Vogel et al., 2010). A drought may be loosely defined as a deficit of water relative to the mean conditions experienced for a particular area (Sheffield and Wood, 2012); however, droughts are characterized by their timescale, intensity, extent and impact, which vary between different types of drought (Keyantash and Darcup, 2002). A meteorological drought is based on a precipitation deficiency that is relative to the normal conditions and this deficit subsequently causes low recharge from the soil to water features, such as rivers and dams, creating a delayed hydrological drought (Tallasken and van Lanen, 2004).

Rainfall is the meteorological variable with the longest time series of data over South Africa and thus rainfall data can be used to evaluate aspects of drought (Malherbe et al., 2015). Rouault and Richard (2003) demonstrated the effective use of the Standardized Precipitation Index (SPI; McKee et al., 1993) for monitoring the intensity and spatial extent of meteorological droughts, at various timescales, for South Africa. The index quantifies the precipitation deficits at various timescales and provides an indication of the drought intensity and extent, based on the historical distribution of rainfall (Malherbe et al., 2015). South Africa experienced at least eight major droughts in the 20th century, namely, 1926, 1933, 1945, 1949, 1952, 1970, 1983, 1992 and in 2003 (Rouault and Richard, 2005; Mussa et al., 2014), with the most severe droughts occurring in 1983, 1992-1995 and 2003-2004 (Mussa et al., 2014). This correlates with extreme droughts over the African continent that occurred in 1972-1973, 1983-1984 and 1991-1992 (Masih et al., 2014). However, the recent drought of 2015 in South Africa stretched into 2016 and it was the worst drought since 1982 (Corke and Whittles, 2015). The drought was concentrated more intensely in KwaZulu-Natal and the Free State, which were subsequently considered to be disaster areas (Essa, 2015).

The impacts of hydrological drought are important, especially when they concern the reservoir levels and streamflow discharge (Malherbe et al., 2015). Vogel (2000) identified a number of hydrological droughts in South Africa from the 1920’s onwards, including the periods 1930-1932, 1950-1953, 1967-1972, 1978-1886 and 1989-1991. An analysis of summer droughts by Malherbe et al. (2015) found that the years with the most extensive extreme and severe hydrological droughts calculated over a two-year period included the following: 1947, 1933, 1966, 1970, 1983, 1992, 1993, 2004 and 2005. A study in the Limpopo River Basin from 1979-2010, using the Standard Runoff Index, identified that the most severe hydrological droughts were experienced in 1982/1983 and 1991/1992 (Trambauer et al., 2014). The severe droughts in the early 1980s and 1990s led to a decrease in crop and stock production and the reservoir water levels dropping dramatically (Harsch, 1992; Vogel, 2000).

The position of South Africa, within the subtropical high-pressure belt, influences the climate variability related to the position and strength of the high-pressure systems over the Atlantic to the west, and the Indian Ocean to the east, during the rainy season (Jury, 1996; Richard et al., 2001; Mulenga et al., 2003; Kane, 2009). Mason and Jury (1997) provide a synopsis of the earlier work on regional circulation patterns and climate variability in South Africa. The El Nino Southern Oscillation (ENSO) is integral in the summer rainfall region where a negative correlation exists between the ENSO and rainfall (Nicholson and Entekhabi, 1986; Ropelewski and Halpert, 1987; van Heerden et al., 1988; Nicholson and Kim, 1997; Rouault and Richard, 2005; Ratnam et al., 2014; Dieppois et al., 2015), especially since the late 1970s (Richard et al., 2000, 2001; Gaughan and Waylan, 2012). The ENSO causes widespread warming and affects the atmospheric stability tropic-wide, on an interannual timescale (Parih et al., 2016). The ENSO events in the equatorial Pacific have long been believed to be the major mechanism for seasonal rainfall variation over southern Africa; hence, recurring droughts are attributed mostly to the influence of El Nino (Mantasa et al., 2008).

ENSO has been attributed to the drought conditions across South Africa. Rouault and Richard (2003) found that 8 out of the 12 droughts that occurred during the 20th century can be attributed to the ENSO conditions, while Gaughan and Waylen (2012) show that the incidence of dry years associated with the El Niño events has been increasing in the last decades. Kurukulasuriya and Rosenthal (2013) found that droughts are projected to increase in intensity and frequency in southern Africa, with ENSO contributing significantly to the climate variability in the region. This confirms the strong relationship between the ENSO and drought events in South Africa. The 2015 drought over South Africa was driven by an ENSO event that initiated widespread dry conditions over much of

South Africa, which is considered to be the worst drought in South Africa since 1982 (Corke and Whittles, 2015). It is thus clear that Southern Africa and South Africa are experiencing multi-decadal trends that are highlighted by declining mean annual precipitation, increasing rainfall variability, drier conditions and an increased number of warm phase El Nino Southern Oscillation events (Gaughan and Waylen, 2012).

1.4 Rationale for the study

The KwaZulu-Natal Drakensberg reaches altitudes of over 3000 m and forms part of the Main Escarpment of southern Africa, which also acts as the watershed border between South Africa and eastern Lesotho (Nel and Sumner, 2006). Precipitation estimates for the escarpment region reach up to 2000 mm/a (Tyson et al., 1976; Schulze, 1979). The KwaZulu-Natal Drakensberg catchments are known to produce nearly twice as much total runoff per unit rainfall than the average for South Africa, as a whole (Schulze, 1979) and a quarter of South Africa's streamflow (Whitemore, 1970). Owing to these facts, the water from the upper catchments is used to supply water for the Tugela-Vaal and Lesotho Highlands water transfer schemes that feed the Vaal Dam and provide water to the Gauteng Province. In the foothills of the Drakensberg, the runoff generated feeds into rivers that supply water used by farmers for irrigation and for domestic use in the surrounding towns. The understanding of rainfall trends and variability in this area is thus of great significance, given the nature of the water demand, both locally and nationally.

Given that this area is known for its high rainfall totals and runoff generation, farms in the central and northern Drakensberg are heavily reliant on irrigated agriculture, because the rainfall in the area is increasingly variable and therefore unreliable. One such example is the Sterkspruit River Catchment area, where farms rely on irrigation and thus need farm dams to store water. The Bell Park Dam that was constructed after the drought in the early 1980s severely impacted the farms. Farms in the Sterkspruit Catchment rely on water from the Bell Park Dam to survive, because the rainfall is not dependable (Stockil, 2017 pers. comm.).

In the summer months, livestock and dairy farmers grow feed and silage to make up for the shortfall in yield during the autumn and spring months, and most of the silage is grown under irrigation (Stockil, 2017 pers. comm.). Crop farmers plant wheat in winter, under irrigation, and plant soya and maize from October to November. If the early seasonal rains do not arrive (late autumn and spring rains) farmers have to adapt their strategies and plant irrigation crops first, and when the rain arrives they then plant the dryland (rainfed) crops. Thus, the reduced rainfall in autumn and spring is mitigated by using water for irrigation from the farm dams. As long as the farm dams have a sufficient amount of water for irrigation purposes in them, the situation in the catchment can be managed, but not without the increasing costs of equipment and electricity to fuel the process (Stockil, 2017 pers comm). However, for the first time since the dam was built in the early 1980s, the Bell Park Dam ran dry in 2015.

Even though the area is crucial for runoff generation, studies investigating the long-term trends of rainfall and variability have only been conducted fairly recently. The work of Nel and Sumner (2006) and Nel (2009) brought about further insight into the limited knowledge of rainfall totals and rainfall variability for the KwaZulu-Natal Drakensberg (1955- 2000), extending the benchmark assessments done by Tyson et al. (1976) and Schulze (1979). The KwaZulu-Natal Drakensberg experienced interannual rainfall variability (Nel and Sumner, 2006) and demonstrated the significant seasonality of monthly rainfall (Nel, 2009). Nel (2009) did not, however, identify the occurrence of drought events. This study will thus update the assessment of the long-term rainfall records to 2015 in the KwaZulu-Natal Drakensberg and it will analyse those records within the sub-catchment that hosts the Sterkspruit River. A temporal analysis of the rainfall trends and variability will be established for the two areas and the low flow and drought periods will be identified.

1.5 Aim and Objectives

1.5.1 Aim

The aim of the study is to determine the temporal rainfall variability trends, the effects on river discharge and the identification of drought periods across the central and northern Drakensberg (1955-2015), as well as at a catchment scale in the Sterkspruit Catchment.

1.5.2 Objectives

1.5.2.1 Objective 1

To determine the temporal trends in rainfall variability and drought within the central and northern KwaZulu-Natal Drakensberg area.

1.5.2.2 Objective 2

To determine the temporal trends in rainfall variability and drought within the Sterkspruit Catchment area.

1.5.2.3 Objective 3

To determine the effects of rainfall variability on river discharge, using two stations located in the central and northern KwaZulu-Natal Drakensberg.

1.5.2.4 Objective 4

To outline the underlying cause of rainfall variability and drought in the central and northern KwaZulu-Natal Drakensberg.

2. LITERATURE REVIEW

2.1 Introduction

The following literature review aims to discuss the pertinent literature and to evaluate the methodology, which is relevant to this research. The focus of this study is on the southern and South African region, and specifically on the local the KwaZulu-Natal Drakensberg. This review includes an explanation of the theoretical and concepts that are relevant for a holistic understanding of climate, rainfall variability and drought, and it provides a sound discussion of the literature that is relevant to this study.

Climate change will be first established as a global and localized problem, especially with regards to climate variability. The drivers behind the climate variability and rainfall patterns in southern Africa will then be explored. Studies on rainfall variability in southern Africa and South Africa will be examined, providing a contextual background of the past rainfall trends. The second aspect of the review defines, characterizes and contextualizes droughts and appraises the previous studies on droughts in southern Africa and South Africa. The different methods used to quantify and classify drought will be explained and supported by global and local examples. Various methodologies will be reviewed, however, the emphasis will be placed on those that are applicable to the characteristics of the region under study and the amount and type of data available.

The main focus of this chapter will be to highlight the past rainfall trends in southern Africa and South Africa, to identify patterns in years of low and high rainfall, to identify the years of droughts, and finally, to introduce the motivation for the study.

2.2 Climate and Climate Variability

Global climate change is one of the most serious environmental challenges that the world is presently facing (IPCC, 2014). It is related to systematic changes of the entire world's weather and climatic patterns, beyond the limits of natural variability, and increased droughts have been identified as being among the consequences thereof (Massa et al., 2014). The Intergovernmental Panel on Climate Change (IPCC) Fourth and Fifth

Assessment Report confirms and reinforces the evidence that climate change is real and poses a serious environmental, social and economic threat (IPCC, 2007a; 2014). According to the IPCC, the recent warming of the climate system and the many of the observed changes have been unprecedented in previous decades and millenia. The atmosphere and ocean have warmed, there has been widespread melting of snow and ice and the global average sea level has risen (IPCC, 2007b, p.5; IPCC, 2014, p.2)

Technically, climate change has been defined as "statistically observable changes in the mean of climatic properties over an extended period of time", and climate variability is defined as "statistically observable changes of the climate over different spatial and temporal scales" (IPCC, 2007b, p.5). These changes in the weather consitions, namely, the overall warming of temperatures, the changes in rainfall patterns, the increasing intensity of extreme events and the increasing frequency and intensity of drought events, consitute a serious global problem (Carter and Gulati, 2014). Climate change is projected to have varying spatial impacts around the world, some of which are already undeniably clear and already affecting ecosystems, biodiversity and human systems throughout the world (Kotir, 2011).

Weather refers to the state of the atmospheric or meteorological conditions experienced at a particular time and place, which can be described for one particular weather station or for a specific area of the earth's surface (Vogel, 2000; Martin, 2005). In contrast, climate is the term given to the prevailing condition of the atmosphere deduced from long periods of weather observation (Martin, 2005). Thus, climate is a generalization, while weather reflects a specific event (Martin, 2005). The analysis of the weather allows for a long-term picture of the climate to be established and used to explain the physical processes underpinning the patterns of climate, in both time and space (Vogel, 2000). World climates are seen as dynamic, interconnected systems that are characterized as much by their variability and propensity for change, as they are by their 'normal' conditions (Vogel, 2000).

The climates of Africa are both varied and varying: varied, because they range from humid equatorial regimes, through seasonally-arid tropical regimes, to sub-tropical Mediterranean-type climates, and varying, because all these climates exhibit differing degrees of temporal variability, particularly with regard to rainfall (Hulme et al., 2001).

The climate of the African continent is controlled by complex maritime and terrestrial interactions that produce a variety of climates across a range of regions e.g. from the humid tropics to the arid Sahara (Christensen et al., 2007). Climate also plays an integral part in determining the resource base for economic development in Africa, particularly for the agricultural and water resources sectors, at regional, local and household scales (Vogel, 2000; Parry, 2007). A country may be determined more by the extremes of climate than by the long-term averages, such as in southern Africa, which is characterised by the oscillation between droughts and floods, making for variable climates (Vogel, 2000). A characteristic of the southern African climate is its variability of the climate at intraseasonal (Pohl et al., 2007; Mapande and Reason 2005), interannual (Lindesay 1988; Mason and Jury 1997; Reason et al., 2006), interdecadal (Tyson et al., 1975; Reason and Rouault 2002) and longer timescales. Understanding and predicting these interannual, interdecadal and multidecadal variations in climate has become the major challenge facing African and African-specialist climate scientists in recent years (Hulme et al., 2001).

2.2.1 Southern African climate

The subtropical high-pressure belt that dominates the weather over Southern Africa lies between the South Atlantic and Indian Ocean subtropical high-pressure cells, in a region that is subject to the interaction of tropical easterly and extra-tropical westerly airflows (Dyson and van Heerden, 2002). Variations in the position and intensity of the two high-pressure systems play an important role in the rainfall distribution over South Africa (Schulze, 1965). The seasonal longitudinal shifts of the subtropical high-pressure cells are especially significant to the climate of southern Africa, with resepct to the Indian Ocean cell (Barry and Chorely, 1992). The mid-latitude westerly circulation, extending northwards to, and in association with, these two high-pressure systems, largely controls the weather of southern Africa (Dyson and van Heerden, 2002). Both these high pressure cells shift westwards and intensify in the southern winter, but because the South Atlantic cell always extends further north than the Indian Ocean cell, it brings low level westerlies to Angola and Zaire in all seasons and high-level westerlies to central Angola, in the southern summer (Barry and Chorely, 1992). As the high-pressure systems migrate southwards during the summer months, the influence of the westerly circulation is diminished (Dyson and van Heerden, 2002). Tropical easterly airflows affect much of southern Africa throughout the year (Barry and Chorely, 1992).

A feature of the climate of southern Africa is the prevalence of wet and dry spells that are associated with broader features of the global circulation (Barry and Chorely, 1992). Above normal rainfall occurs in a as a north-south belt over the region and is associated with a high-phase Walker circulation that has an ascending limb over southern Africa. It is also associated with the strengthening of the Intertropical Convergence Zone (ITCZ), an intensification of tropical lows and easterly waves, often in conjunction with a westerly wave to the south, and a strengthening of the South Atlantic sub-tropical high pressure cell (Barry and Chorely, 1992). Rainfall below normal is associated with a low-phase Walker circulation having a descending limb over southern Africa, a weakening of the ITCZ, a tendency to high pressure with fewer occurrences of tropical lows and easterly waves, and a weakening of the South Atlantic subtropical high-pressure cell (Barry and Chorely, 1992).

The southern African region thus has a dry climate and a high degree of interannual rainfall variability, which impacts greatly on the water resources, on agriculture and the rural communities. Examples include the floods in north-eastern South Africa/southern Mozambique during February/March 2000 and the severe droughts of 1991/1992, 2002/2003 and 2003/2004 over northern South Africa and surrounding areas (Cook et al., 2004).

2.2.2 South African climate

The weather and climate of South Africa are strongly influenced by the position of the country, relative to the global circulation patterns (Vogel, 2000). The country is largely a summer rainfall region (Fauchereau et al., 2009), with winter rainfall occurring over the south-western Cape area (Philippon et al., 2011). The southern areas of South Africa largely receive rainfall from organized synoptic-scale weather systems (Tennant and Hewitson, 2002), whilst rainfall over the northern and eastern interior regions are mostly of a convective and sometimes tropical nature (Malherbe et al., 2012).

For most of the year, the atmospheric circulation over South Africa, especially the central and southern regions, is dominated by extra tropical weather systems, such as cut-off lows, cold fronts and the ridging Atlantic Ocean High (Dyson and van Heerden, 2002). Cold front systems approach from the west and the tropical systems from the east. Cold frontal systems are an important source of rainfall in the coastal areas, but they can penetrate into the interior and bring unseasonal heavy rainfall (Vogel, 2000). This is a result of the meeting of two different air masses, warm moist air and colder drier air, creating a zone characterised by changing pressures, wind and temperatures, and have a north-south orientation (Vogel, 2000). Cold fronts are prominent in winter over the subtropical region of South Africa and may reach as far north as 15°S, where they are associated with upper air westerly troughs or cut off lows (COLs). Cold fronts are normally followed by high pressure systems originating over the Atlantic Ocean. Atlantic Ocean highs, and the position and strength of these highs, regularly determine the severity of a cold front.

In the summer, the ITCZ moves southwards to approximately 17°S (Taljaard, 1994), bringing tropical weather to South Africa's northern regions. At this time of the year, tropical weather systems invade southern Africa in the form of tropical cyclones, tropical lows and easterly waves (Dyson and van Heerden, 2002). Tropical cyclones originate in the Indian Ocean and move in a south-westerly (east to west) direction over the subcontinent, rarely reaching the African west coast, and they bring heavy rain, flooding and powerful winds (Vogel, 2000; Dyson and van Heerden, 2002). Tropical cyclones degenerate to form tropical lows and they favour the large river valleys in their westward migration. (Dyson and van Heerden, 2002). Tropical cyclones seldom penetrate south of 30°S, but they occasionally affect the northern KwaZulu-Natal coast, for example Cyclone Demonia, which caused extensive flooding in northern KwaZulu-Natal (Vogel, 2000). Approximately 43 cold front events occur annually over KwaZulu-Natal (an average of 45 years of data) (Grab and Simpson, 2000).

Mid-latitude weather systems, bringing rainfall to the Cape south of the South African coast, include cold fronts, west-wind troughs, cut-off lows (COLs) and ridging high-pressure systems (Favre et al., 2012; Weldon and Reason, 2014). When rainfall over the region is of a tropical nature, it is usually from Tropical Temperate Trough (TTT) cloud bands that extend over both continental South Africa and the adjacent southwest Indian Ocean (Todd et al., 2004). This weather system is responsible for the bulk of rainfall over the southern African region, casuing about 39% of the rainfall, on average, over the summer rainfall region (Todd et al., 2004; Crimp et al., 1997). Jury and Levey (1993) suggested that COLs are responsible for the autumn rainfall peak, with ridging high-pressure systems driving the spring rainfall peak over the all-year rainfall region of the

Cape. These systems can be extremely hazardous, producing floods and causing consequent damage to infrastructure and sometimes loss of life (Weldon and Reason, 2014).

COLs have also been associated with numerous extreme rainfall events along the Cape south coast (Singleton and Reason, 2007a), causing the 1968 Port Elizabeth floods, where 500 mm fell within 24 hours (Engelbrecht et al., 2015). COLs have caused heavy rainfall events elsewhere as well, for example in September 1987, when the three-day rainfall total along the KwaZulu-Natal coast exceeded 900 mm (Singleton and Reason, 2007b). Heavy rainfall and flooding are not rare in South Africa. Alexander and van Heerden (1991) list 184 noteworthy flood events during the years from 1911 to 1988. Close to 30% of these events occurred over KwaZulu-Natal and 42% over the four northern and central provinces of South Africa (Alexander and van Heerden, 1991). Some 67% of all these flood events occur during the summer rainfall season (October to March) (Engelbrecht et al., 2015).

2.3 South African Rainfall

The ocean influences the progression and strength of the abovementioned weather systems that drive rainfall, thus the distribution of rainfall over South Africa can be related to these systems (Walker, 1990; Vogel, 2000). Over the interior of South Africa, there is a distinct east-west trend in rainfall (Partridge et al., 1997). More rainfall is experienced in the east, as the north-easterly airstreams affect the eastern Highveld and bring annual rainfall totals of around 800 mm, which is concentrated in the summer months (D'Abreton and Tyson, 1994; Vogel, 2000). Rainfall totals decrease towards the west, bringing arid desert conditions to the Kalahari and southern Namibia (Vogel, 2000).

Situated in the subtropics, South Africa mainly receives rainfall during the austral summer, except in the southwest region, which experiences austral winter rainfall, and the south coast of the country, which receives rainfall throughout the seaosns (Grimm and Reason, 2015; Philippon et al., 2011). The seasonality of rainfall along the South African coast varies from west to east. The Cape Peninsula has a strong winter rainfall, the south coast has all-year rainfall and, from the Eastern Cape to KwaZulu-Natal, the rainfall tends towards a summer rainfall maximum (Vogel, 2000). Towards the east, the winter months (June and July) tend to be consistently dry, but heavy precipitation in the early spring and late autumn is not uncommon (Vogel, 2000). Rainfall in the interior occurs during the austral summer months and is most strongly influenced by the tropical northerly airflow, which is enhanced by periodic coupling, together with temperate troughs (D'Abreton and Tyson, 1994). During the warm summer months, intense thunderstorms occur in the Highveld that sometimes bring hail and cause flash flooding (Vogel, 2000).

The winter rain-producing weather systems over South Africa are cold frontal troughs, low pressure systems close to land, cut-off lows (COLs) and long wave troughs (Stander et al., 2016). Heavy precipitation in winter occurs almost exclusively from westerly wind troughs and COLs and nearly all winter rainfall in summer rainfall areas is caused by COLs (Karoly and Vincent, 1998). The temperatures over the interior of South Africa are linked to the high-pressure system over the region. Winter weather in the interior is normally sunny and cloudless, with frost occurring in the early mornings, if the temperatures drop during cloudless nights. Interruptions in the conditions are usually the result of passing cold fronts, or the influx of warm moist air that may result in thunderstorms (Vogel, 2000).

South African rainfall experiences strong interannual fluctuations, with recurrent wet and dry periods, in response to the coupled ocean-atmosphere modes of variability over the Pacific, Indian and Atlantic Oceans (Dieppois et al., 2016). The relationship between the observed precipitation patterns and the associated climate drivers varies at different spatial and temporal scales (Gaughan and Waylen, 2012). Both seasonal and annual high variability characterizes the patterns of rainfall in southern Africa (Eriksen et al., 2008). The rainfall patterns are influenced by large-scale intra-seasonal and interannual climate variability, including El Nino Southern Oscillation events in the tropical Pacific and resulting in frequent extreme weather events, such as droughts and floods (Dore, 2005). A number of studies have confirmed a relationship between rainfall and ENSO in parts of eastern and southern Africa (Nicholson, 1996; Engelbrecht et al., 2015).

2.3.1 Southern and South African rainfall teleconnections

Understanding how the possible climate regime changes may influence future climate variability is critical in Africa (Parry, 2007). Advances have been made in our understanding of the complex mechanisms responsible for rainfall variability (Reason et al., 2005; Warren et al., 2006; Washington and Preston, 2006; Christensen et al., 2007).

Intra-annual and interannual precipitation variability across the African continent is strongly influenced by global climate forcing (Nicholson, 2000; Jury et al., 2004). Recent research has focused on the global scale forcing mechanisms of rainfall and circulation changes across the southern hemisphere.

Two underlying climate oscillations, impacting the timing and amount of precipitation in southern Africa across multiple temporal scales, are the Southern Oscillation and the role of sea-surface temperatures (SSTs), particularly in the South Atlantic and South Indian Oceans (Gaughan et al., 2016). The Southern Oscillation is of particular interest, because it is related to the more familiar El Nino phenomenon that has been used to explain short-term changes to the world's weather patterns in the form of regional drought and flooding. The Southern Oscillation comprises of El Nino and La Nina events that involve the large-scale warming or cooling of the equatorial sea-surface temperatures, together with an associated oscillation of atmospheric pressure over the South Pacific Ocean (Wang et al., 2017). Links between the Southern Oscillation and southern Africa's rainfall have been established in such a way that warm events (El Nino) are commonly associated with below-average summer rainfall over much of South Africa and cold events (La Nina) are typified by above-average rainfall in this region (MacKellar et al., 2014).

Climate exhibits numerous modes of variability in global and hemispheric circulation patterns at intraseasonal and interannual (year-to-year) timescales (MacKeller et al., 2014). The El Nino Southern Oscillation is recognized as the leading mode of interannual variability in the tropics and is driven by variations in sea-surface temperatures (SSTs) in the equatorial Pacific Ocean. The El Niño-Southern Oscillation (ENSO) describes the natural year-to-year variations in the ocean and atmosphere in the tropical Pacific that lead to large-scale changes in sea-level pressures, sea-surface temperatures, precipitation and winds, not only in the tropics, but also across many other regions of the world (Ratman, 2014).

An integral part of the Southern Oscillation is the role that sea-surface temperatures play in modulating the occurrence of low and high phase changes and the associated atmospheric circulation interaction. Temperatures of the oceans around South Africa exert an influence on rainfall over the subcontinent (Rouault et al., 2003). Variations in the summer precipitation over the southern African landmass have been attributed to the anomalous changes in the surrounding oceans (Ratnam et al., 2014). For example, variations in the Pacific Ocean SST, due to El Nino Southern Oscillation (ENSO), are known to exert a remote forcing to cause the variations of the precipitation over this region (Rouault and Richard, 2003; Jury et al., 2004). Rainfall over southern Africa has been shown to be anomalously high when sea-surface temperatures are above normal (Landman and Beraki, 2012). Sea-surface temperatures in the northern Indian Ocean are associated with decreased rainfall over southern Africa in the summer (Jury, 1996). Interannual anomalies of SST thus have a significant influence on the climate over land, and these relationships play an important role in seasonal climate forecasting (Palmer and Anderson, 1994) and in the explanation of droughts (Seager and Hoerling, 2014).

Dry years over southern Africa are often associated with El Nino events in the Pacific (Jury et al., 2004; Mason and Jury, 2007). The incidence of dry years associated with El Niño events has been increasing in the last few decades (Gaughan and Waylen, 2012), although the strength of this association varies geographically and is complicated by other potential interacting climate forcings (Zheng et al., 2010; Manatsa et al., 2011). During many, but not all, El Niño events, precipitation anomalies are negative over southern Africa, south of 15°S (Cook, 2001). It has been shown that severe summer droughts over much of South Africa tends to occur under El Nino conditions (Lindesay, 1988; Reason and Rouault, 2002) and this relationship seems to have strengthened since the 1970s (Richard et al., 2000). Over South Africa, Mason and Jury (1997) found that the influence of ENSO events on rainfall is strongest during summer peak rainfall months December–March, when ENSO events typically have reached maturity and when the tropical atmospheric circulation is usually dominant.

Further, Landman and Beraki (2012) learned that seasonal prediction of summer rainfall in South Africa is prominent during strong ENSO phases. Weldon and Reason (2014) show that rainfall in the South Coast region of South Africa, a region that experiences substantial rainfall variability and frequent severe drought and flood events, is influenced by ENSO. El Nino events were found to bring less rain and most wet years correspond to the mature phase La Niña years (Weldon and Reason, 2014). ENSO also influences South Coast rainfall via increases in the number of cut-off lows in southern South Africa, during the mature phase La Niña years. It must be noted, however, that the relationship between ENSO and South Africa's summer rainfall is very complex, as there are a lot of factors that influence the region’s climate (MacKeller et al., 2014). For example, the synoptic-scale patterns of convection over South Africa are also modulated by ENSO, but different synoptic responses under the same ENSO phase can result in very different rainfall patterns (Fauchereau et al., 2008).

El Niño episodes have weakened since the late 1970s, resulting in important changes in the association between tropical western Indian Ocean sea-surface temperatures and December–February rainfall over South Africa (Landman and Mason, 1999). The sea-surface temperature variability in the equatorial Indian Ocean has been characterized by increasing temperatures. Subsequent to this warming, anomalously warm conditions in the tropical western Indian Ocean have sometimes been associated with wet conditions over parts of southern Africa (Landman and Mason, 1999). The abrupt warming of the tropical Pacific and Indian oceans in the late 1970s is suggested to be partly responsible for increasing the air temperatures over southern Africa, and may have contributed to the prolonged dry conditions (Mason, 2001). The 1982/83 ENSO event exacerbated the prevailing dry conditions over much of the subcontinent (Dent et al., 1987). A return to a wet phase appears to have occurred, despite the persistence of the anomalously high sea-surface temperatures associated with warming of the the late 1970s, and a severe El Niño in 1997/1998 that hardly had any impact (Mason, 2001). The abrupt warming of tropical sea-surface temperatures has been attributed to the enhanced-greenhouse effect, but it may also be indicative of inter-decadal variability (Mason, 2001). These changes in the rainfall–sea-surface temperature in the climate system have important implications for the predictability of southern African rainfall (Landman and Mason, 1999).

The incidences of dry years, associated with El Niño events, has been increasing in recent decades (Gaughan and Waylen, 2012), although the strength of this association varies geographically and can involve other potential interacting climate forcings (Zheng et al., 2010; Manatsa et al., 2011). Subject to the timeline of the research in this study (1955 – 2015), there have been 18 ENSO events in the last 61 years (Climate Protection Centre, 2017). El Nino conditions have occurred in 1982/1983, 1986/1887, 1991/1992 1994/1995, 1997/1998, 2002/2003, 2004/2005, 2006, 2009/2010 and 2015/2016. The 2015/2016 ENSO event is recorded as being the worst in recorded history (Climate Protection Centre, 2017).

ENSO events and sea-surface temperature anomalies in the Indian and South Atlantic Oceans can influence both the tropical and temperate atmospheric circulation and moisture fluxes over the subcontinent and are thus significant influences on rainfall variability (Ratnam et al., 2014). The possibility of changes in the frequency and intensity of ENSO events, and of their predictability, are of direct concern in the southern African region. Because ENSO events are generally associated with significant rainfall anomalies over most of southern Africa (Mason and Jury, 1997), long-term trends in ENSO variability are likely to affect the rainfall climatology of the region. In addition, changes in ENSO predictability will affect a region's ability to mitigate the effects of climate variability (Mason, 2001). In summary, rainfall variations over the country are influenced by a number of interactive mechanisms that are the subject of continued research. The influence of teleconnections and global-scale circulation forcing mechanisms is emerging as one of the key determining factors of local climates. Much of the recent work points to an increase in interannual rainfall variability, particularly of extreme events.

2.4 Impact of Rainfall Variability on Water Resources in Southern Africa

The southern African region is regarded as one of the most vulnerable regions in Africa (IPCC, 2007a). The latest IPCC (2014) report highlighted that southern Africa as one of the areas that will experience changes because many of the countries have scarce water resources and a highly spatial and temporally variable rainfall. It is thus projected that southern Africa, as a whole, could face challenges related to water availability and stress due to climate change, including agricultural productivity and energy use, drought and flood control and municipal and industrial water supply (IPCC, 2014). Its vulnerability is exacerbated by the region's low adaptive capacity, widespread poverty and low technology uptake (Kusangaya et al., 2014). Such impacts filter down to a smaller scale within the specific regions of these countries, and it is at this level that the impacts are most strongly experienced (IPCC, 2014).

Among the many adverse impacts of climate change, the risk to agriculture is considered to be highly significant (Seo et al., 2009), as the majority of the world's population, especially those in the developing countries, depend on agriculture for their livelihoods (World Bank, 2007). Agriculture is very sensitive to weather and climate variables, including temperature, precipitation, light and weather extremes, such as droughts, floods and severe storms (Molua, 2002). A number of countries in Africa already face semi-arid conditions and the impacts of climate change, for example, the reduced length of the growing season, which makes it difficult to establish agricultural practices, thereby worsening likelihood of success (Boko et al., 2007). Yields in some countries are projected to reduce by as much as 50% by 2020 and net crop revenues could fall as much as 90% by 2100, with small-scale farmers bearing most of the consequences (Boko et al., 2007). The actual crop yield, as a percentage of the potential yield, have already decreased by 40% for North Africa and <30% for sub-Saharan Africa (FAO, 2012).

Sub-Saharan Africa has been portrayed as being the most vulnerable region to the impacts of global climate change because of its reliance on agriculture (Kotir, 2011). For example, Batisani and Yarnal (2010) found that the decreasing rainfall trend across Botswana was associated with decreases in the number of rainy days. Both the drying trend and the decrease in rainy days agree with climate change projections for southern Africa, suggesting that climate change is occurring in Botswana (Bastsani and Yarnal, 2010). These results have important policy implications for the government, which must help its dryland farmers to adapt to the changing climate. Its impacts are projected to worsen in the future, as the temperature continues to rise and as precipitation becomes more unpredictable (Batisani and Yarnal, 2010). Therefore, Africa needs to specifically focus its attention on agricultural water management (Valipour, 2015).

Within the climate change matrix, water resources are at the centre of projected climate change impacts. If the observed changes in climate in the last century persist into the future, their impact on water resources is likely to increase in extent, variety and severity (IPCC, 2007a,b). Given the already large spatial and temporal variability of climatic factors in southern Africa (Gallego-Ayala and Juízo, 2011), the impact of climate change on water resources is likely to be more prominent in the near future than previously experienced (Kusangaya et al., 2014). The water sector is strongly influenced by, and sensitive to, changes in climate , including periods of prolonged climate variability (Boko et al., 2007). Several studies have shown that climate change will impact on the availability and demand for water resources. The water demand in South Africa showed an annual growth rate of 1.5% between 1990 and 2010, with 3.5% predicted for urban and industrial use and 1% for irrigation (Otieno and Ochieng, 2004).

2.4.1 Impacts on hydrological response drivers in southern Africa

Several studies in the region have considered the potential impact of climate change on water resources, with a general consensus that, as mentioned earlier, it will affect both the quality and quantity of available water resources in the region (Schulze, 2000, 2005). The following section reviews the impact of climate change on the key climatic drivers of hydrological responses in southern Africa, namely, temperature and rainfall.

Rainfall trends have been extensively studied in assessing the impact of climate change on water resources. The majority of studies concur that, even without climate change, rainfall in southern Africa is extremely variable in time and space (Kusangaya et al., 2014). The extent of the effect of climate change on rainfall in southern Africa is uncertain; however, a decrease in rainfall as a result of climate change is suggested in the IPCC (2007b) report. Climate change models have predicted that by 2050 the interior of southern Africa will have decreased rainfall during the growing season due to reductions in soil moisture and runoff (Kusangaya et al., 2014). For example, Hulme (1992) predicted a 5-10% reduction in rainfall and the IPCC (2001) predicted a 5 to 15% decrease of growing season rainfall in southern Africa. Mazvimavi (2011) projected a 3-23% decrease in rainfall under climate change in southern Africa. Several scholars share the same view (e.g. Hulme et al., 2001; Arnell et al., 2003) and have predicted that the region will be characterized by below-normal rainfall and frequent droughts in the future.

Thus, increased interannual variability, resulting in extremely wet periods and more intense droughts, will continue to occur (Kusangaya et al., 2014). Rainfall patterns are also expected to change in intensity and frequency, resulting in more extreme events and longer periods between rainfalls (Christensen et al., 2007). At the same time, the demand for water in the region is increasing rapidly, due to population growth and economic development (AWDR, 2006). Most livelihoods in the region are still dependent rain-fed agriculture, which will be affected by the reduction in rainfall casued by climate change (Thornton et al., 2009; 2010).

Changes in temperature and rainfall have a direct effect on the quantity of evapotranspiration and on both the quality and quantity of the runoff. Temperature changes can lead to changing patterns of rainfall, the distribution of runoff, soil moisture and groundwater, as well as an increase in the frequency and occurrence of floods and droughts (Schulze, 2011). Consequently, the spatial and temporal availability of water resources can be significantly altered, with any changes in temperature (Kusangaya et al., 2014).

Temperatures over southern Africa have not been constant over the years. Hughes and Balling (1996) found evidence of 1°C warming over South Africa and 0.5°C over southern Africa from 1960-1990. Easterling (1997) found an increase in the annual mean daily maximum temperatures and widespread increases (although some decreases) in annual mean daily temperatures in South Africa between 1950 and 1993. Easterling (1997) also found an increase in the diurnal temperatures (difference between minimum and maximum temperatures) in much of South Africa. In contrast, Hulme et al. (2001) found an opposing trend of decreased diurnal temperatures over South Africa during the 1950s and 1960s. Kruger and Shongwe (2004) discovered that, with few exceptions, stations in South Africa have reported increases in annual mean temperatures for the period 1960-2003 (0.1-0.3 °C), with the strongest warming occurring in the interior, in the autumn months. New et al. (2006) identified that there was an increase in the number of warm spells over southern and western Africa, and a decrease in the number of extremely cold days over this period, over a similar period (1961-2000). Kruger and Sekele (2013) found an overall increase in hot extremes and a decrease in cold extremes over the period 1962-2009, with the strongest changes tending to occur in the western and northern interior of South Africa.

Most studies over southern Africa project a temperature increase of around 3°C by the 21st mid-century. For example, Mujere and Mazvimavi (2012) projected a 3°C maximum temperature increase up to the year 2050 in the Mazoe Catchment in Zimbabwe, Beck and Bernauer’s (2011) projection to the year 2050 showed a 2.9°C maximum increase for the Zambezi Basin, Graham et al. (2011) project a 3°C maximum temperature increase for the Tugela Catchment, in South Africa and Hewitson and Tadross (2011) predict a 3°C maximum temperature increase for South Africa, Swaziland and Lesotho.

There has been a strong warming trend across the African continent over the past 50 to 100 years, with the mean annual temperature rise over Africa likely to exceed 2°C by the end of the 21st century (IPCC, 2014). The modelling of future climate scenarios in southern Africa has predicted an increase in dryness, involving an increase in hot days and a decrease in cool days (Hewitson et al., 2014). This warming will be greatest over the interior and semi-arid margins of southern Africa, the Sahel and central Africa. Projections show that temperature changes will not be uniform over the region; the central, southern land mass extending over Botswana, parts of north-western South Africa, Namibia and Zimbabwe are likely to experience the greatest warming of 0.2-0.5°C per decade (Christensen et al., 2007)

River runoff and water availability are projected to decrease by 10 to 30% in the dry tropics (IPCC, 2007b). Arnell (1999) predicted a reduction in runoff of 26 to 40% in the Zambezi river system, as a result of reduced rainfall and increased evaporation. Evaporative increases of 40%, for example, could result in reduced outflows from reservoirs. In additionally, the projected increased frequency of droughts will most likely increase the frequency of low storage episodes (Desanker and Magadza, 2001), which will inevitably affect, for example, future hydropower generation from such dams as the Kariba and Cabora Bassa (Yamba et al., 2011). The general conclusion from most studies is that streamflow is projected to decrease by 2050. For example, Matondo (2012) projected that the streamflow (for Swaziland) will decrease by up to 40%, Beck and Bernauer (2011) projected a decrease of up to 20% for the Zambezi Catchment, Zhu and Ringler (2010) projected an up to a 35% decrease for the Limpopo catchment, and Graham et al. (2011), projected an up to 18% decrease for Tugela Catchment, in South Africa.

2.5 Rainfall and Rainfall Variability in Southern and South Africa

Southern Africa is a predominantly dry region, with rainfall highly erratic in both time and space (Vogel, 2000). Sixty-five percent of South Africa is a semi-arid and the average rainfall of 600 mm year is below the world average of 860 mm (Otieno and Ochieng, 2004). As a result, South Africa has been classified as a water-scarce country (RSA, 2002). It has been forecasted that it will experience physical water scarcity by the year 2025, with an annual freshwater availability of less that 1 000 m^3 per capita (Otieno and Ochieng, 2004). The country is currently categorised as a water-stressed, with an annual fresh water availability of less than 1 700 m^3 per capita (the index for water stress). The natural availability of water across the country is also uneven and this is compounded by the strong seasonality of rainfall (Otieno and Ochieng, 2004). Notably, an east-west trend in rainfall distribution exists across the country. Rainfall increases towards the east and, in particular, the KwaZulu-Natal Drakensberg catchments are known to produce nearly twice as much total runoff per unit rainfall than the average for South Africa as a whole (Schulze, 1979), and a quarter of South Africa's streamflow (Whitemore, 1970). South Africa is therefore characterised by complex topographical features and marked gradients in vegetation and land cover, and receives most of its rainfall during the austral summer season (December–January–February) (Ratna et al., 2014). Systems, such as tropical temperature troughs, westerly troughs and cut off low-pressure systems and thunderstorms dominate (Ratna et al., 2014).

South Africa is located in the dry subtropics of the southern hemisphere, indicating that large spatio-temporal variability of rainfall is controlled by both tropical and mid-latitude dynamics (Ratna et al., 2014). Over the past decades, precipitation regimes have become more variable, with longer dry periods and an increase in the number of extreme rainfall events (Koerner and Collins, 2014*)*. This variability also has important consequences in terms of extreme events, such as floods and droughts. Much climatological research has focused on the short- and long-term variability of rainfall, recently extending its scope to the occurrence of extreme events, such as droughts. As rainfall is of upmost importance to society, to the economy (agriculture, in particular) and to water resource management/planning, the understanding and prediction of rainfall variability is regarded as a priority (Ratna et al., 2014). Active research on rainfall variability is ongoing in the country and the region. Studies reviewing the rainfall patterns in southern and South Africa are discussed below.

2.5.1 Review of climate trend studies in southern Africa

During the 20th century, the climate of subtropical southern Africa was characterised by a high degree of temporal and spatial variation (Preston-Whyte and Tyson, 1988; Richard et al., 2001). Much of the variability was random, however, with the record, real and significant non-random components being clearly identifiable. Quasi-periodicities in annual rainfall totals over southern Africa have been identified; in particular, an approximately 18-year cycle has been identifed that affects the summer rainfall region of north-eastern South Africa, to the greatest extent, and that may be related to inter-decadal variability in sea-surface temperatures in the eastern equatorial Pacific and central Indian Oceans (Mason and Jury, 1997). In addition to these oscillations, a broad band of variability of the rainfall, of three to six years, occurs throughout southern Africa and appears to be associated with the El Nino Southern Oscillation phenomenon (Preston-Whyte and Tyson, 1988). Several wet and dry years have been identified from meteorological records in the last century. There is a distinct variability in the spatial distribution of wetter and drier conditions that change from year to year; drier spells, however, tend to show a greater areal extent and spatial homogeneity than wet spells (Preston-Whyte and Tyson, 1988) (Figure 2.1).

Rainfall variability and changes in southern Africa over the 20th century were investigated by Richard et al. (2001) and no significant changes in the summer rainfall totals were identified, although, since the beginning of the 1980s, the annual rainfall total had decreased slighty. Thus, summer rainfall showed a change in the intensity of interannual variability (Richard et al., 2001). Fauchereau et al. (2003) found that cumulative rainfall anomalies over the summer season also did not show any trend towards drier or moister conditions during the last century. However, rainfall variability in southern Africa has experienced significant modifications; interannual variability has increased since the late 1960s and, in particular, droughts became more widespread and intense (Fauchereau et al., 2003). A number of studies over the years have identified several notable patterns of rainfall, in particular, the distinctive oscillations over southern Africa and South Africa, both temporally and spatially (Tyson et al, 1975; Ngara et al., 1983; Tyson, 1986; Mason and Jury, 1997; Nicolson, 2000; Rouault and Richard, 2003; Nel and Sumner, 2006; Nel, 2009; Kusangaya et al., 2014).

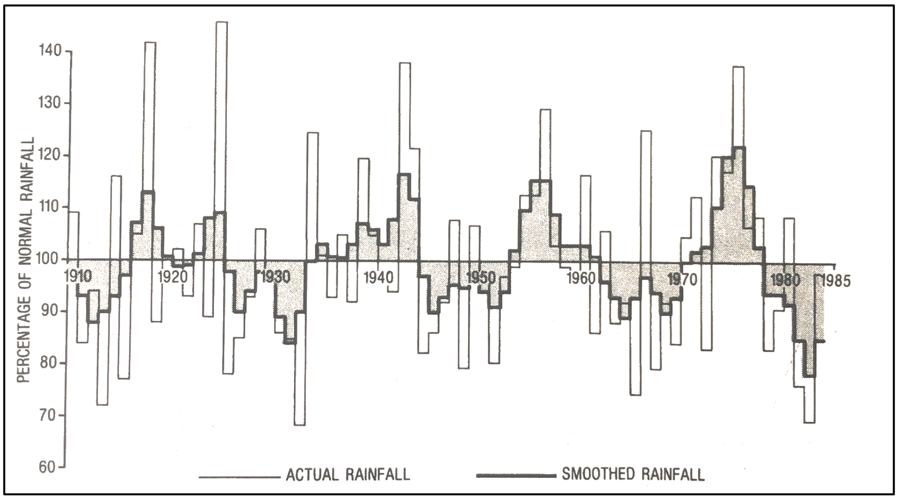


Figure 2.1 Spatially averaged rainfall series for October – September rainfall in the summer rainfall region. (Sourced from: Preston-Whyte and Tyson, 1988).

Other studies have looked at rainfall changes for the southern African region. Ngongondo et al. (2011) found that most of the rainfall stations used in the study in Malawi revealed statistically non-significant decreasing rainfall trends for annual, seasonal, monthly and for the individual months from March to December, at the 95th confidence level. A study in Zambia by Sichingabula (1998) showed increasing 11-year coefficients of variation (CVs) for selected stations and decreasing rainfall trends were observed in southern Zambia after 1975. Mazvimavi (2010) concluded that rainfall in Zimbabwe has a high inter-annual variability. Shongwe et al. (2009) concluded that rainfall trends are characterised by more severe droughts in the southwest of southern Africa, with increased rainfall in the north (Zambia, Malawi, and northern Mozambique). Prior to this, Chamaillé-Jammes et al. (2007) and Joubert et al. (1996) concluded that rainfall in southern Africa showed no consistent or statistically significant trends across the region, however, a decrease in non-statistically significant regionally-averaged total rainfall was found.

Nicholson (2000; 2001) found that in southern Africa there was a shift from the relatively wet conditions of the 1920-1950s, to dry conditions, from the 1970s onwards. Nicholson (2001) showed that there was a 6% increase in rainfall in the 1970s, followed by a reduction of 5% in the 1980. In contrast, an analysis of observed rainfall trends for Zimbabwe (1933–2000) showed that there was no statistically significant rainfall reduction in Zimbabwe (Mazvimavi, 2010). This is in contrast to earlier results from Unganai (1996), who concluded that the areal annual rainfall in Zimbabwe had declined by 10% between 1900 and 1994. Shorter and weaker rain seasons in southern Africa were identified by Morishima and Akasaka (2010), who analysed rainfall data for the 1979-2007 period for southern Africa and concluded that the annual rainfall has decreased over the African continent equator to 20°S, including Madagascar. New et al. (2006) reported that, in the last decades of the 20th century, a spatially coherent increase the in consecutive dry days occurred over much of southern Africa.

These studies and reports suggest that the annual rainfall has not had a clear tendency in the last 20 or 30 years, but most authors concur that the dry periods have increased in duration and intensity in southern Africa (Kusangaya et al., 2014). Changes in rainfall patterns are subject to considerable uncertainty, with regard to the spatial and temporal extent and magnitude of the change (Mazvimavi, 2011). This is largely because rainfall is characterised by high inter-annual variability over southern Africa.

2.5.2 Review of climate trends in South Africa

Numerous studies have investigated climatic trends in South Africa, and they have mostly focused on the temperature and rainfall figures from recording stations. The greatest restriction to such historical studies is the lack of representation of a region's climate (Mackeller et al., 2014). South Africa has a relatively good network of rainfall and temperature recording stations, in comparison to the rest of Africa (Mackeller et al., 2014). It is, however, difficult to detect clear signs of long-term change, given large variability across a range of spatial and temporal scales. This is particularly difficult with regards to rainfall, which is highly variable in both space and time (Mackeller et al., 2014). Because South Africa has a variable mean annual precipitation, few spatially coherent or statistically significant trends in this quantity have been observed (Kruger, 2006; Nel, 2009). The characteristics of how rainfall is distributed throughout the year is important, as these characteristics include the onset and end of the rainy season, the duration of wet and dry periods and the occurrence of extreme rainfall events (Mackeller et al., 2014).

A substantial amount of research in Africa has focused on identifying patterns in rainfall over time and, more particularly, attention has been on rainfall cycles at continental, regional and local scales (Nicholson, 2000; Nel, 2009). A range of short period cycles, in the order of 2-7 years, have been identified in South Africa (Vines, 1980; Jury and Levey, 1993; Kane, 2009), although greater interest has been placed on 20-year cycles (Dyer and Tyson, 1977; Vines, 1980; Jury and Levey, 1993; Kane, 2009; du Toit and O'Connor, 2014). Tyson et al. (1975) showed that a 10-year to 12-year oscillation predominantly affected the southern coastal region and the adjacent inland area. Vines (1980) examined data in the southern and north-eastern parts of South Africa and reported periodicities of 6–7, 10–12 and 16–20 years. Identifying longer cycle periods is often difficult, because data series are limited in length (usually <100 years). More recently, strong evidence of an approximately 20-year rainfall cycle, except for spring rain, was found for stations in the Karoo, over a 123-year period (du Toit and O'Connor, 2014). Analyses have found that rainfall patterns in the area are not random, but are rather driven by cyclical processes and that rainfall is predicted to decrease over the next 20 years. However, it is thought that high variability and complex causal factors will make it difficult to differentiate between natural variation and the possible effects of climate change (du Toit and O'Connor, 2014).

In a preliminary study of daily rainfall time series from the SAWS (South African Weather Service) over the whole of South Africa for the 1920-2000 period, shifts in the daily rainfall distribution were found from around the 1970s (Richard et al., 2001). Increased intense rainfall, accompanied by decreased number of rain days, was found in the south-central part of the country, while the north-eastern part showed a significant decrease in rainfall (Richard et al., 2001). Smakhtina (1998) also showed that the Eastern Cape province of South Africa experienced an increase in heavy rainfall events, with a reduction in the number of rain-days since the beginning of the century.

Recent work by Kruger and Nxumalo (2017) indicates an increase in rainfall over time in South Africa from 1921-2015, particularly in the southern interior, as well as signs of a decrease in rainfall in the far north-eastern parts of the country. This increase in rainfall in the south is reflected in the seasonal trends, where summer rainfall showed an increase, but for the other seasons, most of the country shows no significant historical trends in the annual total rainfall (Kruger and Nxumalo, 2017). Decreases in rainfall from wet spells were found over the eastern and north-eastern regions of the country, while the southern and eastern parts along the escarpment had shorter annual dry spells (Kruger and Nxumalo, 2017). This decreasing rainfall in the east is more pronounced in the work of MacKellar et al. (2014) who found statistically significant decreases in rainfall and the number of rain days over the central and north-eastern parts of the country in the autumn months and significant increases in the number of rain days around the southern Drakensberg are evident in spring and summer. This could suggest that decreases in this particular area have become prominent in recent decades, especially in autumn.

Several studies in South Africa have shown that rainfall in South Africa is characterised by high inter-annual variability. A review by Easterling et al. (2000) noted that extreme rainfall often occurred in the south-western and eastern regions of South Africa in the 20th century. Groisman et al. (2005) also found significant increases in the yearly frequency of heavy rainfall events of eastern South Africa, from 1906-1997. Mason et al. (1999) found an increase in the intensity of extreme rainfall events in the 1961-1990 period relative to the prior 30 years (1931-1960) over much of South Africa. In the largest parts of South Africa, Kruger (2006) found that there has been no real evidence of changes in rainfall, over the past century (1910-2004). There are, however, some identifiable areas (Free State and Eastern Cape) where significant changes in certain characteristics of rainfall have occurred. An increased dry spell duration and a decreased wet spell duration have been observed for parts of the Eastern Cape and the north-eastern parts of South Africa, over the period 1910–2004. New et al. (2006) also showed some evidence of increased rainfall extremes over parts of South Africa for the 1961-2000 period. Mackellar et al. (2014) found statistically significant decreases in rainfall and the number of rain days for the central and north-eastern parts of South Africa in the autumn months and significant increases in the number of rain days around the southern Drakensberg in the spring and summer.

In the KwaZulu-Natal Drakensberg, Nel (2009) demonstrated a shift in seasonality during the period 1955-2000. The mean annual rainfall showed no significant trend, but an increase in summer rainfall, with a decrease in autumn and winter rainfall, resulted in a shorter wet season and increased seasonality. Thomas et al. (2007) also found an increase in early season rainfall in northwestern KwaZulu-Natal, along with a decrease in late-season rainfall between 1950 and 2000. Seasonal shifts were observed in Limpopo for the same period, where there has been a tendency for a later seasonal rainfall onset, accompanied by increased dry spells and fewer rain days. An increased dry spell duration is also evident in much of the Free State and the Eastern Cape, and decreases in wet spell duration have been observed for parts of the Eastern Cape and north-eastern parts of South Africa between the 1910-2004 (Kruger, 2006). Hewitson and Crane (2006) reported rainfall increases in regions where orography is a main contributing factor, and they also found increases in late summer dry spell duration for much of the summer rainfall region from 1950-1999.

More recently, the South African Weather Service (SAWS) has compiled a dataset of monthly average rainfall per province for all nine provinces dating back to 1904 (SAWS, 2016). This dataset was used to calculate the annual total rainfall (January–December) for South Africa over a 112-year period from 1904-2015 and the data showed the annual average rainfall for South Africa to be 608 mm (SAWS, 2016) (Figure 2.2). Over the 112-year period, the lowest annual total rainfall for South Africa was for the period January–December 2015, with an annual total of only 403 mm, thus 2015 was a very dry year in South Africa. There have been 13 years in which the annual total rainfall for South Africa was less than 500 mm (SAWS, 2016). The longest period of consecutive years, in which the annual total rainfall was below the period average of 608 mm per annum, was the six years, starting from 1944 to 1949. If an annual average is calculated for this six-year period, it is 544 mm (SAWS, 2016), highlighting the significantly below average rainfall of 2015.

Southern Africa is one of the regions that is most sensitive to precipitation shifts and variability (IPCC, 2007b) and where the future rainfall distribution, frequency and variation remain uncertain (Gordon et al., 2000). Thus, southern Africa is experiencing multi-decadal trends, which highlights the declining mean annual precipitation, the increasing variability and drier conditions, and an increased number of warm phase El Nino Southern Oscillation (ENSO) events (Gaughan and Waylen, 2012). According to Galvin et al. (2004), the probability of the occurrence of extreme events, such as droughts caused by rainfall variability, is predicted to increase in southern Africa (Galvin et al., 2004).

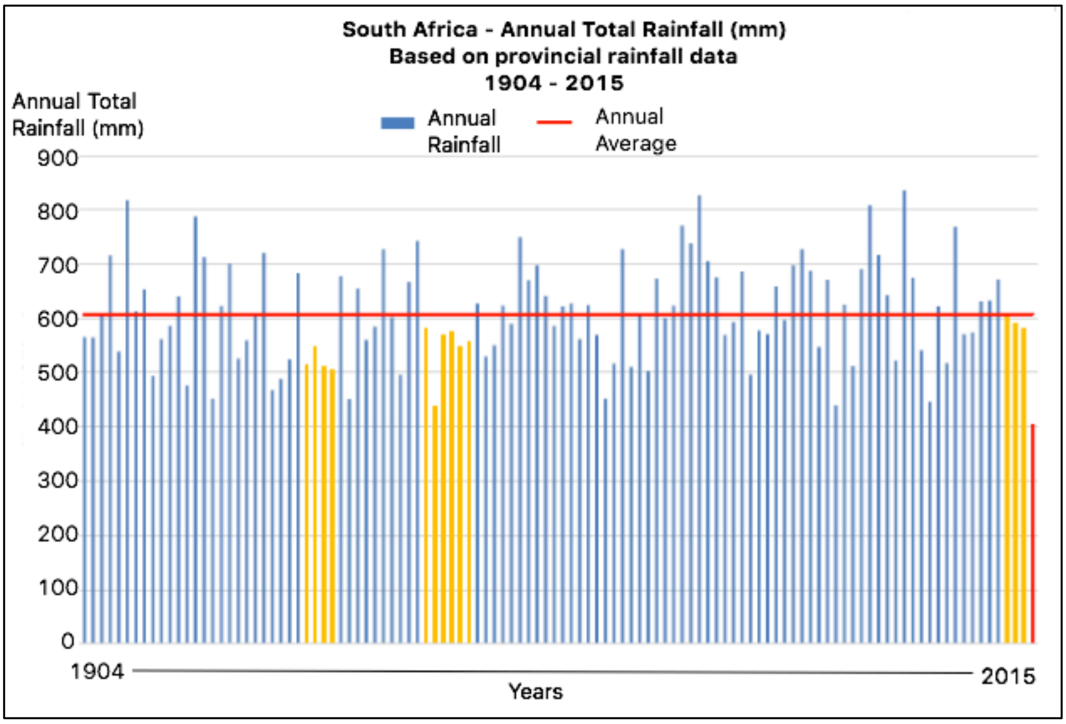


Figure 2.2 Annual total rainfall graph for South Africa from 1904-2015. (From SAWS, 2016: http://www.agrisa.co.za/sa-rainfall-in-2015-the-lowest-on-record-saws/).

2.6 Drought Characterization and Concepts

The global water cycle is expected to change in the 21st century, due to the combined effects of climate change and increasing human intervention (Prudhomme et al., 2014). In a warmer world, the water-holding capacity of the atmosphere will increase, resulting in a change in the frequency of precipitation extremes, increased evaporation and dry periods (Trenberth, 1999), as well as the intensification of droughts (Field, 2012). The rise in water demand and the looming climate change in recent years have resulted in an increased focus on global drought scenarios. Due to the growth of population and expansion of the agricultural, energy and industrial sectors, as well as climate change and the contamination of water supplies, the demand for water has increased greatly (Bates et al., 2008). Water scarcity has been occurring almost every year in many parts of the world, and it has been further compounded by droughts, which affect both the surface water and groundwater resources and which can lead to reduced water supply, deteriorated water quality, crop failure and disturbed riparian habitats (Riebsame et al., 1991). The significant spatial and temporal variability of water resources often results in water deficiencies in different regions and at different times (Zarch et al., 2015). Thus, droughts are highly variable in time and space.

Drought is a normal climate phenomenon; in almost every country, however, the economic, social and environmental impacts are serious and affect more people than any other natural hazard (Edossa et al., 2014) (see definition below). A recent review on droughts and aridity by Dai (2011) indicated that large-scale droughts have occurred frequently over the past 1000 years across the globe. An important finding highlighted that global aridity and drought areas have increased substantially during the 20th century and have contributed towards widespread drying over Africa, southern Europe, East and South Asia, eastern Australia and many parts of the northern mid-high latitudes, since the 1970s, (Dai, 2011). These aridity trends are projected to continue to increase in the 21st century. A study by Sheffield et al. (2012) showed that drought patterns have been increasing over the past 60 years, but not at an alarming rate as projected. Masih et al. (2014) found that droughts have become more frequent, intense and widespread over the past 50 years. The available estimates on drought impacts suggest that there were 642 drought events reported across the world, between 1900-2013, resulting in a huge toll to humanity, killing about 12 million people and affecting over 2 billion (EM-DAT, 2014). The total economic damage has been estimated at USD135 billion (Masih et al., 2014).

Droughts and floods are recurrent phenomena in Africa, due to the variability of rainfall. One-third of the people in Africa live in drought-prone areas and are vulnerable to the impacts of droughts (World Water Forum, 2000). Recent droughts in southern Africa have highlighted the importance of the interannual rainfall variability in these regions. Seymour and Desmet (2009) suggest that long-term drought research is essential in the country and underlined the importance of long-term field observations, experiments, and models for informing agricultural policy and conservation planning (Edossa et al., 2014). It is therefore pertinent to look at drought as a consequence of increased rainfall variability. In order to understand what drives drought and uncover the history of droughts and how to quantify them, relevant literature has been reviewed. The following section focuses on drought conceptualization, its characterization, indices, causes and the occurrence and impacts in southern and South Africa.

2.6.1 Definition of drought

There is no unified definition of drought (Dai, 2011). Drought is therefore a relative phenomenon and definitions must relate to deviations from the long-term mean or accepted condition (Massa et al., 2014). Differences in hydrometeorological and socioeconomic factors, as well as the variable nature of the water demand in different regions across the world, make it difficult to have a precise definition of drought (Mishra and Singh, 2010). Droughts do not have a clear onset, duration, or ending. In addition, drought threshold can vary significantly, depending on the season and region, with the same amount of precipitation having different implications in wet and arid regions (Prudhomme et al., 2014).

The definition of drought is in itself complex and when defining drought, it is important to differentiate between conceptual and operational definitions of drought (Wilhite and Glantz, 1985). Conceptual definitions are those stated in relative terms, namely, drought is a long, dry period (Mishra and Singh, 2011), and they are therefore generally formulated for an overall understanding, to establish drought policy (Wilhite and Glantz, 1985).

Operational definitions of drought objectively define the criteria for the duration and severity of a specific type of drought (Mishra and Singh, 2011).

Some commonly-used definitions include the following: (i) The World Meteorological Organization's (WMO, 1992) glossary provides a definition of drought as a "polonged absence or marked deficiency of precipitation" and a "period of abnormally dry weather that is sufficiently prolonged, for the lack of precipitation, to cause a serious hydrological imbalance." (ii) The UN Convention to Combat Drought and Desertification (UN Secretariat General, 1994) defines drought as a phenomenon that occurs naturally where precipitation has been significantly below normal recorded levels, causing serious hydrological imbalances that negatively affect land resource production; and (iii) Gonzalez and Valdes (2006) concur that drought is the extreme persistence of a precipitation deficit.

Ismail-Zadeh et al. (2014) state that the deficiency of rainfall, over an extended period of time, is a primary factor of all types of droughts. However, contrary to this statement, Lloyd-Hughes (2014) argues that to focus solely on precipitation is to neglect the importance of evaporation and transpiration as moisture sinks, which reduce the amount of water available for use. These definitions also ignore the importance of lateral inflows (stream and groundwater flows) that operate as another source of water in addition to the local precipitation (Lloyd-Hughes, 2014). Further, some of the definitions make no reference to the timing of the precipitation deficits, a factor which is crucial in the determination of many drought impacts (Lloyd-Hughes, 2014).

Lloyd-Hughes (2014) shows that a universal description of drought requires a specific reference to water supply, demand and management, and they stress that the influence of human intervention through water management is key to the definition of drought, in the universal sense, and can only be eliminated in the case of a purely meteorological drought. A better general definition is thus proposed by Gocic and Trajkovic (2014), who define drought as a prolonged period of water deficit that occurs typically when an area receives below average rainfall for several months, and the timing, distribution, and intensity of the deficiency must be viewed in relation to the existing water storage, demand and use. However, Sheffield and Wood (2012: 11) succeed in defining drought both accurately and succinctly as 'a deficit of water, relative to normal conditions'. It would thus seem more appropriate to use operational definitions that are based on the specific type of drought that one is trying to define.

2.6.2 Types of drought

Through using an operational definition of drought, three main physical drought types were established, namely: meteorological, agricultural and hydrological droughts (Zargar et al., 2011). In a broad definition, these droughts occur in a particular order. A precipitation deficiency instigates a meteorological drought, which subsequently causes a soil moisture deficit during the growing period (mostly related to rain-fed agriculture) inducing agricultural drought (Edossa et al., 2014). The low recharge from the soil to water features, such as rivers and dams, causes a delayed hydrological drought. Socioeconomic drought associates a drought with the impact that it has on society, both socially and economically (Edossa et al., 2014).

A meteorological drought is caused by abnormal meteorological conditions, such as the lack of rainfall and high temperature (Yang et al., 2015), and it is thus defined as a lack of precipitation over a region for a period of time (Mishra and Singh, 2010). Of all the recorded meteorological variables, the one with the longest time series over South Africa is rainfall. Rainfall data can therefore be used to evaluate aspects of drought (Malherbe et al., 2015). Several studies have analysed droughts by using monthly precipitation data, where drought is considered as precipitation deficit with respect to average rainfall values (Mishra and Singh, 2010). Agricultural drought refers to a period with a deficit of soil water, causing crop failure (Ramadas and Govindaraju, 2015). A decline of soil moisture depends on several factors, such as precipitation, soil moisture and temperature, which are the common variables adopted for agricultural drought studies (Mishra and Singh, 2010). The drought intensity is categorised by the soil moisture or the state of the vegetation (Yang et al., 2015). The commonly-used indices are the Crop Moisture Index (CMI) (Palmer, 1968) and theWater Deficit Index (WDI) (Moran et al., 1994).

Hydrological drought is related to a period with inadequate surface and subsurface water resources for the established water uses of a given water resources management system. Streamflow data have been widely applied for hydrological drought analysis (Mishra and Singh, 2010). Socio-economic drought differs markedly from the other types, because it associates human activity with elements of meteorological, agricultural, and hydrological drought. It is characterised by the failure of water resource delivery systems to meet the water demands, thereby associating droughts with the supply of, and demand for, an economic good (water) (AMS, 2004) (Mishra and Singh, 2010). The simplest way for monitoring drought conditions is to use drought indices, because they provide a quantitative method for determining the onset and end of a drought event, by using an index value that indicates the level of drought severity (Tabari et al., 2013).

2.6.3 Drought indicators

In addition to the types of drought, drought can be fundamentally characterised into three dimensions: their severity, duration and spatial distribution (Zarch et al., 2015). Duration of a drought depends on the region and can vary from a week, to a few years. The dynamic nature of drought allows for wet and dry spells to be experienced, when considering various timescales (Mishra and Singh, 2010). Severity is either the degree of the precipitation shortfall (magnitude) and/or the severity of the impacts resulting from the deficit (Keyantash and Dracup, 2004). It is generally measured by the departure of some climatic parameter (precipitation), indicator (reservoir levels), or index (e.g. the Standardized Precipitation Index) from the normal, and it is closely linked to duration, in the determination of the impacts (Wilhite and Buchanan-Smith, 2005).

Additional characteristics include frequency and magnitude (cumulative effect) (Zargar et al., 2011). The magnitude of drought impacts is closely related to the timing of the onset of the precipitation shortage, its intensity, and the duration of the event. Droughts also differ in terms of their spatial characteristics (Wilhite and Buchanan-Smith, 2005). The areas affected by severe drought evolve gradually, and regions of maximum intensity shift from season to season. It is rare to find a drought that affects the entire extent of large countries, such as Brazil, China, India, the United States, South Africa or Australia. During the severe drought of the 1930s in the United States, for example, the area affected by severe and extreme drought reached 65% of the country (Wilhite and Buchanan-Smith, 2005).

Drought indices are thus quantitative measures that characterize drought levels by assimilating data from one or several variables (indicators), such as precipitation and evapotranspiration, into a single numerical value (Zargar et al., 2011). Such an index is more readily useable than raw indicator data and these indicators describe the magnitude, duration, severity and the spatial extent of drought (Zargar et al., 2011). Typical indicators are based on meteorological and hydrological variables, such as precipitation, streamflows, soil moisture, reservoir storage and groundwater levels, and several of these indicators can be combined into a single indicator on a quantitative scale, called a drought index (Zargar et al., 2011). A drought index provides a comprehensive picture for drought analysis and decision-making that is more readily useable, compared with the raw data from indicators (Hayes, 2006). Drought indices reflect the varying events and conditions by showing the dryness of the climate, based predominantly on precipitation, or they correspond to delayed agricultural or hydrological impacts, such as low soil moisture or reservoir levels (Heim, 2002).

Common drought indices are categorized, based on the type of impact to which they relate. As a result, different metrics of drought highlight different variables of interest, such as precipitation for meteorological droughts, soil moisture for agricultural droughts, and streamflow for hydrologic droughts. Mishra and Singh (2010), Zargar et al. (2011) and Dai (2011) have presented comprehensive reviews of commonly-used drought indices, including the statistical characteristics of these indices (such as the frequency, number of occurrences, and duration), which are important for short and long-term water management actions. Comparisons of the advantages, disadvantages and applicability of the various drought indices have also been reported in the other literature (Guttman, 1998; Sims and Ramon, 2002; Mishra and Singh, 2010; Zargar et al. 2011; Touma et al., 2015). The focus for this study is on meteorological and hydrological drought, thus these indices are explored further and in more detail.

2.6.3.1 Meteorological drought indices

Meteorological drought indicators are associated with climatological variables, such as precipitation, temperature, and evapotranspiration. They include the Palmer Drought Severity Index (PDSI) (Palmer, 1965), the Standard Precipitation Evaporation Index (SPEI) (Vicente-Serrano et al., 2010) and the Standard Precipitation Index (SPI) (McKee et al., 1993). Precipitation is a widely-used and useful indicator; it can directly measure water supplies, it influences hydrological indicators, and it can reflect drought impacts over different time periods and sectors. Yet meteorological indicators, such as precipitation, can pose analytical challenges because of its temporal and spatial variability, the lack of data and insufficient observation stations.

The Palmer Drought Severity Index (PDSI) is a popular meteorological drought index (Zargar et al., 2011). The PDSI concept views drought in terms of water supply and demand, as opposed to a precipitation anomaly (Zargar et al., 2011). The intensity of short-term drought, over a period of a few weeks, is described as single numerical values that include, inputs of soil moisture, runoff, evaporation and temperature. Precipitation and temperature are used for estimating moisture supply and demand within the soil (Mishra and Singh, 2010). The emphasis is placed on the abnormalities in moisture deficiency, rather than on weather anomalies. The limitations of the PDSI have been well-documented (McKee et al., 1995; Guttman, 1998). According to Guttman (1991), it performs poorly in indicating short-term changes in soil moisture, it has a variable performance across time periods and regions and it does not take into account human water balance impacts, such as irrigation (NDMC, 2006). Thus the PDSI lacks the multiscalar character essential for assessing drought in relation to different hydrological systems, as well as for differentiating different drought types (Vicente-Serrano et al., 2012).

The SPEI was developed by Vicente-Serrano et al. (2010), who wanted an index that could overcome the supposed shortcomings of the SPI and that could not identify the pattern of increase in the duration and magnitude of droughts resulting from higher temperatures. It is based on the SPI, but it incorporates temperatures data and considers the water balance and precipitation. The SPEI thus uses precipitation and potential evapotranspiration, combined with 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 (Zargar et al., 2011). The SPEI is based on a monthly climatic water balance (precipitation minus PET), which is adjusted by using a three-parameter log–logistic distribution. The values are accumulated at different timescales, following the same approach used in the SPI, and they are converted to standard deviations, with respect to average values.

For the SPEI, the importance of variables is negligible and droughts are assumed to be controlled mainly by the temporal variability of precipitation (Vicente-Serrano et al., 2012). A study by Edossa et al. (2014) used the SPEI to characterise drought events in the central region of South Africa and to examine the association between droughts and El

Nino events. The SPEI was formulated based on two meteorological variables, precipitation and temperature, and could account for the possible effects of temperature variability and temperature extremes (Edossa et al., 2014). Where there are no apparent trends in temperature, the SPEI is nearly equivalent to the SPI or other precipitation drought indices (Zargar et al., 2011). In terms of this current study, temperatures in the Drakensberg are relatively low all year round and there is negligible evaporation, thus this index is not completely suitable. The SPEI and PDSI, although commonly-used, are not used in this research because a more appropriate index, the SPI, was found to be more applicable to the area of study.

2.6.3.1.1 Standard Precipitation Index (SPI)

The standard precipitation index (SPI) was originally developed by McKee et al. (1993) and it is another commonly-used index to monitor and analyse drought events. The index quantifies precipitation deficits at variable timescales and provides an indication of drought intensity and duration, based on the historical distribution of rainfall (Malherbe et al., 2015). The SPI is easy to use because it only requires rainfall data, it is easily adapted to the local climate and it can be computed at almost any timescale (Tue et al., 2015). Its simplicity and application, over a wide range of climatic regions, make it an appealing index for determining drought (Hayes et al., 1999).

More specifically, the SPI for any location is calculated by using long-term rainfall data for a desired period and this is fitted to a probability distribution (Mishra and Singh, 2010). SPI compares precipitation to its multi-year average and avoids discrepancies by transforming the distribution of precipitation to a normal distribution, so that the mean SPI for the location and period is zero (McKee et al., 1993). Values above zero indicate wet periods and values below zero indicate dry periods (Zargar et al., 2011). The SPI score for a drought represents how many standard deviations its cumulative precipitation deficit deviates from the normalized average (Drought Watch, 2010).

Its fundamental strength is that is can be calculated over a variety of timescales, thus making it more suitable for comparison of droughts. Mckee's (1993) index can be computed for any time period, but 3, 6, 12, 24 and 48-month periods are typically used. The SPI is also useful because it can reflect changes in different water features, as a precipitation deficit affects different water resources variably and gradually (Zargar et al., 2011). Further, the SPI is able to identify different drought types, since particular systems and regions can respond to drought conditions at very different timescales (Vicetta-Seranno, 2012).

The SPI has been used to construct drought climatology in Europe (Lloyd-Hughes and Saunders, 2002) and Canada (Bonsal et al., 2013) and is used as an operational drought-monitoring tool in the USA (Hayes et al., 1999). A study in Greece was successfully able to classify the years of rainfall shortages to different drought intensities and duration, using the SPI on a spatial and temporal basis (Livada and Assimakopoulus, 2005). Rouault and Richard (2003) demonstrated the use of the SPI for monitoring the intensity and spatial extent of droughts at various timescales for South Africa. They regard the SPI as being consistent with respect to the spatial distribution of rainfall that occurs with great variability in South Africa. The studies found the index useful for helping to develop the climatology of the spatial extension and intensity of droughts, which will assist in understanding the characteristics of drought and providing an indication of the possible reoccurrence of drought at various levels of severity (Rouault and Richard, 2005).

The main criticism of the SPI is the fact that the calculation is based solely on precipitation data (Vicente-Serrano et al., 2012). The SPI index relies on two assumptions: (i) that precipitation variability is a lot higher than other variables, such as temperature and potential evapotranspiration (PET) and (ii) that the other variable has no temporal trend (Vicente-Serrano et al., 2012). Thus, it is ideal in a place such as the KwaZulu-Natal Drakensberg as there is very low evaporation and low temperature variability. Single drought indices clearly have their weaknesses, as they do not encapsulate the complex nature of drought (Mishra and Singh, 2010). Some drought indices specifically reflect one type of impact or application, while others can be configured to correspond to varying impacts, and thus including drought type (Zargar et al., 2011). However, the SPI can be deployed for longer timescales to reflect agricultural and hydrological droughts/impacts (Zargar et al., 2011). For this study, the SPI is used for meterological analysis only.

2.6.3.2 Hydrological drought indices

Several drought indices were developed to characterize hydrological droughts over the years (Dracup et al., 1980; Tate and Gustard, 2000; Smakhtin, 2001; Heim 2002; Tallaksen and van Lanen, 2004). The Surface Water Supply Index (SWSI), designed by Shafer and Dezman (1982), was developed for mountainous regions, where snowmelt contributes to reserves the and incorporated variables, such as snowmelt, streamflow, precipitation and reservoir storage. Shukla and Wood (2008) developed the Standardized Runoff Index (SRI), based on the developing concepts of the SPI, to analyze simulated runoff data on different timescales. They used the SRI, in conjunction with the variable infiltration capacity (VIC) model, to estimate runoff and the SRI. The hydrological drought index is produced by normalizing monthly streamflow data. The Regional Drought Area Index (RDAI) is based on daily streamflow data and was used to represent drought areas in north-western Europe (Fleig et al., 2010; 2011) by identifying trends between hydrological drought and different weather types. Sharma and Panu (2010) suggested the use of a Standardized Hydrological Index (SHI) as a measure for defining and modelling hydrological droughts. The SHI is based on the concept of the SPI, however, SHI is only a standardized entity and not normalized, and therefore depicts the non-normalized character of streamflow. The indices devised for characterizing hydrological drought, in general, are data demanding and computationally intensive. In contrast, the Streamflow Drought Index (SDI), developed recently by Nalbantis and Tsakiris (2009), is a very simple and effective index for hydrological droughts.

2.6.3.2.1 Streamflow drought Index (SDI)

The SDI is analogous to the Standardized Precipitation Index (SPI), which the WMO has declared an official meteorological drought index (Hayes et al., 2011). The SDI was used by Tabari et al. (2013) in north-western Iran and by Rimkus et al. (2013) in the Neman River Basin. The SDI allows one to determine the index for different time periods, 3, 6, 12 and 24 month periods within a range of years. For example, Tabari et al. (2013) used overlapping periods of 3, 6, 9 and 12 months at 14 hydrometric stations in the north-western area of Iran over the period 1975–2009. Some streamflow volume data do not follow normal distributions and must be transformed, using probability distributions such as the log-normal, uniform or exponential. Tabari et al. (2013) found the log-normal distribution best to fit the long-term streamflow data.

2.7 Droughts in Southern Africa

Drought is a part of natural climatic variability on the African continent and is occurring at intra-annual, interannual, decadal and century timescales (Nicholson, 2000). Drought is one of the most devastating natural disasters because of the low socioeconomic status of the sub-regions of many African countries. The annual rainfall of most areas in southern Africa is <500 mm and it exhibits a high degree of spatio-temporal variability (Mason and Jury, 1997; Rouault and Richard 2003). The socio-economic impact of drought is usually severe in regions with an annual rainfall of less than 500 mm (Richard and Poccard, 1998). South Africa has an average of 600 mm rainfall a year but there is large spatial variability across the country, where some areas fall below and exceed the average (Otieno and Ochieng, 2004). Drought threatens the economy of many southern Africa nations, where socio-economic activities depend on rain-fed agriculture (Meque and Abiodun, 2015). Thus rainfall variability over southern Africa has long been the subject of investigation, due, in part, to the importance of rain-fed agriculture in the region, the arid to semi-arid nature of the climate, and the high degree of interannual and interdecadal variability observed over the region (Todd et al., 2004).

The failure of the rainfall regime is likely to cause serious disruption to the natural and agricultural ecosystems and it may lead to a corresponding failure of many sectors of the economy (Manatsa et al., 2008). Southern Africa is largely semi-arid and is characterised by a high seasonal rainfall variability (Nel, 2009) that experiences drought periods with relatively high frequency and severity. With southern Africa and South Africa being water-scarce, droughts are of great importance in the planning and management of water resources (Mishra and Singh, 2010). Hence, drought is a threat to water management and agriculture in southern Africa, especially in the countries where the socio-economy of people depends on agriculture (Jury and Mwafulirwa, 2002).

According to the SPEI data analysis, there was a statistically significant increase (at the 99% significance level) in the areas under all categories of drought (moderate, severe and extreme droughts) for the African continent between 1901-2011 (Masih et al., 2014). This is supported by studies conducted at continental scale (Dai, 2011; 2013), as well as at country and regional level (Richard et al., 2001; Manatsa et al., 2008; Touchan et al., 2008; Kasei et al., 2010). Extreme continental drought years were identified as being from 1972-1973, 1983-1984 and 1991-1992 and none of the previous droughts during the 20th century were as widespread and intense as these were. In addition, many severe and prolonged droughts have been recorded in the past, for example, the 1970s and early 1980s droughts in western Africa (Sahel), the 1999-2002 drought in northwestern Africa, the 2010-2011 drought in eastern Africa (the Horn of Africa) and the 2001-2003 drought in southern and south-eastern Africa, to name a few (Masih et al., 2014).

DEWFORA (2012) reported that in the period 1980-2000, the southern African region was struck by four major droughts, notably in the years 1982/1983, 1986/1987, 1991/1992 and 1994/1995. The drought of 1991/1992 was the most severe in the region in recent history. After 2000, important droughts included the years 2002/2003/2004 and 2005/2006 (DEWFORA, 2012). Manatsa et al. (2008) identified droughts in Zimbabwe, based on SPI estimation from the regionally averaged rainfall for the period 1900-2000, where 1991-1992 was considered to be the most extreme drought of the 20th century. A study by Mussa et al. (2014) analyzed whether groundwater can be used as an emergency source of water in cases of severe droughts in the Crocodile Catchment area in the Mpumalanga Province of South Africa from 1940-2011. The study used the SPI and SRI drought indicators to identify meteorological and hydrological droughts, respectively. Three severe droughts occurring in 1983, 1992-1995 and in 2003-2004 were found and these droughts were also noticed in much of the rest of South Africa and its neighbouring countries. The 1992-1995 drought was identified as the most severe drought in the past 70 years, where the upper and lower areas of the Crocodile Catchments were the most affected (Mussa et al., 2014).

In the past few decades, southern Africa has experienced severe droughts, with extensive socio-economic impacts (Meque and Abiodun, 2015). Figure 2.3 shows a graph adapted from a report by Gommes and Petrassi (1996) for the Food and Agriculture Organization of the United Nations (1996), which indicaes the rainfall indexes for the southern Africa group of countries (Botswana, Lesotho, South Africa and Swaziland) for the years 1960-1993. The rainfall indexes of 1972-1973, 1982-1984 and 1991-1992 have been encircled in red, to highlight the low values, indicating that a drought occurred in the region during these years. In particular, the 1982-1984 drought event stretched across southern Africa and it was at the time the most severe drought ever experienced in the summer rainfall regions of southern Africa since the 1920s (Dent et al., 1987; Gommes and Petrassi, 1996). These countries were also severely affected by the 1991-1992 drought, which was the most severe after the 1981-1895 droughts (Gommes and Petrassi, 1996).

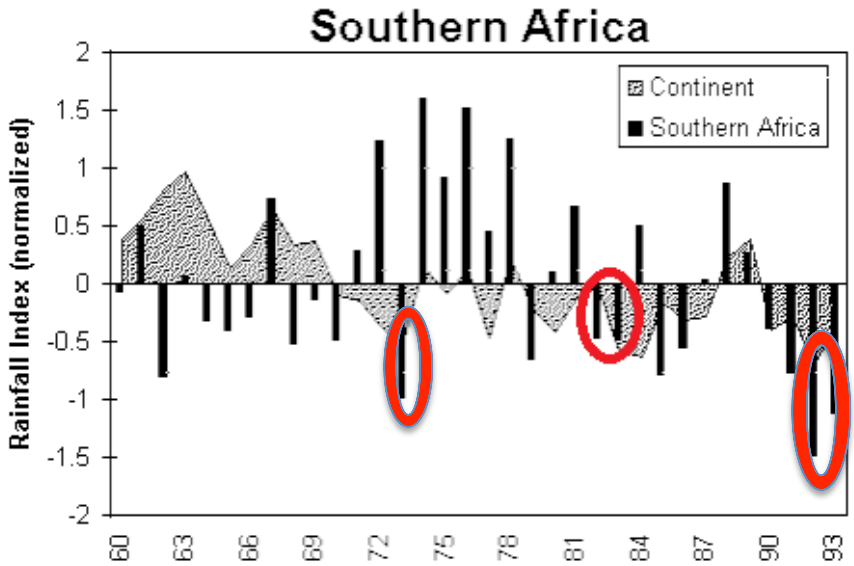


Figure 2.3 Graph indicating rainfall indexes for the southern African group of countries for the years 1960 - 1993. Adapted from Gommes and Petrassi (1996).

The major consequences of the early 1980's drought in southern Africa were reduced agriculture and increased malnutrition (Vogel, 2000). The impact on agriculture and food production in the South African Development Community (SADC) countries during this drought saw economic and social conditions deteriorating and the real regional output was estimated to have fallen by 1.3 percent in 1981 (Vogel, 2000). From 1982 to 1984, the maize yields south of the Zambezi Valley declined to 10% of the historical values and 80% of the livestock perished (Makarau, 1995). In 1991-1992, one of the most extensive and long-lasting droughts in southern Africa produced widespread dry conditions across the whole region (Vogel and Drummond, 1993). The availability of water became critical, as 90% of small dams dried up and the rainfall was reduced to 50% below the normal value (Jury and Mwafulirwa, 2002). The drought destroyed crops and caused large agricultural and economic losses in southern Africa (Calow et al., 2010).

The most common source of episodic droughts around the world is the El Niño/Southern Oscillation (ENSO) (Trendberth et al., 2013). El Niño events have caused major droughts over Australia, Indonesia, south-east Asia, parts of Africa and the north-east of Brazil (Trendberth et al., 2013). This is a result of the main rainfall systems in the tropics moving off-shore over the tropical Pacific, combined with the much warmer than normal waters, often leaving weakened monsoons behind (Trendberth et al., 2013). The atmospheric circulation creates favourable conditions for drought, often through teleconnections, which can be initiated by sea surface temperature anomalies.

Masih et al. (2014) compiled a summary of the selected literature reviewed in their study on droughts on the African continent. The section on southern Africa is presented in Table 2.1 and indicates the drought years. In this summary, Masih et al. (2014) made an effort to avoid the duplication of similar studies and yet provide geospatial and temporal coverage. Meteorological drought remains the main subject of most of the studies, followed by agricultural drought. These research studies have a common thread and show that, despite the regional differences in the factors causing drought, the El Nino Southern Oscillation (ENSO) and SSTs are the main influencing factors across the continent.

Nicholson and Kim (1997) studied the correlation between precipitation and ENSO in the Pacific and found that 15 of the 20 analysed rainfall events appeared to be modulated by the ENSO. Their results suggest that the southern part of Africa is negatively correlated with the warm ENSO. Upon analysing the rainfall variability in agricultural production in Zimbabwe, Phillips et al. (1998) found a decrease in precipitation during the El Nino phase. Edossa et al. (2014) found a clear indication that most of the droughts identified by the SPEI follow the El Nino events, when looking at drought over the central region of South Africa. Richard et al. (2001) examined droughts during the summer rainfall periods (January – March) of 1950 to 1988 in southern Africa and found that droughts during the 1970–1988 period were intense and widespread, compared to those during 1950–1969. ENSO was the main governing factor for drought between 1970-1988; however, this observation requires a word of caution because droughts may not occur during the El Niño periods, as happened during 1925–1926 and 1997–1998. For the droughts during 1950-1969, regional oceanic and atmospheric anomalies (the southwestern Indian Ocean SST) were named as the main causes (Richard et al., 2001; Fauchereau et al., 2003).

Table 2.1 Summary of literature pertinent to droughts in southern Africa adapted from work by Masih et al. (2014).

References for Southern Africa	**Drought enlisted by region/country/basin from 1900 to 2013**
Belbase and Morgan (1994)	Southern Africa – Botswana: 1978–1979, 1982–1987, 1991–1992
Edossa et al. (2014)	South Africa (central region) Second half of 20th century: 1964, 1968, 1970, 1982, 1983, 1984, 1992/1993
Manatsa et al. (2008)	Southern Africa – Zimbabwe: 1902–1903, 1911–1916, 1926–1927, 1941–1942, 1963–1964, 1972–1973, 1982–1984, 1986–1987, 1991–1992
Msangi (2004).	Southern Africa: 1902, 1909–1911, 1917–1918, 1921–1922, 1925, 1929, 1933–1934, 1939–1940, 1953, 1969, 1972–1973, 1976, 1980–1982, 1984–1985
Mussá et al. (2014)	South Africa, Crocodile River catchment: 1945, 1951, 1958, 1966, 1970–1971, 1978, 1983–1984, 1992–1995 and 2003–2004
Malherbe et al. (2015)	South Africa: 1922, 1926, 1927, 1933, 1949, 1966, 1970, 1983, 1992, 2004
Richard et al. (2001)	Southern Africa: 1951, 1960, 1964, 1965, 1968, 1970, 1973, 1982, 1983, 1987
Rouault and Richard (2003)	South Africa: 1926, 1933, 1945, 1949, 1952, 1970, 1983, 1992
Rouault and Richard (2005)	Southern Africa (south of 10°S): 1906, 1916, 1924, 1933, 1949, 1970, 1983, 1984, 1992, 1993, 1995, 1996, 2002, 2003, 2004
Vogel et al. (2010)	Southern Africa: 1982–1993, 1991–1992, 1994–1995, 2001–2003

Rouault and Richard (2005) studied the temporal and spatial extent of the droughts in southern Africa from 1901-1999. The ENSO conditions were attributed to eight out of these twelve droughts that occurred during the 20th century (Table 2.1), which confirms the strong relationship between the ENSO and the drought events in southern Africa. They found that the area of the African continent under drought has increased significantly, especially after the 1980s. Malherbe et al. (2015) confirm that, through SPI and SPEI time series, the significant decadal modulation of the impact of ENSO on the South African summer rainfall and on the occurrence of extreme droughts in the northeastern region. However, research over and above that of ENSO, is needed to determine the causes of the decadal variability of drought occurrence. Kurukulasuriya and Rosenthal (2013) found that droughts are projected to increase in intensity and frequency, especially in southern Africa, where the El Niño/Southern oscillation is contributing to significant climate variability in the region.

2.7.1 Drought in South Africa

More than 90% of South Africa is either arid or semi-arid, with drought being a typical feature of the climate (Hoffman et al., 2009). Thus, drought is a regular and recurring feature of the South African climate (Rouault and Richard, 2005), and it is an endemic feature of agriculture. Large parts of the country are exposed to moisture stress and most agricultural activities take place under arid to semi-arid conditions (Bruwer, 1993; Rouault and Richard, 2005). Model predictions indicate that there has been a reduction in the mean annual rainfall, an increase in the interannual variation, and an increase in the frequency of extreme events, including drought, particularly in the winter-rainfall region of southern Africa (Tadross et al., 2005). Such predictions raise the question of whether there isevidence in the climate records that the annual rainfall has already declined and that the incidence of drought has increased over the past 100 years in the winter-rainfall region of South Africa (Hoffman et al., 2009).

South Africa has experienced at least eight major droughts in the 20th century, namely, 1926, 1933, 1945, 1949, 1952, 1970, 1983 and 1992 (Rouault and Richard, 2005). Mussa et al. (2014), whose study extended to the year 2013, found similar years of drought, including 2003- 2004. Of these droughts, the most severe were experienced in 1983, 1992-1995 and 2003-2004. The most severe to extreme droughts, based on the SPI, occurred in 1933 (48% of the country experienced severe to extreme drought), followed by 1949 (46.5%), 1992 (33%), 1970 (28%), 1926 (27%), 2004 (26%), 1983 (26%), 1927 (22%), 1922 (21%) and 1966 (21%) (Malherbe et al., 2015). The findings for the period ending in 2000, are in agreement with the findings by Rouault and Richard (2003). In relation to the African continent, Masih et al. (2014) found that extreme droughts, over the period 1901-1999, occurred in 1972-1973, 1983-1984 and 1991-1992. Many of the droughts may be short-term and they may be followed by a recovery during subsequent years of higher rainfall, and in some cases, they can trigger substantial ecological and socioeconomic changes (Bomhard et al., 2005).

Droughts at a yearly timescale have a strong impact on agriculture and the hydrological impacts are important, especially concerning reservoir levels and streamflow (Malherbe et al., 2015). An analysis of summer droughts at a 24-month time scale would include the impacts of hydrological and agricultural droughts. Malherbe et al. (2015) found that the years with the most extensive extreme and severe droughts over two years include 1947, 1933, 1966, 1970, 1983, 1992, 1993, 2004 and 2005. Vogel (2000) found that hydrological droughts occurred in South Africa between 1920 and 2000, namely, in 1930-1932, 1950-1953, 1967- 1972, 1978-1986 and 1989-1991.

Reservoir water levels dropped dramatically during the early 1980s and 1990s and, in some cases, levels dropped so low that it became necessary to augment water in some dams, such as the Vaal Dam (Vogel, 2000). These dams are essential for supplying water to Gauteng and shortages can have a major effect on industrial productivity, national power generation and livelihoods (Vogel, 2000). While the levels of dams, such as the Vaal Dam, evoke national concern, the levels of other smaller dams, often located in remote rural areas, are also drastically reduced during drought periods. Other sources of water that serve boreholes and pumps, such as groundwater, are also placed under strain.

The Eastern Cape has experienced deficiencies in water resources due to periodic droughts throughout the past 20th century (Jury and Levey, 1993). The most notable period occurred between 1940 and the 1970s, followed by a decade of heavy rainfall and downward trends in rainfall to 1993 (Jury and Levey, 1993). During the late 1940s, the droughts were concentrated over the central and southern parts of the country, and in the mid-late 1960s, they were focused over the northern and north-eastern interior (Malherbe et al., 2015). The early 1980s, 1990s and 2000s were expressed strongly over the northern and eastern to north-eastern areas (Malherbe et al., 2015).

The drought in the Eastern Cape in the early 1990s prevailed for 18 months (Jury and Levey, 1993). Droughts in the region have been found to recur at intervals of 3.45 and 18.2 years. The recurring droughts have had consequent unfavourable impacts on agriculture production and water resources. In particular, in early 1992, dam levels dropped to hold only 30% of their capacity (Jury and Levey, 1993). Malherbe et al. (2015) found three major spectral peaks, based on the SPI time series, in quaternary catchments of South Africa, namely, 17-20 year, 10-12 year, and 4-7 year cycles. The 17-20 year drought cyclicity was found to reach significant values and is the most prominent, as it correlates with the 17-20 year rainfall cycle (Tyson et al., 1976; Jury and Levey, 1993; Kane, 2009). The wet conditions experienced over northeastern South Africa in the late 1990s, followed by predominant dry years over 2002-2008, and then a reappearance of wet conditions since 2011 seemingly follow this cycle (Malherbe et al., 2015). The 10-12 year and 4-7 year cycles are not precluded, as they could be related to significant regional variability.

2.8 Summary

This review has established the background literature on rainfall variability and drought in southern Africa and South Africa, highlighting the prevalence of increasing variability in rainfall and the resultant dry conditions. The theory and main concepts have been established, as well as evidence of the existence rainfall variability and drought and the causes thereof. The reviewed literature serves as the background on which the study was developed. Rainfall variability and the effects on hydrology in the KwaZulu-Natal Drakensberg has not been extensively studied or updated, nor has the occurrence of drought conditions in the area. Identifying the recent trends in rainfall provides a platform for a comparison with previous research, conducted by Nel (2009), which will allow for a better understanding of rainfall in the area. A catchment scale analysis of the rainfall trends in the area has not been conducted before, nor have the effects of altitude been distinguished, at ths scale. Rainfall in the Drakensberg is vital, both regionally and locally, and this identified knowledge gap therefore provides a sound motivation for this research.

3. STUDY AREA

3.1 Introduction

The topography of south-east Africa is characteristiced by a high-elevation interior, with a prominent erosional escarpment (the Great Escarpment). The escarpment is an important boundary that separates the coastal lowlands and fold-mountains around the perimeter of southern Africa, from the high continental plateau and semi-arid river systems of the interior (Moore et al., 2009). The KwaZulu-Natal Drakensberg is part of the Great Escarpment of southern Africa that extends as a passive margin around the sub-continent (Nel and Sumner, 2008) and is situated between 160 km and 240 km inland of the eastern coast of South Africa. The Drakensberg Escarpment creates a natural watershed, which is the boundary between eastern Lesotho and South Africa (between 28.50°S and 28.50°E; 30.50° S and 29.50° E). The study area for this research is located in the central and northern parts of the KwZulu-Natal Drakensberg area, which is highlighted by the red line in Figure 3.1.

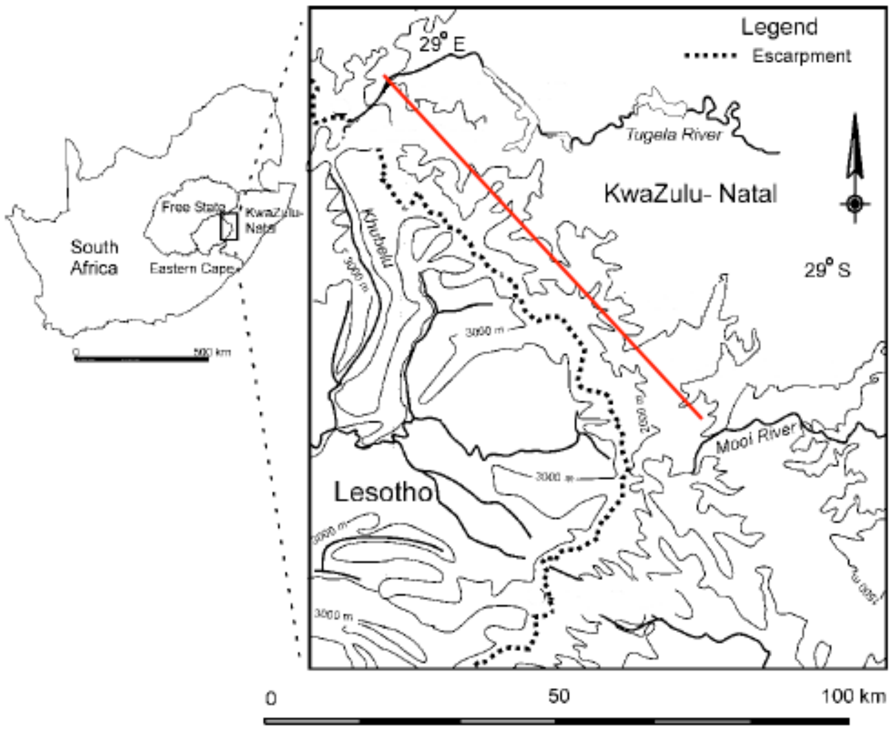


Figure 3.1 Location map of the Drakensberg Escarpment showing the study area, the central and northern KwaZulu-Natal Drakensberg, highlighted by the red line (Adapted from Nel, 2008).

3.2 Topography and Geology

The topography of the continental divide between the Indian and Atlantic Oceans drops to the east, into South Africa, and slopes more gently to the west, into Lesotho (Zunckel, 2003). The mean elevation in the east rises progressively from the coastline to around 1600 m, at approximately 120 km inland, and then rises more gradually up to around 2000 m at the base of the Drakensberg Escarpment, approximately 150 km inland of the coastline (van der Beek et al., 2002). The Drakensberg Escarpment reaches over 3 400 m a.s.l, with the highest peak (3482 m a.s.l.) lying 4 km back from the edge of the escarpment in eastern Lesotho (Killick, 1990).

The geology of the Drakensberg comprises of two clearly-defined areas that are consistent throughout the mountain range, east of the escarpment. The main escarpment is comprised of the summit area, which consists of a basalt scarp of volcanic origin rising from about 2000 m to in excess of 3000 m (Sumner et al., 2009). The uniform layer of lava is superimposed onto sandstone and is continuously being eroded, to expose the underlying interpolated sand and shale strata of the Karoo system, which is known, in this context, as the 'Little Berg' (Sycholt, 2002). The sandstone formation found between approximately 1400 m to 1900 m is made up of several layers of different types of sandstone and this is known as the 'Little Berg' (Figure 3.2). These sediments lie nearly horizontal inland and beneath the basalt, but are tilted westward, exposing progressively older strata (approximately 160-220 million years old), until the basement is reached, 30 km from the coast (van der Beek et al., 2002).

The Little Berg consists primarily of three sandstone formations, namely, the Clarens, Elliot and Molteno, overlaying outcrops of the Beaufort series in the valley floors (Figure 3.2). Differences in sandstone formations are distinguished by varying degrees of sand, clay and mud content. The upper Beaufort layers lie at 1380-1480 m a.s.l and consist of mudstones, siltstones, shales and sandstones (Moore and Blenkinsop, 2006). Over this lies the Molteno formation (1480-1580 m a.s.l.), consisting of white, gritty, coarse-grained sandstones (Brown et al., 2002). This is overlain by the Elliot formation, which is a red mudstone layer reaching 1710 m a.s.l., and above this, reaching 2000 m a.s.l., lies the uppermost sequence of fine-grained sandstone known as the Clarens Formation, which forms the lower slopes of the Drakensberg (Sumner et al., 2009). Differences in the resistance to erosion of individual units in these sedimentary successions is expressed by secondary escarpments and ridges that extend for up to 10 km or more beyond the main escarpment face (Brown et al., 2002).

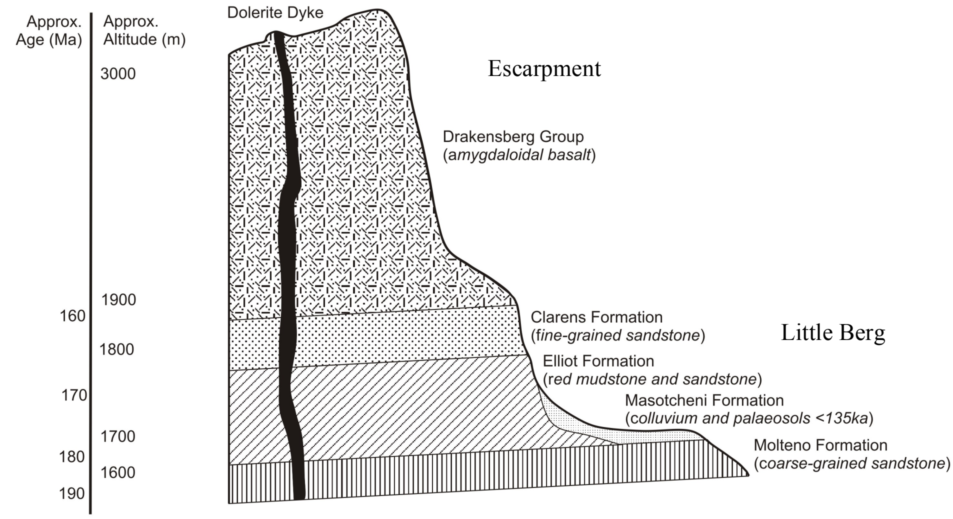


Figure 3.2 West-East cross-section showing the geology of the Drakensberg escarpment and Little Berg (Adapted from Sumner et al., 2009).

3.3 Drainage

South Africa's most valuable source of water is the eastern escarpment region of the KwaZulu-Natal Drakensberg and the Lesotho Highlands (Nel, 2009). Relatively linear river systems, which have their headwaters in the escarpment region, drain the area seaward of the continental divide (van der Beek et al., 2002). The Drakensberg escarpment represents the catchment boundary between the interior catchments of eastern Lesotho, which are home to the headwater tributaries of the Orange River, and the steep inner catchments of the KwaZulu-Natal rivers that supply the rivers flowing to the east coast, namely, the Tugela, Mooi and Mkomazi Rivers (Nel and Sumner, 2006; Nel, 2009). The local relief increases rapidly from the coast and reaches a maximum at the Drakensberg Escarpment and across the highly-dissected Lesotho Highlands (Summerfield, 1991).

The KwaZulu-Natal catchments are known to produce nearly twice as much total runoff per unit rainfall than the average for South Africa as a whole (Schulze, 1979) and a quarter of South Africa's streamflow (Whitemore, 1970). These high rainfall totals and runoff yields supported the construction of two interbasin transfer projects, the Tugela-Vaal (1981) and the Lesotho Highlands Water Project (1986 treaty), that rely on the upper catchments to supply water to the Gauteng Province (Nel and Illigner, 2001; Matete and Hassen, 2006). More locally, the runoff generated feeds into rivers that supply water used by farmers to irrigate and for domestic use in the surrounding towns. The function of this region as a water catchment is vital, as it is one of only a few in southern Africa where the long-term annual precipitation exceeds evaporation (Zunckel, 2003). Biologically and hydrologically important wetlands provide ecosystem services, by sustaining the flow of high quality water into the major rivers, the Orange River and the Tugela River, that supply interbasin transfer schemes (Zunckel, 2003).

3.4 Climate

The weather and climate of South Africa is strongly influenced by the location of the country, relative to the global circulation patterns (Vogel, 2000). The weather is dominated, all year round, by a belt of subsiding air from the Hadley Cell and the interior is characterized by a high pressure and north-easterly inflow of tropical air (Vogel, 2000). South Africa lies within the subtropical belt of high pressure, which accounts for its dry climate (Preston-Whyte and Tyson, 1988). Cold front weather systems from the west and tropical systems from the east influence and the day-to-day weather and the oceanic influences have an important role in influencing the strength and progression of these systems (Vogel, 2000).

Cold frontal systems are made up of different air masses of warm and cold air and, when they meet, a front of changing pressures, wind directions and temperatures is formed (Harrison, 1984; Vogel, 2000). The circumpolar vortex dominates the circulation of the atmosphere in the mid-latitudes and is made up of winds that carry the weather systems with them, such as low-pressure systems, and influence the upward or downward movement of air thus, controlling the conditions for stability (descending air) or instability (uplift) (Vogel, 2000). The low-pressure systems move in an easterly direction across the country, bringing rain to the south-west and southern coasts (Preston-Whyte and Tyson 1988; Vogel, 2000).

The cold fronts move in a west-northwesterly to east-southeasterly direction and are generally deflected parallel to the line of the South African escarpment (Tyson et al., 1976). Cold fronts frequently cross over coastal areas and penetrate inland and are also an important source of rainfall (Preston-Whyte and Tyson 1988). Cold-frontal rainfall is normally associated with low-level stratocumulus clouds (Nel et al., 2010). Warm fronts are seldom experienced in southern Africa, as they tend to form in the south of the country and take up a west-east alignment (Vogel, 2000). Another important weather system, affecting the east coast, is that of tropical cyclones (Harrison, 1984). These originate in the Indian Ocean and move in a south-westerly direction onto the mainland, bringing heavy rain and winds. They seldom penetrate below 30°S but have been known to affect northern KwaZulu-Natal (Vogel, 2000).

The distribution of rainfall is thus related to these systems. Over the interior, there is a distinct east-west trend in rainfall (Vogel, 2000). North-easterly airstreams affecting the eastern Highveld bring rainfall totals of approximately 800 mm that are concentrated in the summer months (D'Abreton and Tyson, 1994). Rainfall trends decrease towards the west, which receives around 125 mm near the coast, and this provokes arid and desert conditions (Vogel, 2000). The Highveld experiences thunderstorms in the warm summer months. The temperature over the interior of South Africa is linked to the high-pressure field over the region, subsiding air from the tropics, together with a descending Hadley Cell, bringing clear conditions (Partridge et al., 1997).

Altitude and the surrounding oceans influence the general climate, however, it is important to note that there are regional variations (Sycholt, 2002). In the KwaZulu-Natal Drakensberg, the rainfall average differs vastly from South Africa's yearly average of approximately 600 mm. It rises from about 800 mm near the foothills, to over 1000 mm on the escarpment (Schulze, 1979; Nel and Sumner, 2008). The major sources of precipitation over the Drakensberg consist of two types, namely, large-scale thunderstorms and orographically-induced storms, that develop in the extended summer period (Tyson et al., 1976). The Drakensberg is a summer rainfall dominated area (Nel and Sumner 2006). The El Niño/Southern Oscillation influences the summer rainfall variability of the KwaZulu-Natal Drakensberg, with a statistically significant correlation existing between the summer rainfall in the Drakensberg and lagged the Southern Oscillation Index (Nel, 2009).

In winter, approximately 40 cold fronts affect KwaZulu-Natal annually (Grab and Simpson, 2000), bringing occasional rain and snowfall to the Drakensberg. Cold fronts and the associated cut-off lows account for about 80% of the snow cover over the Drakensberg (Mulder and Grab, 2009). In the Drakenberg, the escarpment zone was found to influence the effectiveness of the winter cold fronts in producing substantial rainfall at high altitude (Nel et al., 2010). The deflection of the cold front path, together with the low level of cloud could possibly be the reason for the high altitude areas receiving less winter rainfall, at lower intensities and energy, than in the foothills (Nel et al., 2010). Heavy frosts occur during the winter months, between April and September, and snow is likely to fall at any time of the year on the KwaZulu-Natal escarpment and in Lesotho (Sycholt, 2002). Snow does, however, fall more frequently between April and September and, on average, eight times a year (Tyson et al., 1976; Mulder and Grab, 2009) and it is usually restricted to the summit plateau and nearby slopes of the escarpment (Bussmann, 2006). It is estimated that snow contributes a water equivalent of 100 mm per annum above the escarpment (Nel and Sumner, 2005), although the exact contribution of snow to the water budget is unknown.

Above 3000 m, the climate supports marginal periglacial conditions, with a mean annual air temperature in the region of 5°C (Nel and Sumner, 2008). Temperatures are warmer in the eastern foothills, where fluvially incised valleys expose the sandstones and precipitation increases to around 1200 mm p.a. (Sumner and Nel, 2006). Recurrent droughts force plants to sustain water stress. The study area is thus a high elevation summer rainfall area, where winters are cold and mostly dry, with frequent frosts and snow on the mountains. Summers are warm, with regular thunderstorms that bring most of the high annual rainfall (Scott, 1993).

3.5 Precipitation

Precipitation is highly variable, given the complex topography of the Drakensberg. The two main rain-producing systems that bring summer rainfall to the Drakensberg are large-scale line thunderstorms and orographically-induced storms (Tyson et al., 1976). Winter rain is caused by the development of frontal systems, as closed low-pressure cells in the western Atlantic that move across southern Africa in a west-northwesterly to east-southeasterly direction (Tyson et al., 1976). Rainfall in the Drakensberg is seasonal and the summer months, November to March, account for 75% of the annual rainfall, and the four winter months, May to August, account for less than 10% (Nel and Sumner, 2006; Nel, 2009). Rainfall stations in the Drakensberg are known to experience 16-18 rain-days in the months of December and January (Tyson et al., 1976; Nel, 2008). Annual rainfall shows a cyclic variation of between 10 and 20 years (Nel, 2009).

Mean annual rainfall is estimated to vary between 800 mm in the north-east and south to over 1000 mm in the escarpment (Schulze, 1979; Nel and Sumer, 2008). Previous estimates of escarpment rainfall (1800 and 2000 mm) (Tyson et al., 1975; Schulze, 1979) have been challenged. Data collected from the southern and northern Drakensberg by Nel and Sumner (2005; 2008) suggest that these earlier estimates of rainfall totals on the escarpment may be an over-estimation. Sene et al. (1998) found that the escarpment edge receives about 1600 mm to less than 600 mm in the rain shadow zone, west of the escarpment. For example, Letseng-la-draai (3050 m a.s.l.), located in the rain shadow, receives 713 mm of rainfall per annumn, whereas Mokhotlong Town (2377 m a.s.l.) located about 35 km to the west of the escarpment, receives only about 575 mm of rain per annumn (Killick, 1978). Altitude is not necessarily the only important factor influencing rainfall in mountainous areas. A negative correlation between the altitude of the rainfall stations and the station's eastward distance from the escarpment was found in a study in the Drakensberg region (Nel and Sumner, 2006). The mean annual rainfall decreased by 54 mm for every 10 km from the escarpment. However, a significant relationship was found between station altitude and station distance from the escarpment in an eastward direction. It is unclear whether the two factors work together to affect rainfall totals in the Drakensberg, or if both altitude and the distance from the escarpment influence the spatial distribution of mean annual rainfall independently (Nel and Sumner, 2006).

Historical trends have shown that rainfall variability is increasing (Mason, 1996). An increase in the variability of the distribution of monthly rainfall, indicating increased seasonality, was highlighted in a study assessing the long-term rainfall trend in the KwaZulu-Natal Drakensberg (Nel, 2009). The strengthening of rainfall seasonality in the Drakensberg was identified through a significant decrease in autumn rainfall. The active crop-growing season in KwaZulu-Natal is October and changes to the seasonal regime could thus affect the late crop-growing season (Schulze, 1979; Stockil, pers.comm.). The El Nino Southern Oscillation influences the summer rainfall variability of the KwaZulu-Natal Drakensberg, suggesting that an increase in the frequency and intensity of ENSO events should negatively affect the rainfall in the Drakensberg (Nel, 2009). The early 1980s saw much of South Africa suffering from drought conditions, owing to low rainfall (Malherbe et al., 2015). In the Drakensberg area particularly, the farmers battled to grow crops when the local farm dams ran dry (Stockil, pers. comm.).

The correlation between high-intensity rainfall and thunderstorm activity in South Africa is strong. Frontal rainfall during winter gives rise to non-erosive, low-intensity drizzle, whereas erosive events, are linked to summer thunderstorms (Schulze, 1978; 1980). Thus, the seasonality of erosive events can be related to the source of precipitation (Nel and Sumner, 2008). There is a decrease in summer erosive rainfall and kinetic energy with altitude, which is possibly linked to the development of the orographic storms, where peak rainfall is reached before mazimum altitude (Nel et al., 2010).

3.6 Temperature

Generally, the southern Drakensberg is colder and drier than the northern portion (Bussmann, 2006). In the foothills (lower altitudes) of the Drakensberg, the mean monthly temperatures range from 20°C (January) to 10°C (July) and on the Lesotho plateau from 15°C (January) to 4°C (July) (Tyson et al., 1976). Mean annual temperatures along the summits are 5-6°C and seasonally range from approximately 10°C to 0°C (Grab, 2013). Frost days vary from 30-40 days in the foothills to >120 days in the high mountain valleys (Grab, 2013) thus supporting alpine flora and a marginally periglacial physical setting (Knight and Grab, 2015). At lower altitudes in the Central Drakensberg (1920 m a.s.l.), Sumner and Nel (2006) recorded air minimum temperatures of –3.3°C and mean air temperatures of 4.3°C over the winter of 2001.

Temperatures in headwater catchments, at and above the escarpment, are derived either as extrapolations from lower altitudes below the escarpment, or from relatively short-duration field monitoring (Nel and Sumner, 2008). Above 2600 m a.s.l., the environmental lapse rate is estimated to be between 2.6°C/km and 3.0°C/km (Grab, 1997a) and the mean annual air temperature (MAAT) above 2800 m is assumed to be in the region of 3°C to 7°C (Boelhouwers 1994; Grab, 1997b; 2002). The absolute minimum temperature exceeds -20°C at the highest altitudes. Therefore, periglacial conditions with thufur, needle ice, patterned ground, and soil movements are common at higher elevations (Grab, 1997b; Nel and Sumner, 2008).

3.7 Soil and Vegetation

The classification of soil types in the Drakensberg reveals a very dissimilar pattern, indicating that the soils have been formed under continually-changing conditions (Bester, 1988). In the Maloti-Drakensberg Park, Carbutt and Edwards (2015) found soils derived from nutrient-rich flood basalts to be high in organic matter and clay fractions. Small-scale studies at the Cathedral Peak station (Schulze, 1974; Granger, 1976) describe the Little Berg soils as ferrallitic (having a high iron content), structureless and acidic, due to a high degree of leaching, with a generally low availability of phosphates (Bester, 1988). Ferrallitic soils occur when rock is turned into clay, in conditions with high rainfall, through chemical weathering that rapidly breaks down the rock to form silica residues. The silica is subsequently removed through leaching, which releases iron, giving the soil its red colour (seen in the Elliot formation) and acidity (Bester, 1988; Werger, 1978). The soils, particularly those derived from basalt, are very acidic and are highly structured. Being volcanic in origin, they have a high mineral content (Irwin and Irwin, 1992). Low temperatures, high rainfall and steep topography result in nutrient leaching and a low pH content of soils. Carbutt and Edwards (2015) suggest that there will be an increased pool of nutrient-poor soils available under a warming regime.

Ecologically, the Drakensberg region is classified mostly as a grassland biome (Mucina and Rutherford, 2006), a classification that extends to the most high central plateau of South Africa, including KwaZulu-Natal, the Free State and the Eastern Cape (Sycholt, 2002). There are four bioregions in the grassland biome and the Drakensberg grassland bioregion is the one that is specific to this study. It the highest-lying bioregion on the Lesotho highlands and the immediate surrounds in KwaZulu-Natal, stretching southwards along the high-lying area of the escarpment in the Eastern Cape Province (Mucina and Rutherford, 2006). It is the grassland bioregion with the fewest number of vegetation types and has much lower temperatures, with a high incidence of frost, compared to the other grassland bioregions (Mucina and Rutherford, 2006). The summer rainfall grassland biome of the cooler, elevated interior is poorly represented elsewhere in Africa and is home to a wealth of species that are limited to southern Africa (Mucina and Rutherford, 2006).

Three altitudinal zones of vegetation are apparent in the KwaZulu-Natal Drakensberg region (Sycholt, 2002). Killick (1963) originally divided the vegetation into altitudinal zones that correspond closely with the physiographic features, namely, Montane (1280-1829 m), Sub-alpine (1829-2865 m) and Alpine (above 2865 m). In the Sub-alpine belt, grasslands dominated by *Themeda triandra* and various scrubs are common. The Alpine belt consists of heath communities (*Ericas and Helichrysum*), while grassland patches are generally broken up by frost action. The majority of the weather stations in this study are situated in the Montane vegetal belt. This belt extends from the valley floors to the lowermost basalt cliffs at the edge of the Little Berg. The greater part of this belt is occupied by Afro-montane grasses, such as tussock grassland (Bester, 1988; Sycholt, 2002). Species of *Protea* are scattered throughout the grassland to form Protea Savanna, in places (Bester, 1988). Examples of montane forests are Gudu Bush (Royal Natal National Park), Didima and Rainbow Gorges (Catherdral Peak Reserve).

O'Connor and Bredenkamp (1997) identified *Themeda triandra* and *Trachypogon spicatus*, as the dominant species for north- and south-facing slopes, respectively. Acocks (1975) provides a more specific description of the dominant species for the Drakensberg, as the tufted grass; *Themeda triandra* usually occurs in open grassland, whilst *Trachypogon spicatus* occurs on the woodier southern facing slopes. Major plant community differentiation in the area is usually related to climate (rainfall and temperature) and is associated with altitude. Although there are indigenous species in the area, these are now concentrated mainly in the reserves, with very few speicies remain outside of them.

3.8 Landuse and Landuse Change

The landuse within the Drakensberg area is related to altitude, with holiday resorts and conservation areas situated mostly above approximately 1500 m a.s.l. and farmland and small towns are found in the lower-lying areas (Nel and Sumner, 2006). The landuse in the central and northern Drakensberg consists of settlements, conservation, agriculture and tourism (Figure 3.3). The habitat types, including montane grassland, wetland, patches of dense forest and bushland, are increasingly being converted, on a large scale, for agricultural purposes (Ramesh and Downs, 2015).

Agriculture dominates this high rainfall region, with commercial farming, including livestock ranching and dairy production under private tenure, together with the use of indigenous grassland to support pastures, maize, plantation, forestry, as well as communal tenure (maize and rangelands) (O'Connor, 2005). The type, intensity, and success of the various agricultural practices relate to the past and prevailing political dynamics, whereby the areas dominated by (mostly white) commercial farmers exhibit successful commercial practices that vary between livestock farming, irrigated crops and afforestation (Zunkel, 2003). The necessary infrastructure for supporting farming activities has been developed, such as the building of dams/impoundments and irrigation systems (Zunkel, 2003).

Given the Drakensberg's high mean annual precipitation (MAP>800 mm), intensive agriculture is feasible (O'Connor, 2005). This relatively high rainfall ensures the viability of dryland crops, such as maize, potatoes and soybeans (Zunckel, 2003) and has resulted in the decline of livestock on rangeland, in favour of dairy farming and forestry (O'Connor, 2005). However, with the recent increasing variability in the rainfall (e.g. Nel, 2009), farmers can no longer completely rely on the rain (Stockil, pers. comm.). The indigenous grassland is grazed mainly by sheep and dairy cows from spring through to mid-summer (February) and by cattle to the early winter (May) (O'Connor, 2005). The pastures grown in summer last a long time (>30 years), as they are grazed and grown naturally and they are mown two to four times a season for hay. In winter, the annual pastures of perennial ryegrass occur mainly for the dairy cows, are usually irrigated and herbicides are used for weed control in maize and ryegrass (O'Connor, 2005).

In addition to the agricultural land use practices, this region has a well-established tourism industry (Zunckel, 2003). The boundary of the uKhahlamba Drakensberg Park (uDP), first inscribed on the list of World Heritage Sites in November 2000, was extended in 2013 to include the Sehlabathebe National Park in Lesotho (Ndlovu, 2017). The new transboundary World Heritage Site was named the Maloti-Drakensberg Park. It has one of the most richly-painted rock art areas south of the Sahara Desert, and it is among the most comprehensively researched areas in southern Africa (Ndlovu, 2017). The area has been a major centre for tourism in South Africa since the first half of the 20th century, owing to the outstanding natural beauty of its mountain flora and fauna and its exceptional richness in rock art (Duval and Smith, 2013). A variety of tourism markets exist, from upmarket resorts to camping facilities, local trading stores and adventure tours, which see an influx of people in the holiday season. Most people venture to enjoy the natural heritage, the spectacular scenery, hiking, climbing, fishing and other outdoor sports (Duval and Smith, 2013).

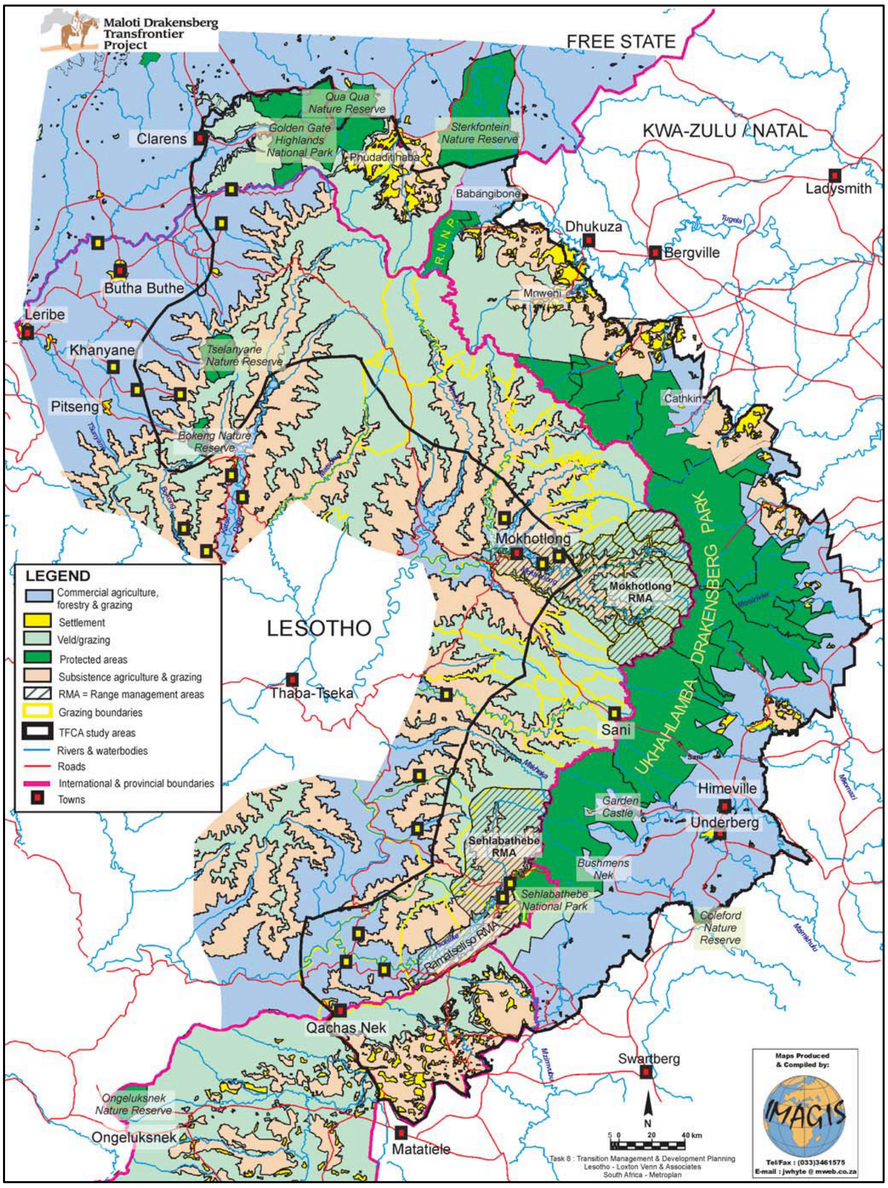


Figure 3.3 Map of the Maloti–Drakensberg Transfrontier Conservation and Development Area between Lesotho and South Africa, showing current land use (Sourced from Zunkel, 2003).

Overgrazing and related soil erosion is generally the most pressing issue in the sustainability of mountain landscapes along the escarpment, and the high Drakensberg sector in particular (Knight and Grab, 2015). Natural grasslands are being lost to exotic timber plantations, a form of extensive dryland cropping (Zunkel, 2003). Although this is currently happening on a small scale, with less than 5% of the surface area transformed thus far, stock theft and market dynamics are making afforestation a more attractive alternative for private landowners (Zunkel, 2003). The negative impact of this land use type is well documented and entails the complete transformation and subsequent loss of grasslands and the reduction of streamflow (Zunkel, 2003). Furthermore, the construction of new roads and dams, with the associated quarrying, and the need for more water storage facilities, leave many wetlands and mountain slopes permanently damaged (Knight and Grab, 2015). Additional planned developments include the building of new tourist lodges, the construction of wind turbines, a cableway from the foothills to the escarpment summit and additional diamond mining excavations, which will all negatively impact on the aesthetic landscape appeal (Knight and Grab, 2015).

A potential conflict of interest in the Drakensberg foothills arises from the demand of this area for agriculture and its importance as a water catchment (Walters et al., 2006). More than half the wetlands have also been destroyed, or eliminated, by various practices over the past 100 years, including cultivation (Sycholt, 2002). The wetlands are well-suited to agriculture because of the fertile soils and water availability (Walters et al., 2006). The delivery of high quality water in the Drakensberg is ensured, in part, by the wetlands that cover 5-20% of the area (Begg, 1989). When it rains, the water is stored underground in vleis and pans and gets released slowly throughout the year. Many of these water reserves are also being drawn off by exotic species of pine, blue gums and wattles that have been introduced for forestry and plantation purposes (Sycholt, 2002). Where these alien plants have been established, such as in the foothills of the central and southern Drakensberg, the grassland diversity has been destroyed, reducing the water flow of rivers in winter and reducing the ability of grassland to combat soil erosion (Sycholt, 2002). Invasive alien trees pose a threat to the natural habitat, by overtaking the indigenous plants and using a lot of water. This includes the spread of two alien wattle species (*Acacia mearnsii* and *Acacia dealbata*) in the rural parts of the Drakensberg region (de Neergaard et al., 2005).

3.9 Summary

The climate and rainfall of the KwaZulu-Natal Drakensberg represent a background of a summer rainfall dominated region, with altitudinal variations in rainfall and an increased strengthening of seasonality. There is, however, a limited knowledge of recent long-term rainfall trends in the area, which further makes rainfall variability trends unclear. Owing to this lack of knowledge, no analysis has explored the occurance of drought years and the subsequent affects on drainage and discharge in the region. In particular, it has also been noted that rainfall variations at a valley scale (over shorter distances) are also unknown.

4. WEATHER STATION DATA AND METHODS

4.1 Introduction

When compared to other African countries, South Africa has a good historical rainfall record, which is useful when studying the trends of rainfall over time, for example, over multiple decades (MacKellar et al., 2014). However, the country is characterized by high variability in terms of climate and this extends to rainfall, which contributes to the complexity of long-term trends and changes (MacKellar et al., 2014). Depending on the area of study in South Africa, rainfall data is available, either from the South African Weather Service (SAWS), or from privately run rainfall stations, including farmers and home owners, and this data can sometimes extend back to the late 1800s (Goodall and Ranwashe, 2012). Alternatively, there are nationwide rainfall databases from which to access rainfall data, including the South African Environmental Observation Network (SAEON) (SAEON, 2017), WeatherSA (2017) and the South African government websites, such as the Department of Water and Sanitation (Department of Water and Sanitation, 2017).

This study is conducted at two spatial scales; the first is across the central and northern Drakensberg and the second is at a catchment scale within the Drakensberg area containing the Sterkspruit River. Station data for the central and northern KwaZulu-Natal Drakensberg region were obtained from the South African Weather Service. In assessing the temporal trends of rainfall and rainfall variability in the foothills of the Drakensberg, this study only used data from well-established SAWS stations, with long-term rainfall measurements covering most of the last century and providing a good geographical coverage of the central and northern study region. The Sterkpruit Catchment rainfall data were obtained from local farmers and homeowners in the catchment area who had access to long-term and reliable data. This area of localized study was chosen as it is where the Bell Park Dam is situated. The overall study area is found in the greater drainage region of the Tugela River.

4.2 Central and Northern Drakensberg Data

Data for the stations in the Drakensberg were obtained from five South African Weather Stations. In part, this study is building on the research of Nel (2009), who used 11 SAWS rainfall stations, however, some stations have been discontinued and updated records were unavailable for all 11 stations used in Nel's (2009) study. Thus, reliable data sets were obtained from the SAWS for five stations located in the central and northern Drakensberg. Figure 4.1 illustrates the geographical location of the stations and Table 4.1 highlights the specific historical records, co-ordinates, altitudes and the different station types.

Table 4.1 Information on the selected rainfall stations in the central and northern KwaZulu-Natal Drakensberg.

Rainfall Station	Historical record	Latitude (S)	Longitude (E)	Altitude (m.a.s.l.)	Type of rainfall gauge
Cavern	1947 – 2015	28° 38′	28° 58′	1980	Manual
Royal Natal National Park (Royal Natal)	1948 – 2015*	28° 41′	28° 57′	1392	Automated (1988)
Bergville	1934 – 2015	28° 44′	29° 21′	1128	Manual
Cathedral Peak	1941 – 2015	28° 57′	29° 12′	1448	Manual
Giant's Castle	1948 – 2015	29° 16′	29° 31′	1754	Automated (2006)
*Data missing in 1959, 1966, 1979, 1989					

At some of the stations, daily rainfall is measured manually from a raingauge at 08h00, by SAWS volunteers, including farmers, police officers and conservation personnel. However, the Giant's Castle and Royal Natal National Park's rainfall gauges are now automated (Table 4.1). This means that the station captures, calculates, logs and saves meteorological data automatically, which allows for it to be accessed electronically. They are often run off batteries that are powered by solar panels. In earlier reports, a 95-98% accuracy was noted for station data (Schulze, 1979; Dent et al., 1990). Photographs of each station can be seen in Figure 4.2.

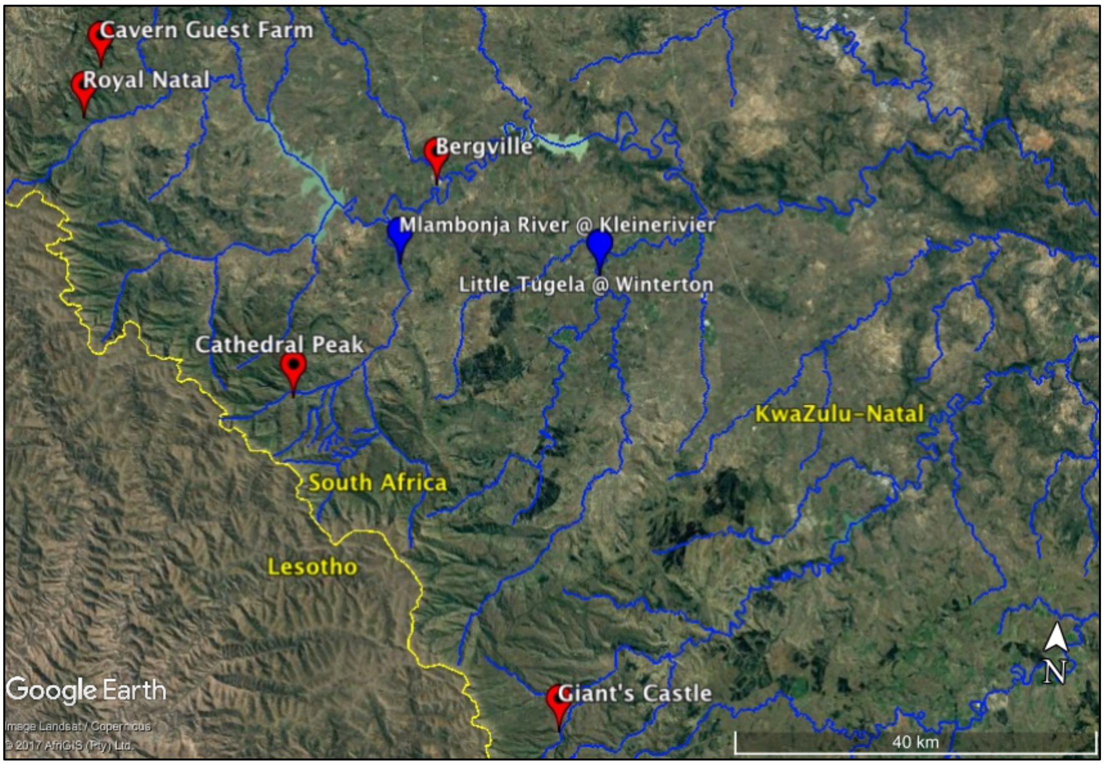


Figure 4.1 Location of the rainfall stations in the central and northern Drakensberg (Author's own, Google Earth 2017).

Figure 4.2 Photographs of the rainfall gauges used in the central and northern Drakensberg. Top row: Cavern, Cathedral Peak, Giants Castle; Bottom row: Royal Natal National Park, Bergville. (Photos courtesey of: Cavern Drakensberg Resort, Cathedral Peak Hotel and South African Weather Service, respectively).

4.3 Sterkspruit Catchment Data

The location of the Sterkspruit Catchment, in relation to the central and northern Drakensberg rainfall stations, is seen in Figure 4.3. This catchment is of the Sterkspruit River is found in the Tugela River drainage region 'V', and in the quaternary catchment V13B. The Bell Park Dam is situated in this quaternary catchment. The Sterkspruit River starts in the highlands and connects to the Little Tugela River near Winterton (Figure 4.3). The Little Tugela is a tributary of the Tugela River and thus continues downstream, to join the Tugela River. The rainfall stations chosen for the study provide a distribution along the Sterkspruit Catchment, except for one farm, Reeves, which lies 11 km downstream of the catchment boundary. Four rainfall stations were selected to represent the area and Table 4.2 provides information specifically relating to each station.

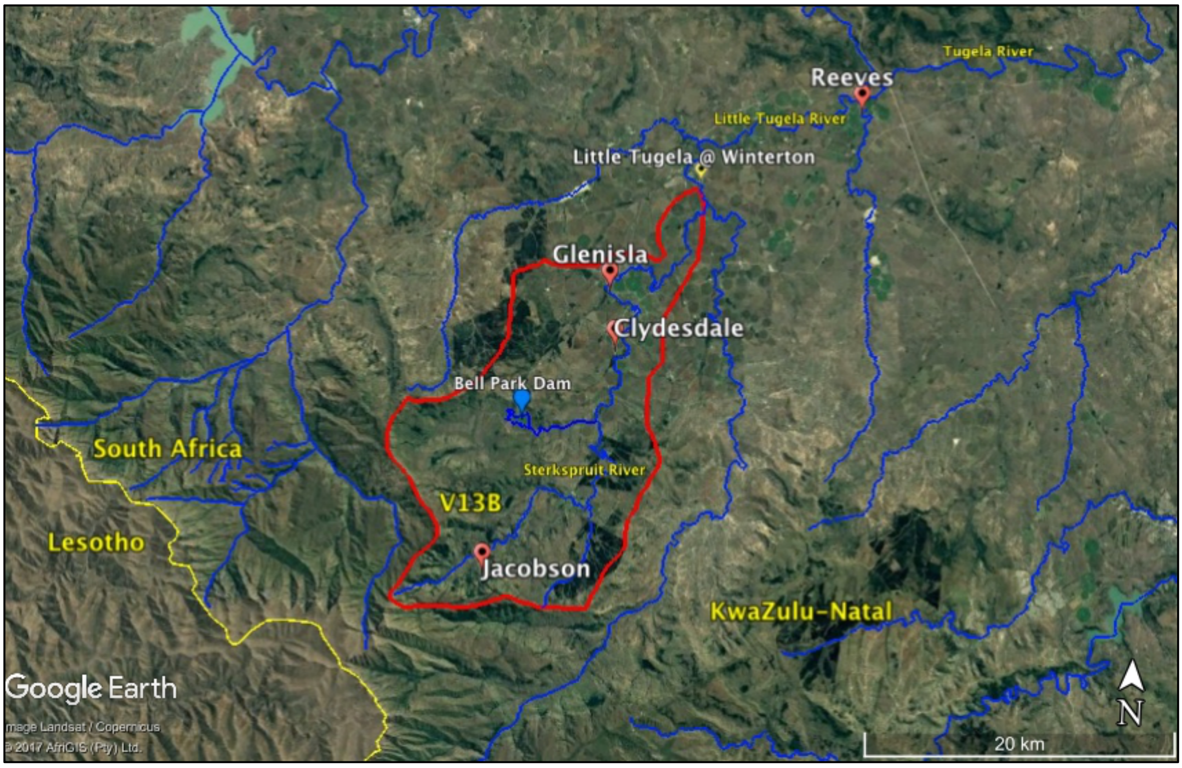


Figure 4.3 Location of the rainfall stations in the Sterkspruit Catchment (V13B) and discharge station (Authors' own, Google Earth 2017).

Table 4.2 Spatial distribution of selected rainfall stations in the Sterkspruit Catchment, KwaZulu - Natal.

Rainfall station	**Historical record**	**Latitude (S)**	**Longitude (E)**	**Altitude (m a.s.l.)**	**Type of rainfall gauge**	**SAWS/ Land owner**
Reeves	1957 – 2015	28° 46′	29° 38′	990	Manual	Land owner
Glenisla	1974 – 2015	28° 52′	29° 29′	1077	Automated (2015)	SAWS
Clydesdale	1968 – 2015*	28° 55′	29° 24′	1135	Manual	Land owner
Jacobson	1966 – 2015	29° 02′	29° 29′	1443	Manual	Land owner
*Data missing from 2006 – 2010						

Glenisla Farm is a third-generation family farm that produces dairy and maize. The rainfall gauge has been situated in the same open area since the establishment of the farm and measurements are recorded at the same time each day (Figure 4.4). Mr Reeve’s rainfall gauge is secured to his gatepost, in the open, with no obstructions and is measured at the same time every morning (Figure 4.4). It must be noted that the data from 1956-1963 were obtained from a neighbouring farm 1 km away, which had already been established prior to Reeves’ farm. Clydesdale farm is currently in its third generation of farming, producing mainly beef cattle, soya and maize, and wheat is grown in the winter. The rainfall gauge has remained in the exact same position since the beginning of the available rainfall data in 1968 (Figure 4.4). The rainfall is checked daily at 08h00 and recorded by the farmer. Dr Jacobson is a homeowner, whose property is situated just below Monks Cowl. The rainfall gauge is perfectly situated in the open and is measured and recorded every morning at 07h00 (Figure 4.4).

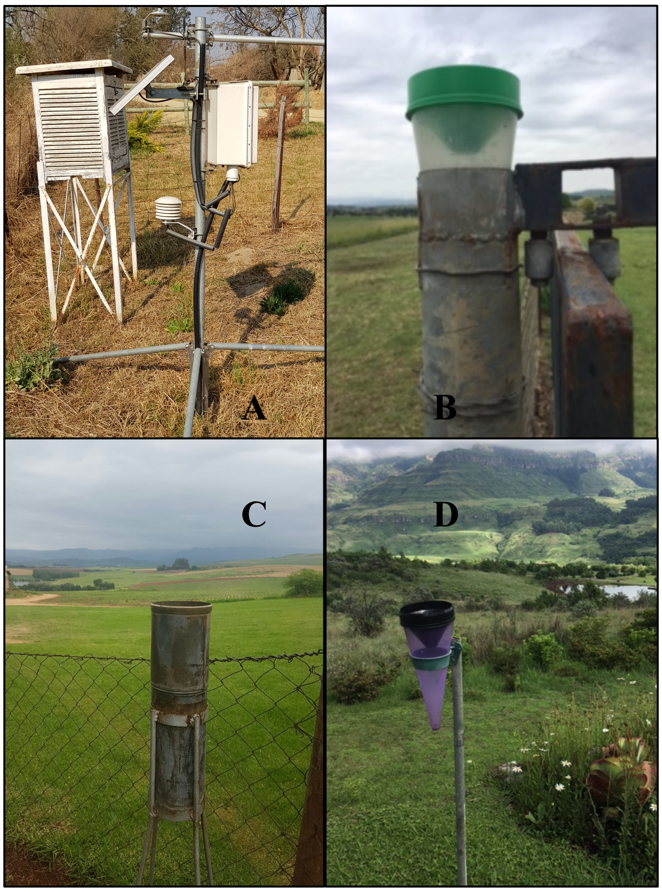


Figure 4.4 Rainfall stations photographed on-site at (a) Glenisla (b) Reeves (c) Clydesdale (d) Jacobson (Photographs: Catherine Smart, 2016).

4.4 Discharge Data

Discharge data for the wo stations in the Drakensberg were obtained from the website of the Department of Water and Sanitation (DWS) (http://www.dwa.gov.za/Hydrology/Verified/hymain.aspx: Accessed August 2016). The stations and the data are run and collected by the DWS and the data are presented as monthly volumes in million cubic meters. The data selected for this study consisted of the surface water level data at discharge stations in two rivers in the central and northern Drakensberg, namely, the Little Tugela River at Winterton and Mlambonja River at Kleinerivier. These two stations were chosen as they are the nearest to the Sterkspruit River.

The Little Tugela River at Winterton, Station Number V1H010, has a catchment area of 782 km^2, and discharge data available from 1964 to the present, and it is situated at the bottom of the quaternary catchment area V13B (28° 49' 5.27" E and 29° 32' 42.18" S)

(Figure 4.3). The Mlambonja River Station Number V1H041 has a catchment area of 434 km^2, and data available from 1976 to the present, and is situated within the Drakensberg study area (28° 48’ 45.4”E and 29° 12’ 38.19” S). These stations will give an indication of how the streamflow of the rivers respond in times of low and high rainfall in the Drakensberg.

4.5 Methodology

This research has followed an inductive thread of reasoning, however, some aspects of the analysis are more deductive, due to the quantitative nature of working with rainfall data (Elo and Kyngäs, 2008). The data collected in this research is secondary data, namely, from rainfall records that have already been collected. There are challenges, when conducting a desktop analysis of secondary data, and it is important to ascertain the reliability of the rainfall and climate data to be used (WFP, 2009).

The methods are, for the most part, carried out at two scales, namely, in the general central and northern Drakensberg region and at a catchment scale within this area, the Sterkspruit Catchment. The shared methods include the trends in interannual rainfall variability, intra-annual variability and seasonal variability, the Southern Oscillation and rainfall, the discharge patterns, as well as the identification and characterization of meteorological and hydrological droughts. A detailed account of data analysis can be found below.

While Nel’s (2009) study included analysis for the period from 1955-2000, the current data analysis for the Drakensberg study area covers the period from 1955-2015. The data analysis for the catchment stations is carried out individually, from their historical starting date of data capture, until 2015. A grouped analysis of the stations is carried out from 1974-2015, where data is consistently available for all the stations. A secondary timeframe of analysis is carried out in both sections for the the period 2000-2015, in order to ascertain if any changes have contributed to the overall long-term trends and patterns this century. The methods are identical in both the Drakensberg and Sterkpruit Catchment scale study areas. A statistical analysis, regarding correlation statistics and linear regressions to find r and P values, was computed using a free online program called JASP Statistics (Available: free download: https://jasp-stats.org). All the other analyses were computed using Microsoft Excel 2010.

4.5.1 General rainfall trends

The general rainfall trends were first established for the area, where the average monthly rainfall distribution was plotted and the average seasonal rainfall (summer, autumn, winter and spring) totals were calculated. The mean annual rainfall of the stations was then plotted against their altitudes and a linear trend was applied, to ascertain the relevant degree of significance. Linear regressions are used to ascertain if there is a trend within the data. The degree of significance used throughout the study is where the P value falls within the 95% confidence level ($P \leq 0.05$).

4.5.2 Trends in interannual rainfall variability

To test for trends in annual rainfall, the data at each station in the Drakensberg region for the period 1955-2015 was used to calculate an arithmetical mean for the region and this mean for each year was plotted. This was carried out across the stations in the catchment area for the period 1974-2015. The yearly rainfall totals for each station were averaged and plotted, showing the mean annual rainfall at each station. Linear regressions were applied, to discern the stations with a degree of significance. Each station in the Drakensberg catchment study area was plotted individually, displaying the trend in annual rainfall over time and linear regressions were again applied. An analysis was also carried out at both scales for the period 2000-2015 in order to ascertain if the last 16 years contributed significantly to the overall trends.

The absolute deviation of annual rainfall from the mean annual rainfall (absolute deviation) was used to analyse the long-term trends in interannual rainfall variability at each station for the periods 1955-2015 and 1974-2015. A linear regression was applied to all data to discern any temporal trends with a related degree of significance. An analysis was also carried out at both scales for the period 2000-2015, in order to ascertain if the last 16 years contributed significantly to the overall trends. The rainfall deviation from the mean annual rainfall was also analysed for the two study areas and plotted in a bar graph, to discern which years had above or below average rainfall, and it was tested for a degree of significance. However, these values are the same as the MAR statistical values, thus the results can be found in Appendix A.

4.5.3 Trends in intra-annual rainfall and seasonal variability

4.5.3.1 Precipitation Concentration Index

Rainfall in the Drakensberg is seasonal and the seasonality can be described through the monthly rainfall totals as a percentage of the total amount of rainfall (Nel and Sumner, 2006). A modified version of the Precipitation Concentration Index (PCI) was applied to define intra-annual variability and to quantify its temporal and spatial trends (de Luis et al., 2000; Ceballos et al., 2004):

$$\mathrm{PCI} = 100 \frac{\sum_{i=1}^{12} P_i^2}{\left(\sum_{i=1}^{12} P_i\right)^2}$$

where Pi is the precipitation of the month, *i.* Values below 10 indicate a uniform distribution of rainfall throughout the year. PCI values from 11 to 20 indicate a seasonal trend and values above 20 indicate a considerable variability of the distribution of monthly rainfall (Ceballos et al., 2004). Therefore, an increase in the PCI value over time indicates an increase in the variability of the distribution of monthly rainfall. The PCI was calculated for each station from 1955-2015 and from the relevant years in the catchment, until 2015 to analyse the long-term trend in intra-annual rainfall variability. A linear regression was applied to the station data to quantify any temporal or spatial trends with the related degree of significance at the 95th confidence level. An analyse was also carried out at both study areas for the period 2000-2015, in order to ascertain if the last 16 years contributed significantly to the overall trends.

4.5.3.2 Seasonality

An increase in the PCI values indicates an increase in the seasonality of monthly rainfall. The average summer (November–March), winter (May–August), spring (September–October) and autumn (April) rainfall were measured at each station in the Drakensberg and catchment over the relevant periods, to assess the seasonality and to quantify the changes in seasonal rainfall. The classification of seasons is done, based on the hydrological year, as the rainy seasons run through the calendar year (Nel, 2009). The onset or end of a rainfall season, either on average or in individual years, seldom coincide with the calendar months (Schulze, 1979; Nel, 2009). The seasons are standardized to allow for consistency throughout the study. For this study, only temporal trends are analysed and this delimitation is purely a functional one.

The years and each season were plotted and trend analyses was carried out to ascertain which seasons are significant to the 95th confidence interval ($P \leq 0.05$). The average seasonal rainfall over time was also plotted for each season (summer, winter, autumn and spring) for both the Drakensberg and catchment stations. Liner regressions were applied, to assess the degree of significance. An analyse was carried out at both scales for the period 2000-2015, in order to ascertain if the trends this century contributed significantly to the overall trends.

4.6 Southern Oscillation Index and Summer Rainfall

In order to investigate the effect that the ENSO has on summer rainfall totals in the KwaZulu-Natal Drakensberg, the regional summer rainfall (November to March) has been determined annually for the period 1955-2015, as a five-station arithmetical mean. The SOI was calculated with the method used by the Australian Bureau of Meteorology is the Troup SOI, which is the standardized anomaly of the Mean Sea Level Pressure difference between Tahiti and Darwin (Troup, 1965). The SOI data were retrieved from the Internet on the 24 August 2016 from the Australian Bureau of Meteorology's website: http://www.bom.gov.au/climate/current/soihtm1.shtml. Allan et al. (1997) compared the Troup and several other SOI series, and it was concluded that they are all very highly correlated (Chiew et al., 1998).

Non-lag correlations of the linear regressions of the rainfall against Troup SOI were calculated to quantify the strength of the Southern Oscillation index on rainfall (Chiew et al., 1998; Nel, 2009). The correlations are calculated for the total rainfall amounts in the summer season, using the actual data values. To test if there is a non-lagged relationship between summer rainfall and the SOI, the average SOI from November to March was compared with the summer rainfall (November–March), over the same period in the Drakensberg study region (see e.g. Nel, 2009). For forecasting the Australian seasonal rainfall in some regions and seasons, it has previously been found that lagged indices of the Southern Oscillation are useful (McBride and Nicholls, 1983; Stone et al., 1996; Chiew et al., 1998).

The Pearson Correlation parametric test was applied, to test the strength of the relationship with the related degree of significance. To forecast early summer rainfall from SOI, Hyden and Sekoli (2000) assume that one month is needed to model the forecast, to distribute the results and to act on them. Therefore, the SOI was computed for the preceding periods lagged by at least one month. A similar methodology was used to test if there was a lagged relationship between the SOI and the summer rainfall in the Drakensberg and to see if this correlation was strong enough to be used as an indicator for seasonal forecasting. All lagged correlations were tested for significance, where $P \leq 0.05$. This analysis was also conducted at the catchment scale.

4.7 Identification and Characterization of Meteorological Droughts

4.7.1 Standard Precipitation Index (SPI)

The Standardized Precipitation Index (SPI) was chosen for the analysis of meteorological droughts. This index for, evaluating a precipitation deficit, was suggested by McKee et al. (1993). The calculation of SPI accommodates data from a large number of stations and can be used for regional drought analysis. The nature of the SPI allows one to determine the rarity of a drought or an anomalous wet event, at a particular time scale, for any location in the world that has a precipitation record (Cheval, 2015). Thus, the SPI was designed to quantify the precipitation deficit for multiple timescales and it allows the monitoring of the intensity and spatial extent of droughts at different timescales.

The SPI has been favourably reviewed by Keyantash and Dracup (2002), in comparison to other indices, and has been integrated into the indices used by the Drought Monitor in the USA. Rouault and Richard (2005) show how the SPI can be used operationally to detect the start of a drought, its spatial extent and its temporal progression, and that it can be used worldwide because it is linked to percentage occurrence and is only based on rainfall (Rouault and Richard, 2003). The theoretical framework of the SPI described by McKee et al. (1993, 1995) was substantially enhanced in the years that followed. Edwards (1997) and

Guttmann (1999) provided a detailed methodology for its computation (Cheval, 2015).

This current study adopts the method, based on work done by Rouault and Richard (2005), who used the SPI to do a retrospective analysis of the spatial extent of droughts in southern Africa, to determine the extent of meteorological drought. They showed the potential of using the SPI for South Africa. The SPI is consistent with regards to the spatial distribution of rainfall that occurs with great variability in southern Africa, due to the geographical location, orography and ocean influence (Rouault and Richard, 2005). The timing of maximum precipitation, or the mean annual precipitation, and onset of the rainy season, also vary greatly in southern Africa and within South Africa itself.

Hayes et al. (1999) showed that, for some regions, a good rainfall for one month can create the impression that the drought is over, but if the SPI values are not above a certain value (typically -1) at all scales, a drought will still affect a region, one way or another. However, during the dry season, large negative or positive SPI values may be associated with precipitation totals that are not very different from the mean; and because this is a period with little rain, these mean totals will be small, and relatively small deviations on either side of the mean could cause large negative or positive SPI values (Hayes et al., 1999). Nevertheless, it is important to have a drought index for longer timescales. A normal three-month period could occur in the middle of a longer-term drought and will only be visible at a longer time scale (Vicente-Serrano et al., 2010). Monthly precipitation totals are normally used for the calculation of the SPI (Vicente-Serrano et al., 2010).

4.7.2 Methodology to estimate SPI

In this study, the SPI was calculated on a 12-month time scale that corresponds to the past 12 months of observed precipitation totals. The 12-month period gives an indication of where prolonged droughts exist, or where below-normal rainfall has occurred over a period of one year. Initially, the variability of precipitation totals is described by gamma distribution, and it is then transformed to a normal distribution (McKee et al., 1993; Edwards, 1997). All analyses of the SPI were carried out using the Microsoft Excel 2010 program. The method of Edwards (1997) is followed (see Appendix B).

The SPI is a dimensionless probability index, and its computation for a given location and period was conducted in the following sequence:

(a) Data sets were fitted to a probability density function (PDF). The selection of the parametric distribution determines the accuracy of the SPI values. McKee et al. (1993) recommended gamma distribution, which is followed in this study. The gamma distribution is defined by its frequency or probability density function. The computation of the SPI involves fitting a gamma probability function to a given frequency distribution of precipitation totals for a station. The alpha and beta parameters of the gamma probability density function are estimated for each station for the given timescale (12 months) and for each month of the year.

(b) The resulting parameters are then used to find the cumulative probability of an observed precipitation event for the given month and the timescale for each station (Edwards, 1997).

(c) Further, the cumulative probability distribution is transformed into a standard normal distribution, with a mean of zero and variance of one, which is the SPI value (Sönmez et al., 2005; Angelidis et al., 2012). The SPI is normalized to a station because it accounts for the frequency distribution of precipitation, as well as the accompanying variation at the station. In addition, the SPI is normalized in time, because it can be computed at any number of timescales, depending upon the impacts of interest to the analyst. No matter what the location or time scale, the SPI represents a cumulative probability, in relation to the base period for which the gamma parameters were estimated (Edwards, 1997).

4.7.3 Standard Precipitatin Index (SPI) Analysis

SPI values were plotted against time for the period 1955-2015, for the Drakensberg stations and from the individual data starting dates up to 2015, for the catchment stations. This allows for the identification of wet and dry years over a 60-year period. Linear regressions were applied, to ascertain the relevant degree of significance. The percentage of the occurrence of each drought category was then calculated and tabulated with the SPI data, in order to ascertain how frequently each category of drought occurred. A separate analysis was conducted over the period 2000-2015 at both scales, which separates the recent droughts. This was done to contextualize the current century.

The SPI values were then compared to the drought intensity categories, as shown in Table 4.3. Various drought intensities have been defined, in order to describe the rainfall conditions (Mckee et al., 1993; Komuscu, 1999; Agnew, 2000). In the present study, the classification for SPI values by McKee et al. (1993), as adapted by Hayes et al. (1999), is used to define the drought conditions. Table 4.3 shows the classification of wet and dry events. An SPI value of -2 or more indicates an extremely dry period and an SPI value of +2 or more indicates extremely wet conditions.

Table 4.3 SPI Classification for wet and dry rainfal periods (McKee et al., 1993; Hayes et al., 1999)

SPI Value	Drought category
> 2.00	Extremely wet
1.5 to 1.99	Very wet
1 to 1.49	Moderately wet
- 0.99 to 0.99	Near normal
- 1.00 to -1.49	Moderately dry
- 1.50 to -1.99	Severely dry
< - 2.00	Extremely dry

4.8 Discharge

4.8.1 River discharge in relation to rainfall

Surface water level data from two rivers, the Little Tugela at Winterton and the Mlambonja River at Klienerivier, were used to statistically compare years of low rainfall with the corresponding discharge in the two rivers. The analysis assumes that the flow and water level records are free from problems caused by sediment accumulation, erosion and the local changes of gauge sites (Marengo et al., 1998).

Yearly discharge was calculated from the monthly data, in order to allow for a time series trend analysis. The overall discharge over time was plotted for both stations and linear regressions were run, to determine the statistical significance (Chiew et al., 2003). The trends for 2000-2015 were also analysed for each station. The discharge of the stations was plotted against the yearly total summer rainfall for each year for the Drakensberg stations and the catchment stations. Linear regressions were then run, to test for the relative degree of significance (Chiew et al., 2003). Lastly, the overall average discharge for both stations was compared to the summer (November–March) and winter (May–August) discharges and the statistical significance was tested using a linear regression.

The average discharge data for both stations was then correlated, to ascertain the relative degree of significance between rainfall and discharge, at both the Drakensberg and catchment scale. The overall average Drakensberg rainfall was computed against the average summer and winter discharges, the Drakensberg summer rainfall was computed against the summer discharge and the Drakensberg winter rainfall was computed against the winter discharge. Comparisons were made by looking at the 'r' and P values where $P \leq 0.05$.

4.9 Hydrological Drought

4.9.1 Streamflow Drought Index (SDI)

Based on the SPI developing concepts, the SDI was developed by Nalbantis and Tsakiris (2009) for characterizing hydrological drought. The index shows the standardized cumulative streamflow volume at a selected timescale. Analysis of the streamflow data from the two discharge stations, the Little Tugela at Winterton and the Mlambonja at Kleinerivier, occurs over a timescale of 12 months. All analyses were computed using Microsoft Excel 2010.

Firstly, the sum of the discharge data over a 12-month period was calculated, using the hydrological year for South Africa, which runs from November through to October (Nel, 2009). Analogous to the calculation of SPI, the SDI is computed by firstly fitting a suitable probability distribution function to the frequency distribution of aggregated streamflow data, summed over the time scale of interest (Nalbantis and Tsakiris, 2009; Hong et al., 2015). In order to do this, the appropriate distribution is needed and several studies have recommended using the log-normal distribution to fit the streamflow data (Rimkus et al., 2013).

Following this reccommendation, the data was subsequently transformed into a logarithmic dataset. The SDI was then computed, by subtracting the average of the streamflow transformed data from the transformed streamflow data for each year, and it was then divided by the standard deviation of the transformed data. The SDI values obtained were then compared to the relevant drought categories, as described by Nalbantis and Tsakiris (2009) in Table 4.4. Positive SDI values reflect wet conditions (not shown), while negative values indicate a degree hydrological drought. The probability of each type of drought occurring for each of the two stations was calculated, to see how often a specific drought type occurred over the years.

Table 4.4 Definition of states of hydrological drought, based on SDI (Nalbantis and Tsakiris, 2009).

SDI Values	Description
SDI≥0.0	Non-drought
−1.0≤SDI<0.0	Moderate drought
−1.5≤SDI<−1.0	Mild drought
−2.0≤SDI<−1.5	Severe drought
SDI> −2.0	Extreme drought

The dry years are then noted and used for comparison against the rainfall trends for the two areas (the catchment and the Drakensberg) and compared against the meteorological drought analysis, to confirm which years were dry and how long and how often these dry years occurred. The analysis was also conducted over the period 2000-2015.

5. RESULTS

5.1 Introduction

This chapter has three broad sections, which consists of the rainfall trend analysis, drought analysis and hydrology of the Drakensberg. The former two sections are carried out at both the central and northern Drakensebrg scale and the Sterkspruit Catchment scale and the hydrology is generalized for the Drakensberg region. The rainfall analysis is comprised of the establishment of the rainfall attributes of the area, followed by the presentation of the rainfall and seasonality trends and the influence of the El Nino Southern Oscillation. The second section comprises an analysis of meteorological droughts in the area and thirdly, an investigation into the hydrology of the area, to establish the discharge patterns and the occurrence of hydrological drought.

The time period of analysis is specific to the two spatial areas of the study. The central and northern Drakensberg rainfall analysis is conducted over the period 1955-2015 at all stations. The time period of analysis at the Sterkspruit Catchment scale is conducted over the period 1974-2015 for a grouped station data analysis, becasue 1974 is the earliest concurrent year across the stations. An analysis of the individual station data at the Sterkpruit Catchment is conducted from the earliest year from when the rainfall measurements commenced, until the year 2015.

In order to identify any trends occurring this century a trend analysis was also undertaken for the period 2000-2015 at both scales. Isolating the trends that have occurred this century, helps to provide a comparison with Nel's (2009) work, which was undertaken from 1955 to 2000. Thus, extending the analysis up to 2015 will help to establish if any recent changes in rainfall trends are contributing to an overall change in rainfall variability in the area.

5.2 Central and Northern Drakensberg

5.2.1 General rainfall attributes

The rainfall regime in the Drakensberg study region is classified by the hydrological year, namely, November–March (summer), April (autumn), May–August (winter) and September–October (spring). From the data of the five stations, it is clear that this is a summer rainfall region, with 75.22% of the rainfall occurring in summer, 13.68% in spring, 6.54% in winter and 5.28% in autumn (Figure 5.1). A summary of the stations' mean annual rainfall and average seasonal rainfall is provided in Table 5.1.

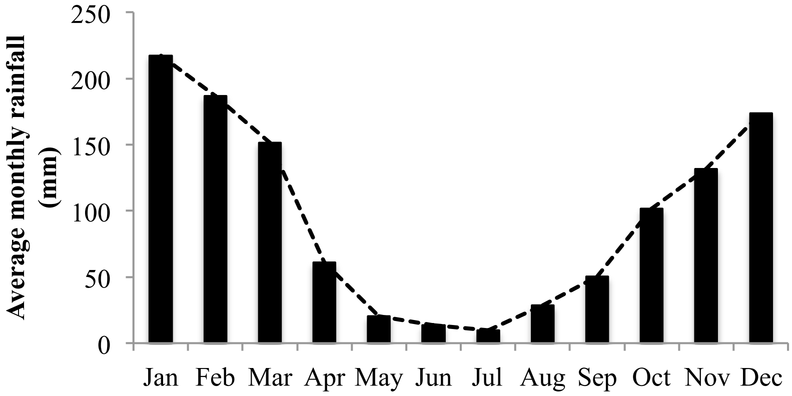


Figure 5.1 Average monthly rainfall average for five combined stations in the Drakensberg for the period 1955-2015.

Table 5.1 Summary of results for the Drakensberg stations for the period 1955-2015.

Rainfall Station	Altitude (m.a.s.l.)	MAR (mm)	Average summer rainfall (mm)	Average winter rainfall (mm)	Average autumn rainfall (mm)	Average spring rainfall (mm)
Cavern	1980	1338	201	23	65	90
Royal Natal	1392	1274	192	20	67	82
Bergville	1128	757	114	13	43	49
Cathedral Peak	1448	1216	188	20	68	76
Giant's Castle	1754	931	140	15	51	63

The Cavern station has the highest mean annual rainfall, followed by the Royal Natal station and Bergville has the lowest mean annual rainfall (Table 5.1). Rainfall over the Drakensberg stations increases with the increasing altitude; however, this relationship is not significant to the 95% confidence level (Figure 5.2).

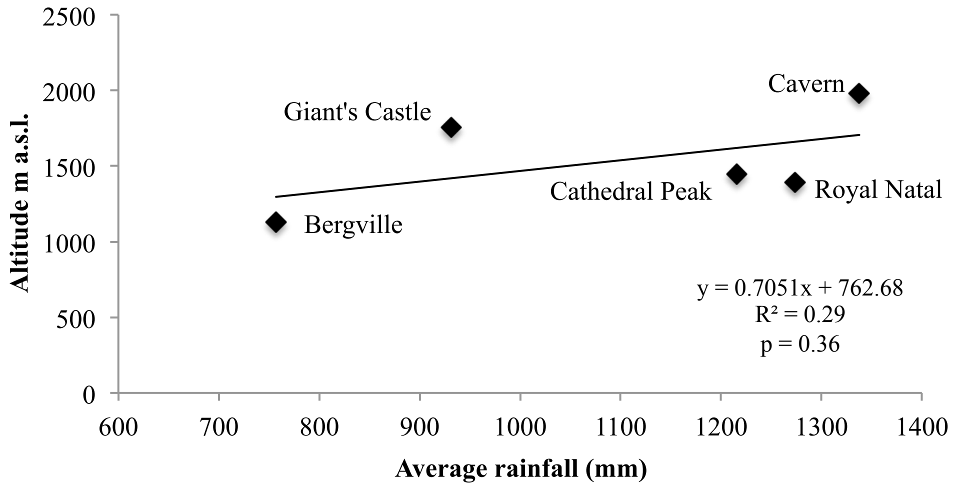


Figure 5.2 Average rainfall for each station in relation to altitude for the period 1955-2015.

5.2.2 Trends in interannual rainfall variability

The time series of annual rainfall for the five individual stations is depicted in Figure 5.3. All stations, except Bergville, show a decreasing trend in rainfall over time and the trend at Giant's Castle, Cathedral Peak and Royal Natal is statistically significant. Giant's Castle has a highly significant decreasing trend ($P<0.001$) and a rainfall cycle that averages 14 years between rainfall peaks. Cathedral Peak displays a rainfall cyclicity, averaging 10.5 years, and Royal Natal has a rainfall cyclicity occurring approximately every 12 years between rainfall peaks. All stations show dry years around 1980, 1982, 1992, 1994, 2003 and 2007. All stations, except Giant's Castle, depict a dry year in 2015. The minor peaks in rainfall across all the stations occur, on average every three years.

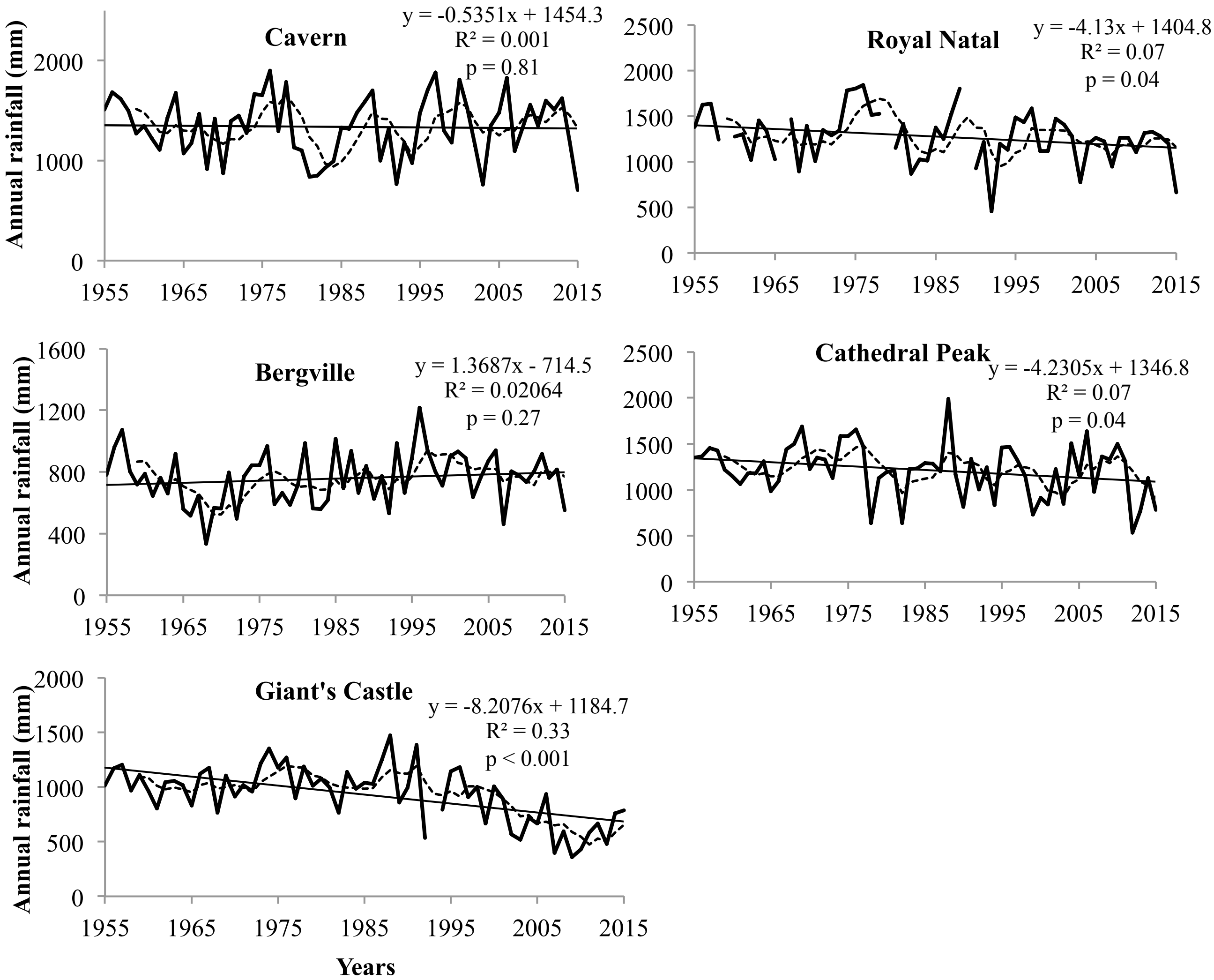


Figure 5.3 Time series of annual rainfall for five stations in the Drakensberg for the period 1955-2015 showing the overall trend (solid line) and the 5-year moving average (dashed line).

The analysis of the annual rainfall over the period 2000-2015 showed no change in the overall rainfall trends for any of the stations, except at Bergville, where the overall average rainfall decreases (Figure 5.4). None of these trends are significant to the 95% confidence level.

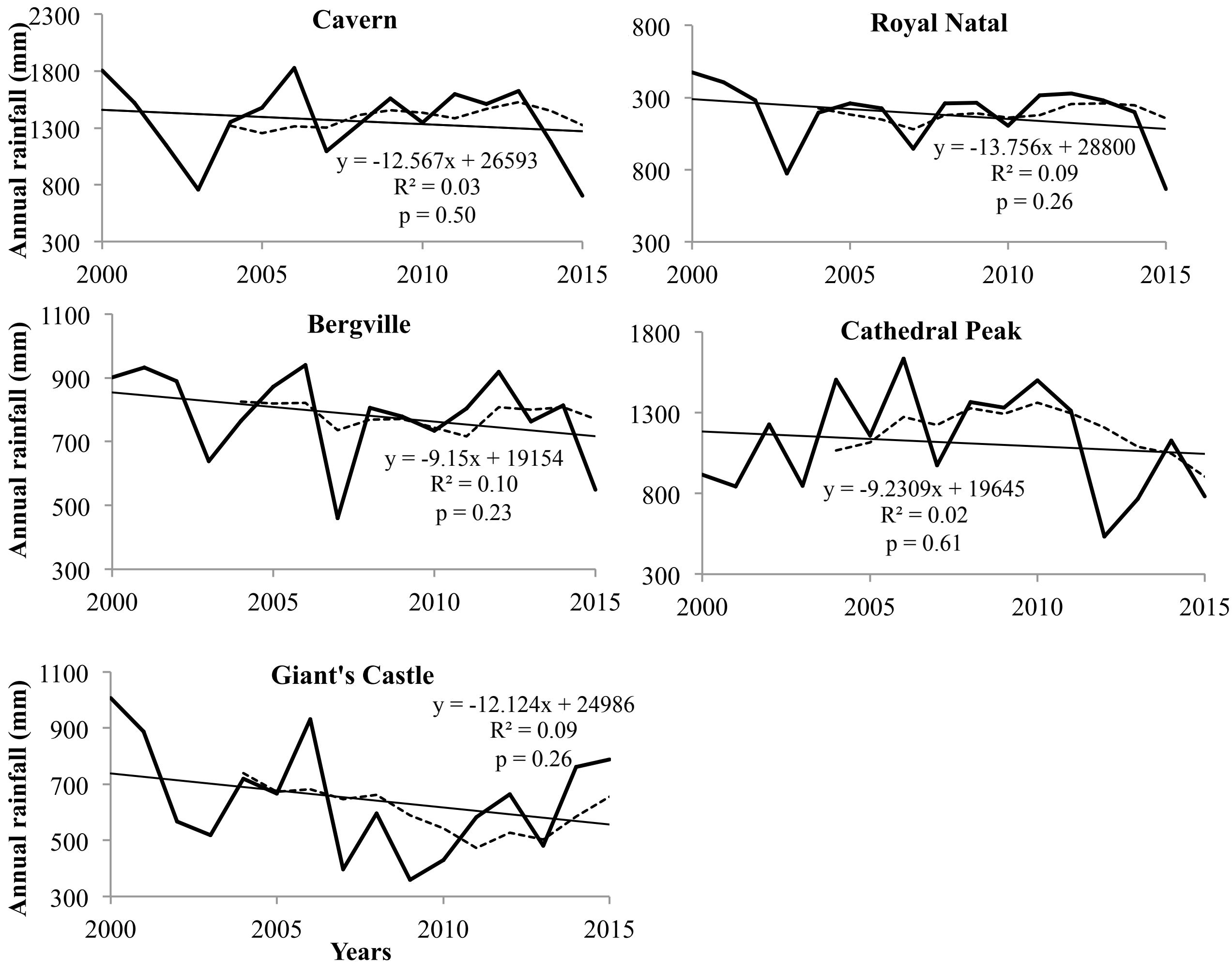


Figure 5.4 Time series of annual rainfall for five stations in the Drakensberg showing the trendline (solid line) and 5-year moving average (dashed line) for the period 2000-2015.

A linear trend of the mean annual rainfall for the combined stations in the Drakensberg, from 1955-2015, shows a statistically significant decrease in the mean annual rainfall over time (P=0.04) (Figure 5.5). The most predominant dry years observed including the following: 1982 (736.86 mm), 1992 (658.50 mm), 1994, 2003 (707.46 mm), 2007 (775.02 mm) and 2015 (698.16 mm). The years of high rainfall include 1957 (1397.66 mm), 1974 (1446.96 mm), 1976 (1525.94 mm), 1988 (1503.12 mm), 1996 (1401.26 mm) and 2006 (1311.78 mm), and these years are always followed by a decrease in rainfall in the subsquent years. It must be noted that small-scale fluctuations in rainfall show that a cyclicity of rainfall peaks, occurring approximately every three years.

The data show that it can be observed that there are four major cycles of rainfall peaks occurring over time, with an average cyclicity of 12 years, namely, 1957-1976, 1976-1988, 1988-1996 and 1996-2006. Five major cycles in rainfall troughs were identified, including 1965-1970, 1970-1982, 1982-1992, 1992-2003 and 2003-2015. There is an average cyclicity of 10 years for the low rainfall.

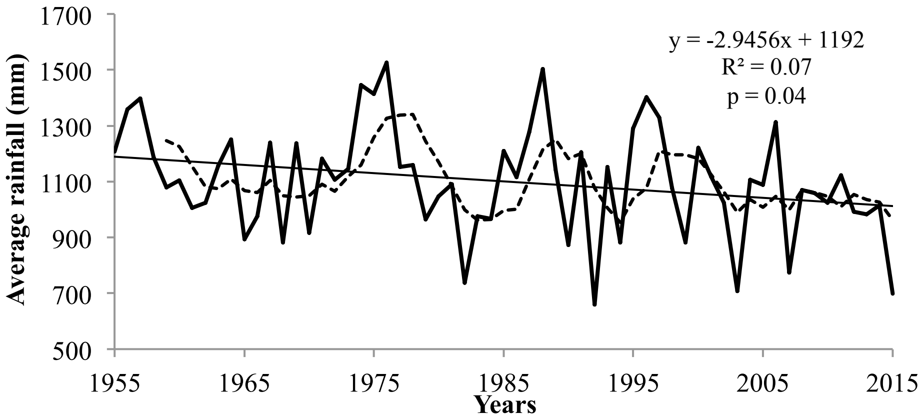


Figure 5.5 Time series of mean annual rainfall for five combined stations, showing the trendline (solid line) and 5-year moving average (dashed line) for the period 1955-2015.

The analysis of the last 16 years of annual rainfall data still depicts a decreasing rainfall trend; however, this is not significant at the 95% confidence level (Figure 5.6). The data is slightly less variable, with a strengthened r squared value. The last 16 years did not have a substantial effect on the overall rainfall trend for the 61-year period of analysis.

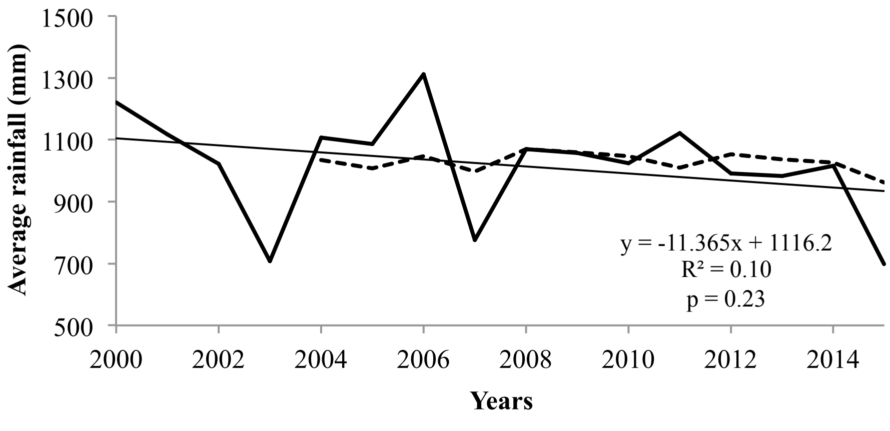
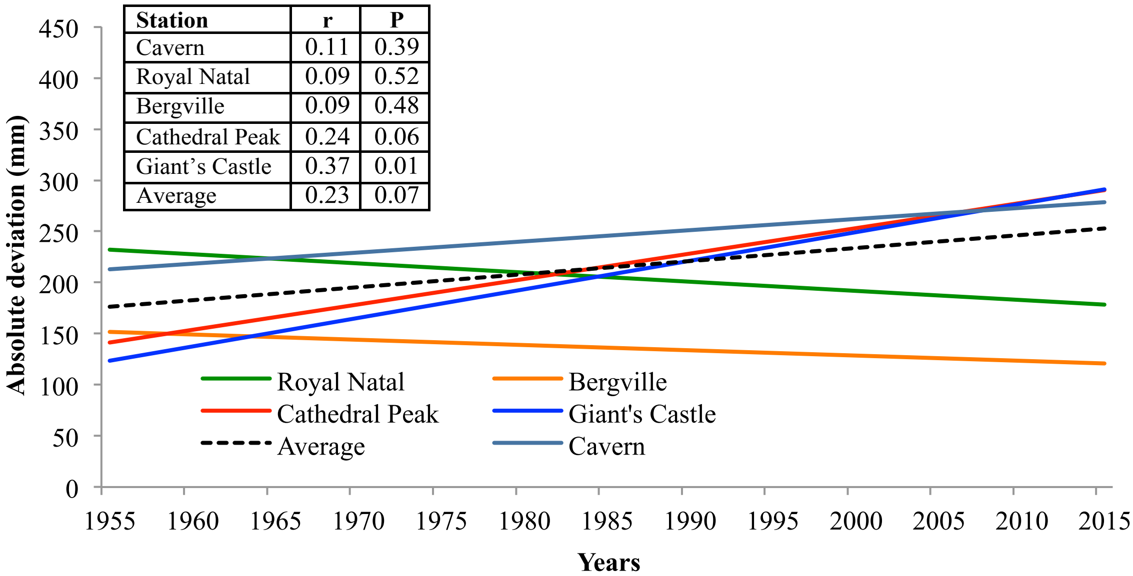


Figure 5.6 Time series of mean annual rainfall for five combined stations showing the trendline (solid line) and 5-year moving average (dashed line) for the period 2000-2015.

All stations were analysed for trends in absolute deviation of rainfall from the mean for the period 1955-2015 (Figure 5.7). All stations, except for Royal Natal and Bergville, show an increase in the variability of annual rainfall over time. Only Giant's Castle has a statistically significant trend (P=0.01). The rainfall seems to have become more variable in the most recent 30 years, with the highest deviation reaching 900 mm in 1993.



Station	r	P
Cavern	0.11	0.39
Royal Natal	0.09	0.52
Bergville	0.09	0.48
Cathedral Peak	0.24	0.06
Giant's Castle	0.37	0.01
Average	0.23	0.07

Figure 5.7 Linear trends of the absolute deviation of rainfall from the mean measured at five stations for the period 1955-2015.

The linear trend of the average rainfall for the five combined stations from 2000-2015 depicts a trend in increasing rainfall deviation from the mean, indicating an overall increase in interannual variability (Figure 5.8). All stations, except Cavern and Bergville, have an increasing trend but none of these trends are significant to the 95% confidence level.

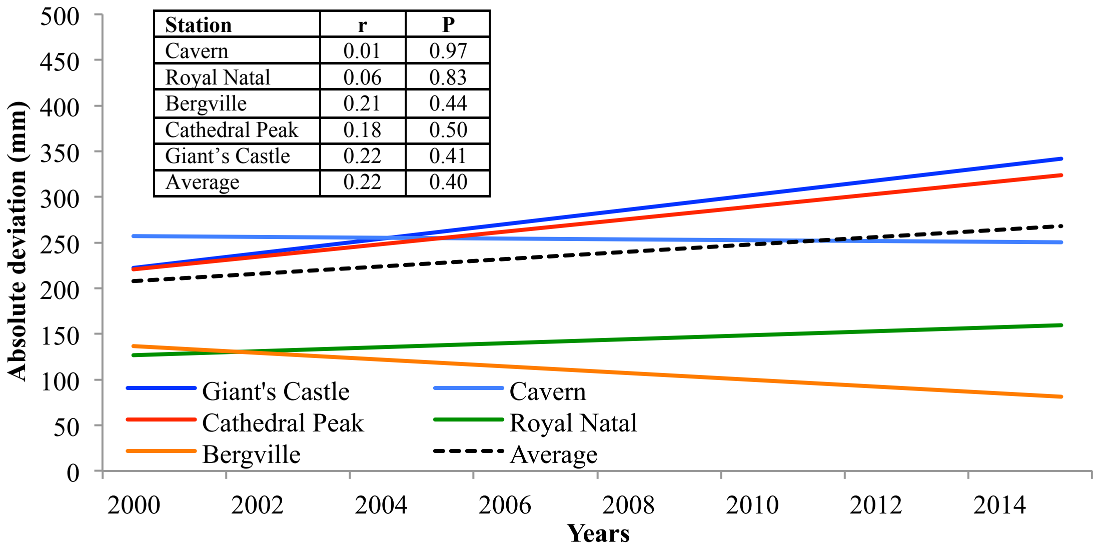


Figure 5.8 Linear trends of the absolute deviation of rainfall from the mean measured at five stations for the period 2000-2015.

5.2.3 Trends in intra-annual rainfall variability

An increase in the Precipitation Concentration Index (PCI) indicates an increase in the intra-annual variability of monthly rainfall. All stations show an increase in PCI over time, which is statistically significant at Giant's Castle ($P<0.001$) and Cathedral Peak ($P=0.01$), indicating that the seasonal nature of rainfall measured at all stations is strengthening (Figure 5.9). From the linear regression, the PCI increased by 4.5 values, from 14.4 to 18.9, at Cathedral Peak and by 3.5 values, from 13.5 to 18.9, at Giant's Castle.

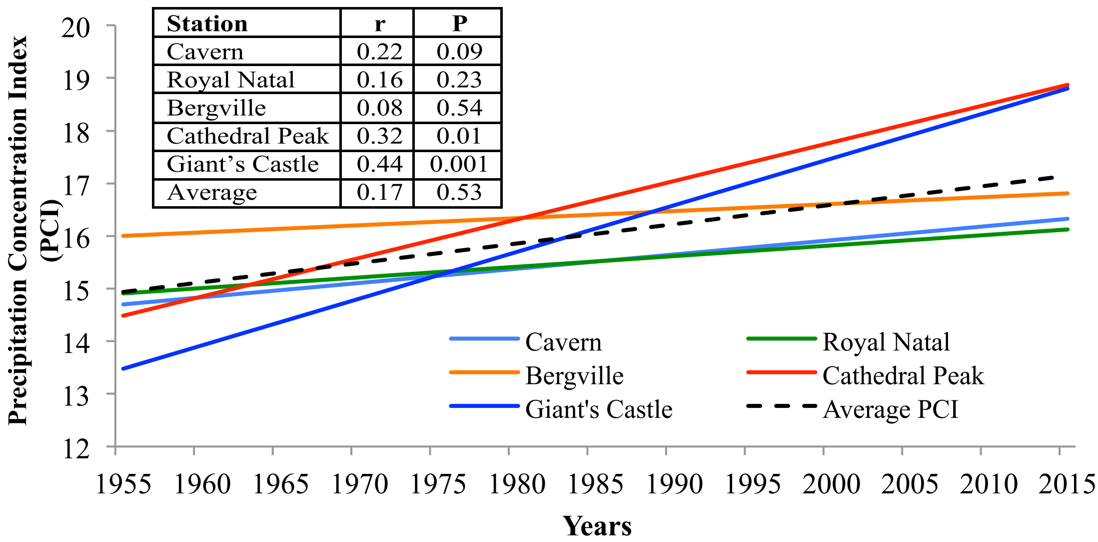


Figure 5.9 Linear trends of Precipitation Concentration Index measured at five stations for the period 1955-2015.

An analysis of the post-2000 data reveals that the increasing PCI trend remains for all stations (Figure 5.10). For this period, Royal Natal (P=0.003), Cathedral Peak (P=0.01), Cavern (P=0.03) and the combined station average PCI (P<0.001) are all statistically significant to the 95% confidence level.

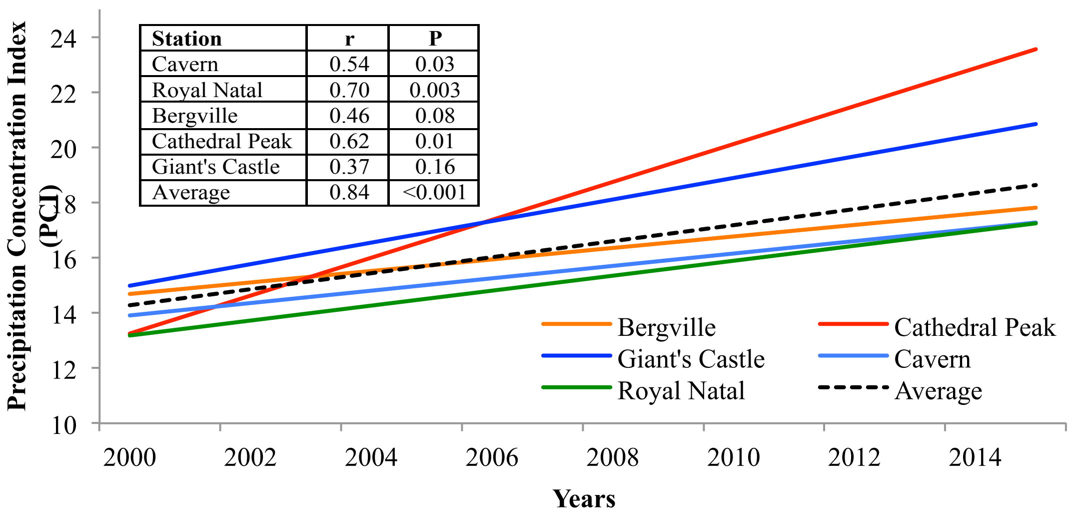


Figure 5.10 Time series of the Precipitation Concentration Index measured at each station in the Drakensberg for the period 2000-2015.

5.2.4 Seasonality

Analysis of the seasonality of rainfall at the individual stations is shown in Figure 5.11. All seasons show a decreasing trend in rainfall over time. In summer all stations, except Bergville, have decreasing rainfall, with Giant's Castle being the only station with a significant decrease in summer rainfall (P<0.001). All stations show decreasing rainfall trends over time in winter and Giant's Castle has a significant trend (P=0.05). In autumn, all stations, except Bergville, have decreasing rainfall and Cathedral Peak (P=0.05), Giant's Castle (P < 0.001) and the combined trend (P=0.05) have statistically significant trends. Spring rainfall depicts a decrease in rainfall over time across all stations, but no trends are significant at the 95% confidence level.

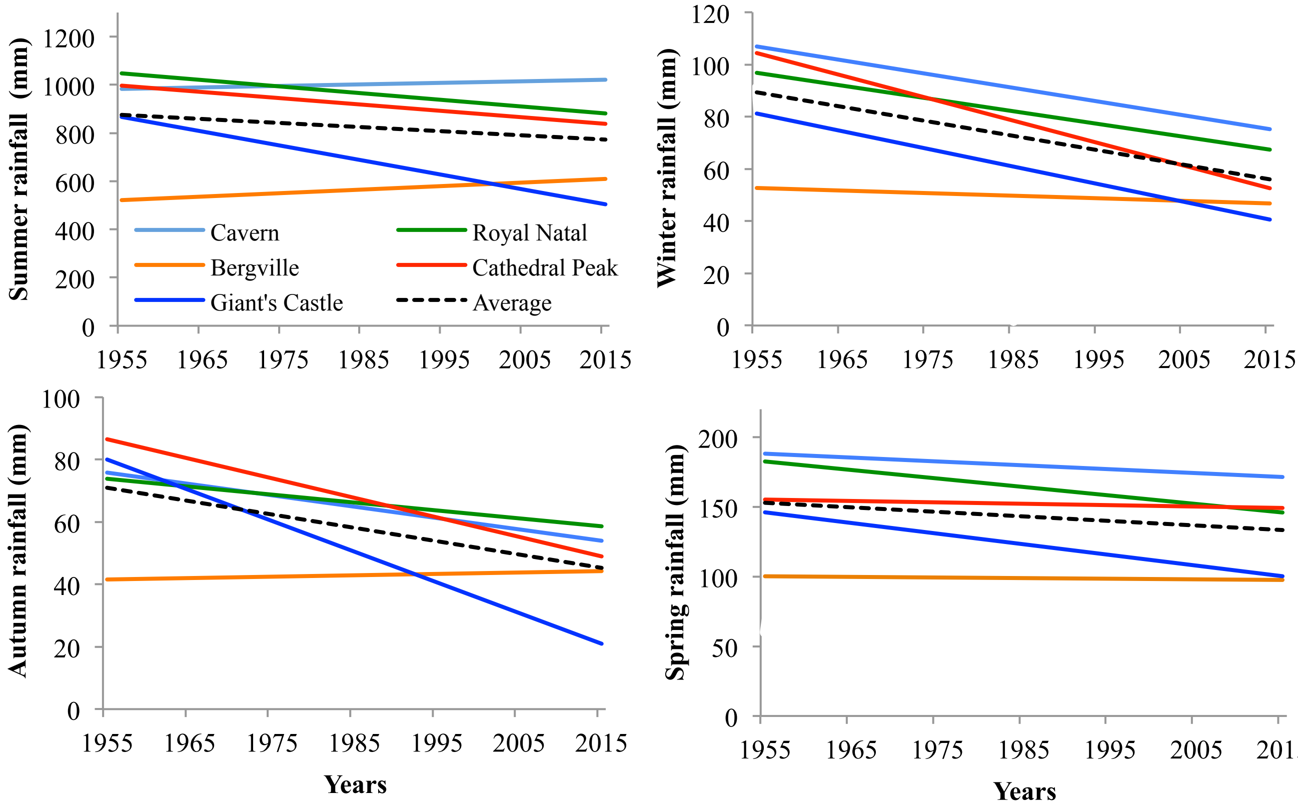


Figure 5.11 Time series of linear trends of seasonal rainfall at the five stations for the period 1955-2015.

In summer (except Bergville), winter and spring, all individual stations have decreasing rainfall for the period 2000-2015 (Figure 5.12). Giant's Castle has a significant decreasing winter rainfall trend, where P=0.01. In autumn, Cavern and Cathedral Peak display increasing rainfall trends, whilst the other stations have decreasing trends. The Bergville station has decreasing trends in autumn and spring, in contrast to the 1955-2015 data, where these seasons had increasing rainfall trends over time. The seasonal trends of the combined stations are all decreasing over time, but are not statistically significant to the 95% confidence level.

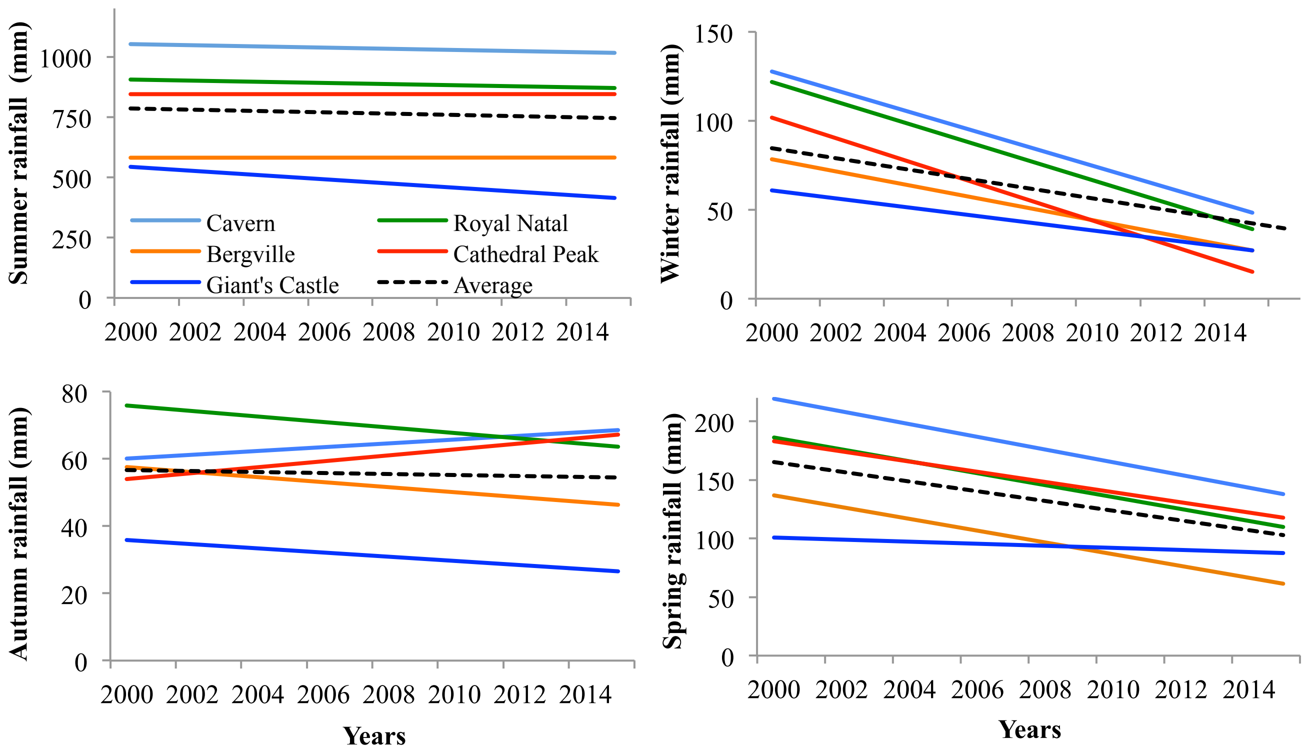


Figure 5.12 Time series of linear trends of seasonal rainfall for the five stations for the period 2000-2015.

An overall decrease in average summer, winter, spring and autumn rainfall for all five stations from 1955-2015 is observed (Figure 5.13). Autumn is the only station with a statistically significant (P=0.05) decrease in rainfall over time. From 2000-2015, all seasons continue to have decreasing rainfall over time, however, none of these trends are significant to the 95% confidence level (Figure 5.14).

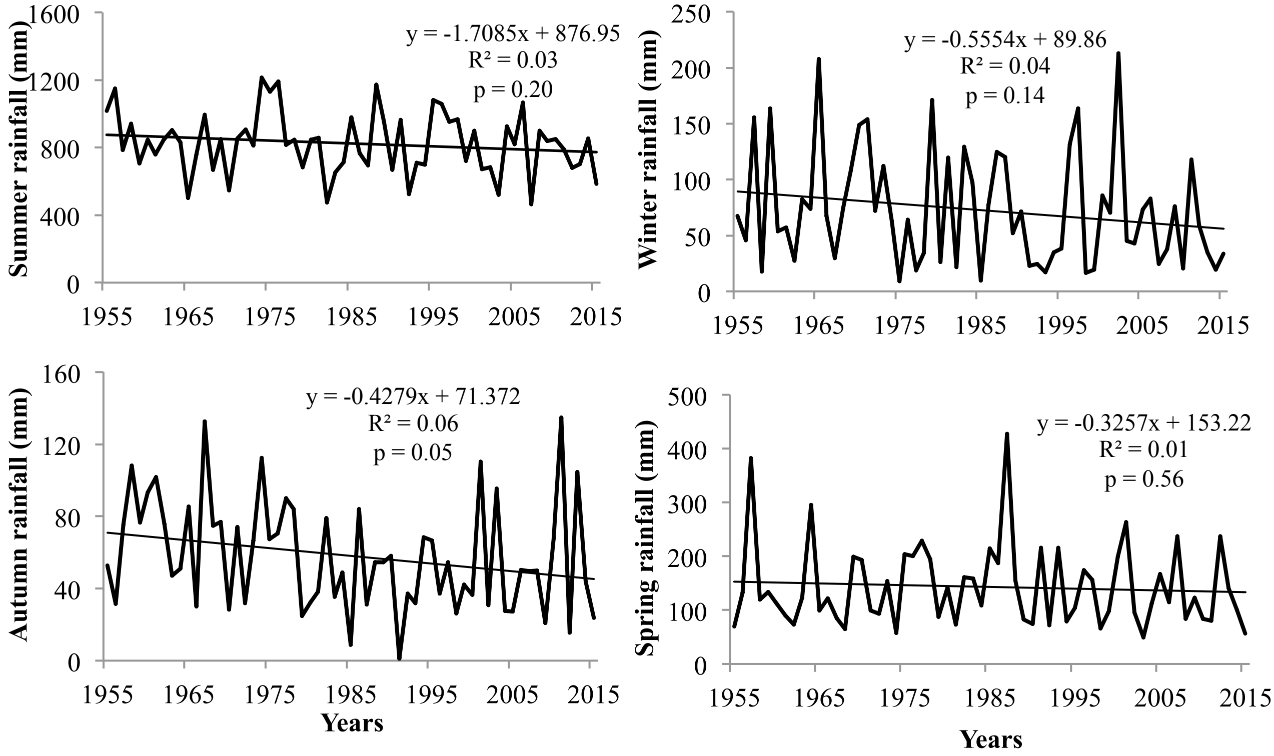


Figure 5.13 Time series and linear trends of summer, winter, spring and autumn rainfall for five combined stations for the period 1955-2015.

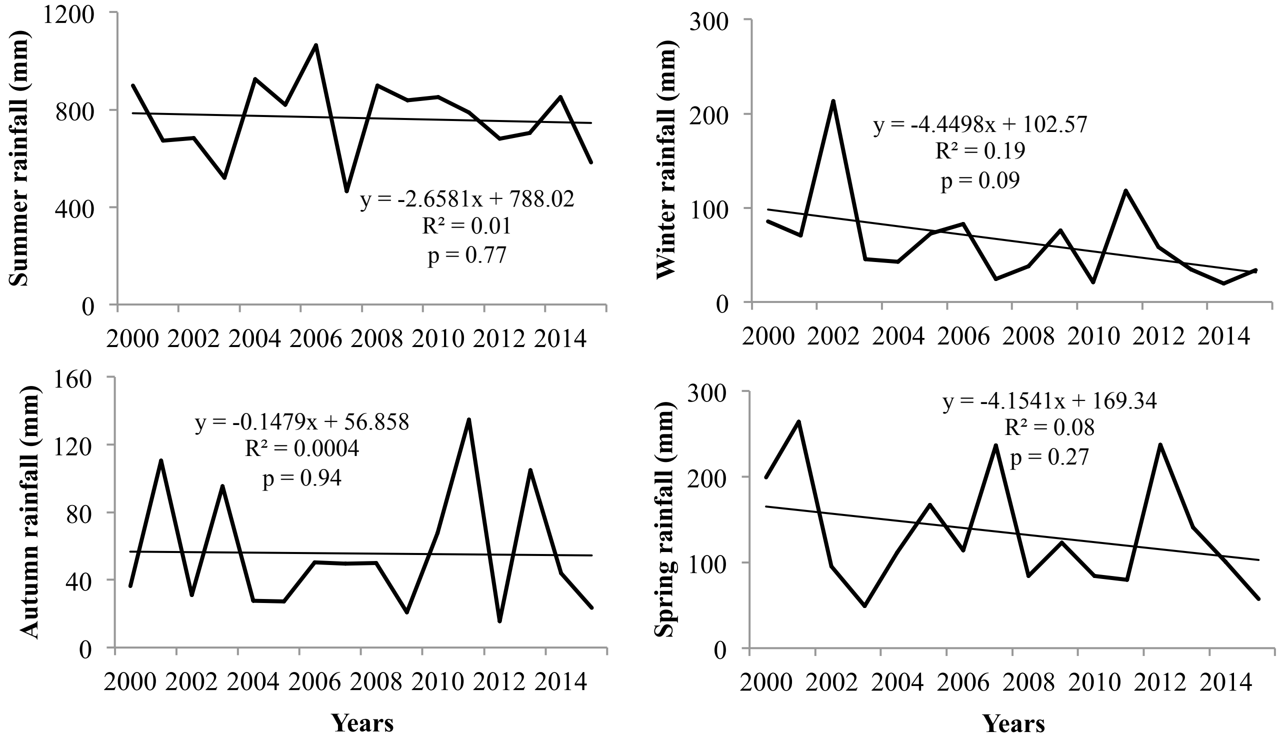


Figure 5.14 Time series and linear trends of summer, winter, spring and autumn rainfall for five combined stations for the period 2000-2015.

5.2.5 Southern Oscillation and summer rainfall

A statistically significant correlation ($r = 0.47$, $P < 0.001$) exists between summer rainfall in the Drakensberg region and the SOI (Table 5.2). The correlation of the summer rainfall is highly significantly. All lagged correlations that were tested are highly significant for $P<0.001$. The correlation coefficients between summer rainfall and the preceding months are all above 0.4, with the highest lagged correlation ($r = 0.45$) from the four months preceding the start of the summer rainfall season (July to October)**.**

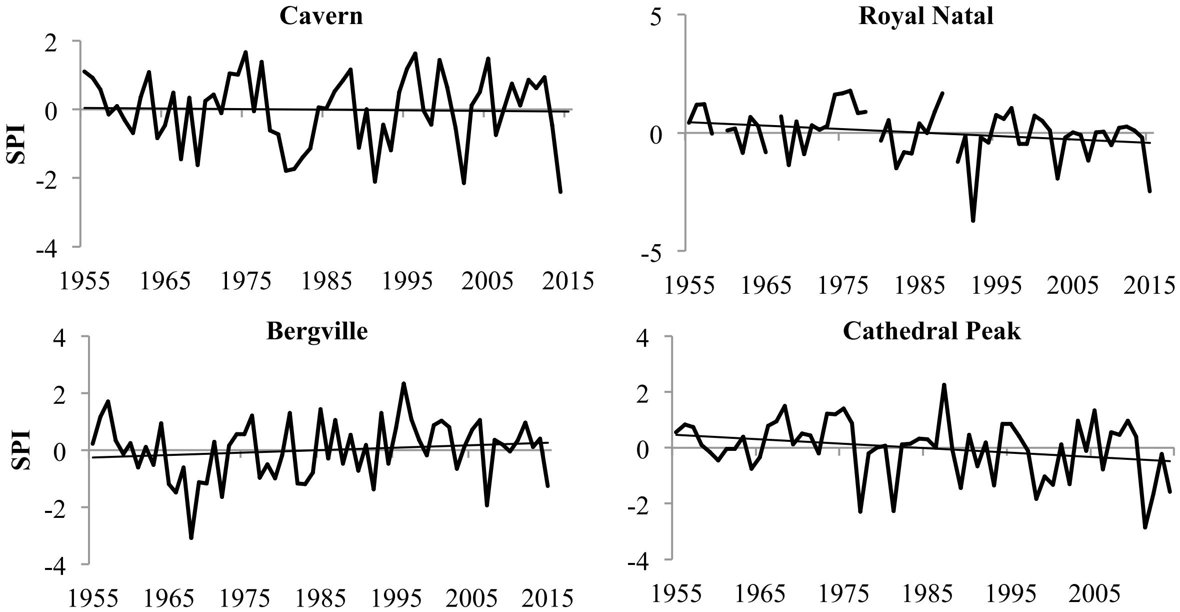
Table 5.2 Correlation coefficient *r* with the relevant level of significance *P* between standardized regional summer rainfall and the mean SOI values for certain periods.

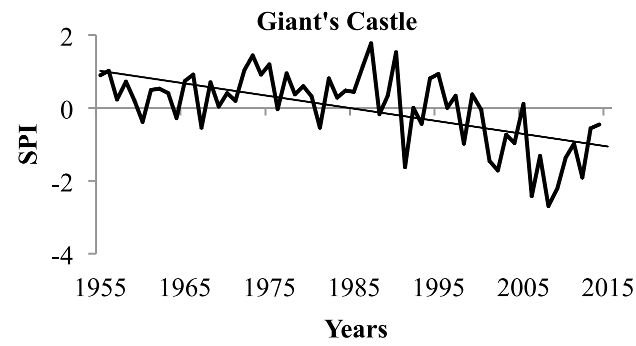
Rainfall Period	**Period of SOI values (non-lagged)**	**r**	**p**
November - March	Nov + Dec + Jan	0.39	0.002
	Nov + Dec + Jan+ Feb + Mar	0.47	<0.001
	Period of SOI values (lagged)		
	May + Jun + Jul + Aug + Sep	0.42	<0.001
	Jun + Jul + Aug + Sep	0.45	<0.001
	Jun + Jul + Aug + Sep + Oct	0.44	<0.001
	Jul + Aug + Sep	0.45	<0.001
	Jul + Aug + Sep + Oct	0.45	<0.001

5.3 Drought Analysis: Central and Northern Drakensberg

5.3.1 Standard Precipitation Index

The SPI classification can be found in Table 5.3. Four of the five stations in the Drakensberg show a trend in decreasing SPI values from 1955-2015 and Bergville has an increasing trend (Figure 5.15). Royal Natal, Cathedral Peak and Giant's Castle are statistically significant at the 95th confidence level, with Giant's Castle having a highly significant trend ($P<0.001$). All stations showed drying conditions in the early 1980s and 1990s, for example, Cavern experienced severe to extremely dry conditions, Bergville experienced moderately dry conditions, Royal Natal had extreme dryness in 1994 (-2.97) and severe dryness in 1982, Giant's Castle experienced 'near normal' conditions in 1982 and severe conditions in 1991, and in 1981 Cathedral Peak was extremely dry (-2.27), with moderately dry years in 1990 and 1993.





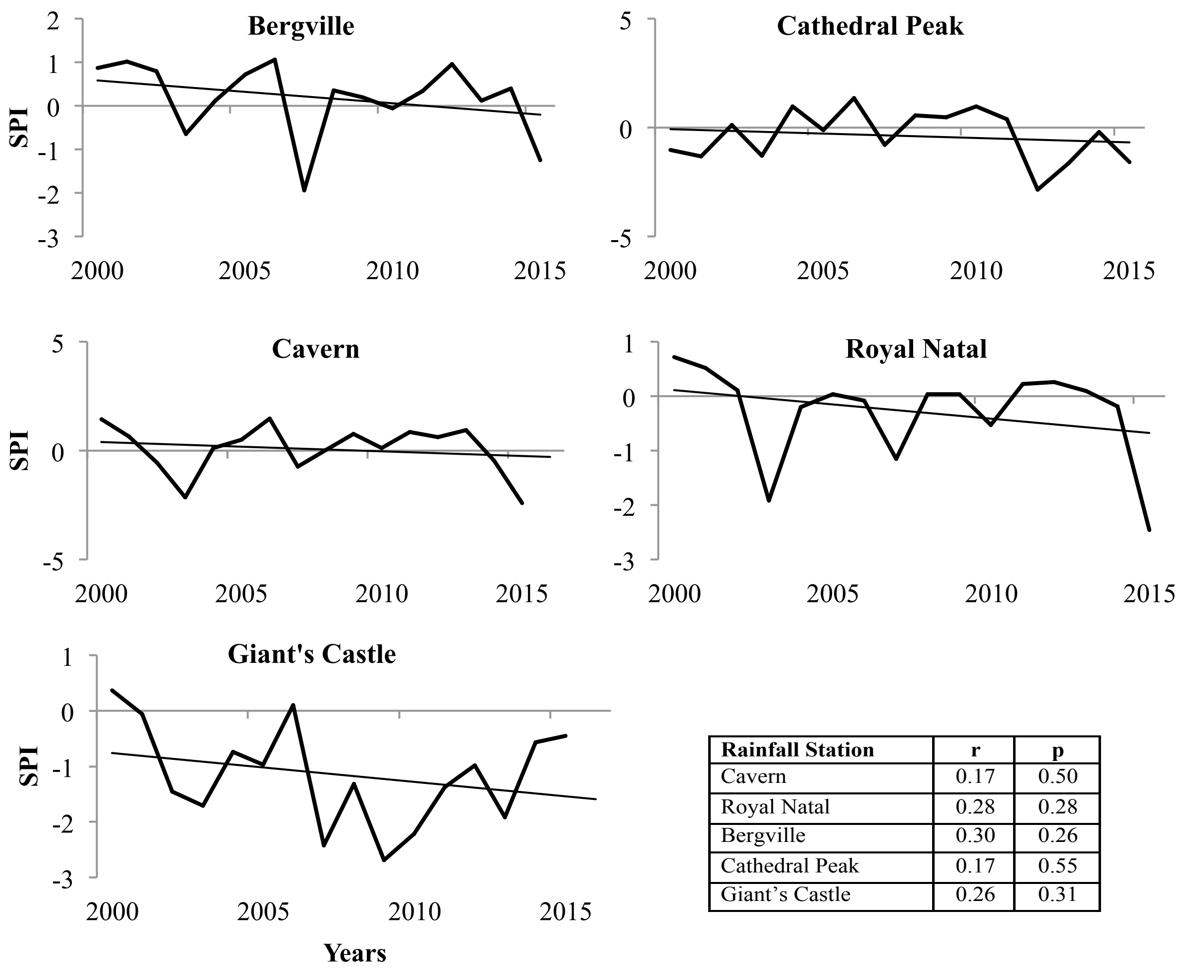
Rainfall Station	r	p
Cavern	0.04	0.72
Royal Natal	0.26	**0.05**
Bergville	0.39	0.24
Cathedral Peak	0.28	**0.03**
Giant’s Castle	0.60	**<0.001**

Figure 5.15 Time series analysis of Standard Precipitation Index for five stations for the period 1955-2015.

Another dry period experienced across the stations occurred around 2002-2003 and 2006-2007 (Figure 5.16). All stations, except Cathedral Peak, experienced drier conditions in these years. In 2002, Cavern was classified as extremely dry and in 2003, Royal Natal and Cathedral Peak were classified as severely dry. In 2007, Royal Natal was moderately dry and Bergville was severely dry, whereas Cathedral Peak shows near normal conditions. Giant’s Castle experienced extremely dry conditions in 2006 and 2008 and moderately dry conditions in 2007. In 2015, dry conditions were experienced at all stations. Cavern and Royal Natal faced extremely dry conditions, with SPI values of -2.41 and -2.46

respectively, Bergville showed moderately dry conditions (-1.25), Cathedral Peak (-1.58) experienced severely dry conditions and Giant’s Castle had near normal conditions (-0.45).

The SPI results computed for 2000-2015, for the stations in the Drakensberg, show a decreasing SPI over time for all stations (Figure 5.16). However, none of these trends are significant to the 95% confidence level. The low SPI values around 2003, 2007 and 2015 are easily discerned.



Rainfall Station	r	p
Cavern	0.17	0.50
Royal Natal	0.28	0.28
Bergville	0.30	0.26
Cathedral Peak	0.17	0.55
Giant’s Castle	0.26	0.31

Figure 5.16 Time series analysis of the SPI for five stations for the period 2000-2015.

Across the stations, the extremely dry years are noted as 1982, 1992, 2003 and 2015 (Table 5.3). Near normal conditions occur 75.4% of the time; however, a higher occurrence of severely dry years, rather than very wet years, is noted. Over the 61-year period, more dry periods than wet periods occurred in approximately the last 30 years, indicating an increase in the number of dry years over time (Figutr 5.17).

Table 5.3 Percent occurrence of each drought category for the average SPI values for five rainfall stations in the Drakensberg.

SPI Value	Drought category	% Occurrence 1955 – 2015	% Occurrence 2000 – 2015	Years
> 2.00	Extremely wet	0.0	0.0	None
1.5 to 1.99	Very wet	6.6	0.0	1974. 1975, 1976, 1988
1 to 1.49	Moderately wet	8.2	6.3	1956, 1957, 1996, 1997, 2006
- 0.99 to 0.99	Near normal	68.9	75.0	42 in total
- 1.00 to -1.49	Moderately dry	8.2	0.0	1965, 1968, 1990, 1994, 1999
-1.50 to – 1.99	Severely dry	1.6	6.3	2007
< - 2.00	Extremely dry	6.6	12.5	1982, 1992, 2003, 2015

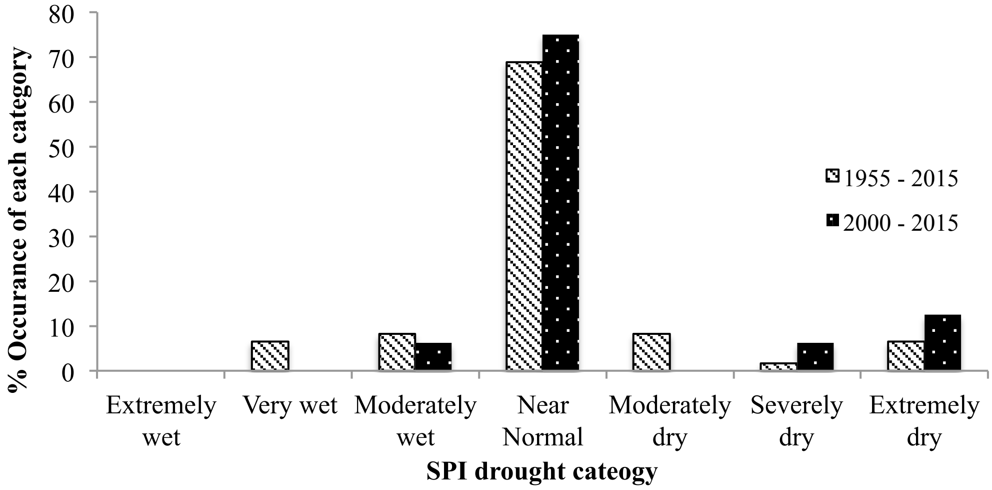


Figure 5.17 Percent occurrence of SPI drought categories for the periods 1955-2015 and 2000-2015 for the average SPI values for five stations in the Drakensberg.

5.4 Sterkspruit Catchment

As previously mentioned, the Sterkspruit catchment stations are analysed individually over the time period, which is consistent with the beginning of each station's rainfall data measured up until 2015. A grouped station analysis is computed from 1974-2015, where data for all the stations is consistently available. Analysis is computed on the hydrological year, November–March (summer), May–August (winter), April (autumn) and September and October (spring).

5.4.1 General rainfall attributes

A summer-dominated rainfall trend occurs in the catchment, with 75.78% of the rainfall falling from November to March (summer) (Figure 5.18). Very low rainfall totals are experienced in winter, with only 5.84% of the rainfall falling in this period. Spring accounts for 13.13% of the rainfall and autumn 5.23%. Table 5.4 is a summary of the rainfall characteristics of the stations. Again, it is evident that this is a summer rainfall region, with every station receiving the most rainfall in summer. In the catchment area, there is a significant positive relationship between altitude and average rainfall at the 95% confidence level (P = 0.003) (Figure 5.19).

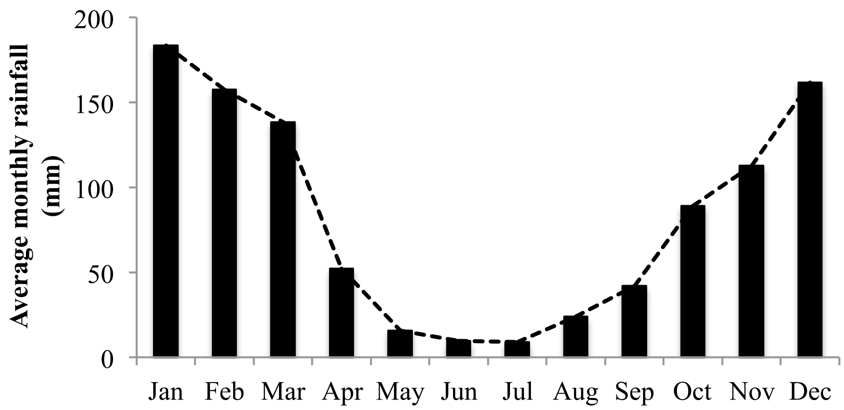


Figure 5.18 Average monthly rainfall for four combined stations in the catchment for the period 1974-2015.

Table 5.4 Summary table of the Sterkspruit Catchment stations.

Rainfall station	**Historical record**	**Altitude (m a.s.l.)**	**MAR (mm)**	**Average summer rainfall (mm)**	**Average winter rainfall (mm)**	**Average autumn rainfall (mm)**	**Average spring rainfall (mm)**
Glenisla	1974	1077	870	114.3	11.5	35.2	52.1
Clydesdale	1966	1115	828	135.7	13.1	46.6	59.9
Jacobson	1968	1443	1457	129.2	11.8	53.4	53.2
Reeves	1957	990	754	224.1	21.7	73	96.8

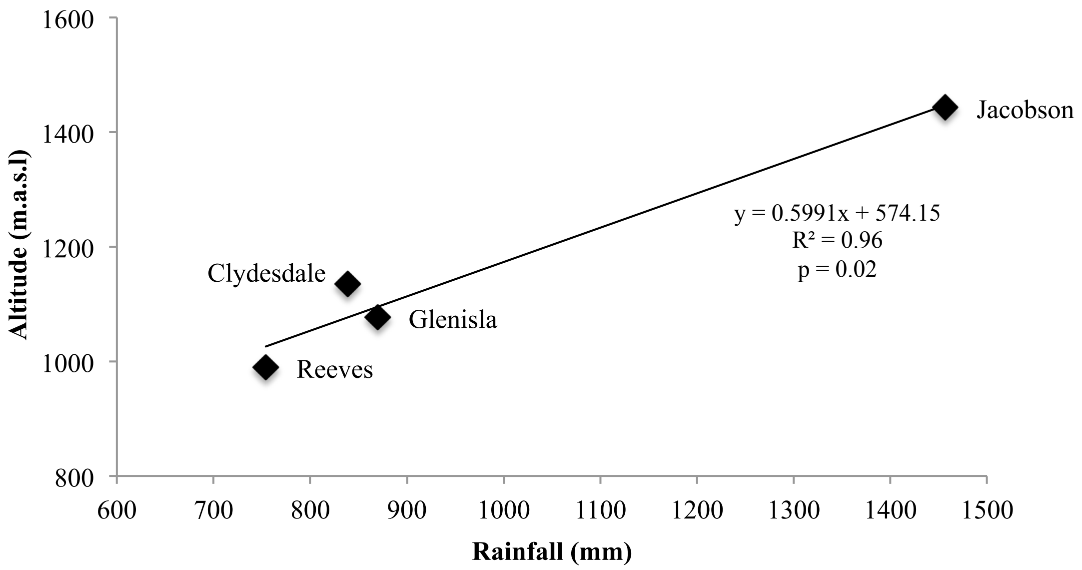


Figure 5.19 Average rainfall for each station in relation to altitude for the period 1974-2015.

5.4.2 Trends in interannual rainfall variability

Annual rainfall time at all four stations depicts an increase in rainfall over time. The Glenisla and Jacobson stations have significant increasing rainfall trends, where P=0.04 and P=0.01 respectively (Figure 5.20). All stations (where data is available) show a below average rainfall around 1968, 1982-1984, 1992-1994, 2002-2003, 2006-2007 and in 2015. Glenisla has an average cyclicity of eight years between rainfall peaks, namely, 1981-1988, 1988-1996, 1996-2005, 2005-2011 after which, the rainfall declines in 2015, which has the lowest recorded rainfall over the time period. Reeves' rainfall has an average cyclicity of nine years between rainfall peaks, as shown by the moving average in 1958-

1967, 1967-1974, 1974-1987, 1987-1995, 1995-2005 and 2005-2012, with a decline in 2015. Clydesdale has an average of nine years between rainfall peaks, namely, 1974-1981, 1981-1988, 1988-1996 and 1996-2011 after which, the trend decreases towards 2015. Jacobson, on average, displays a seven-year cyclicity in rainfall peaks, 1975-1988, 1988-1996, 1996-2006 after which, the trend decreases towards 2015.

The analysis of the post-2000 trends indicates that all four stations have decreasing rainfall trends over time (Figure 5.21). Jacobson was the only station with a significant decreasing trend to the 95th confidence level (P = 0.04). This is in stark contrast to Figure 5.20, where all the rainfall trends are increasing over time, indicating that rainfall patterns have changed in the last 16 years.

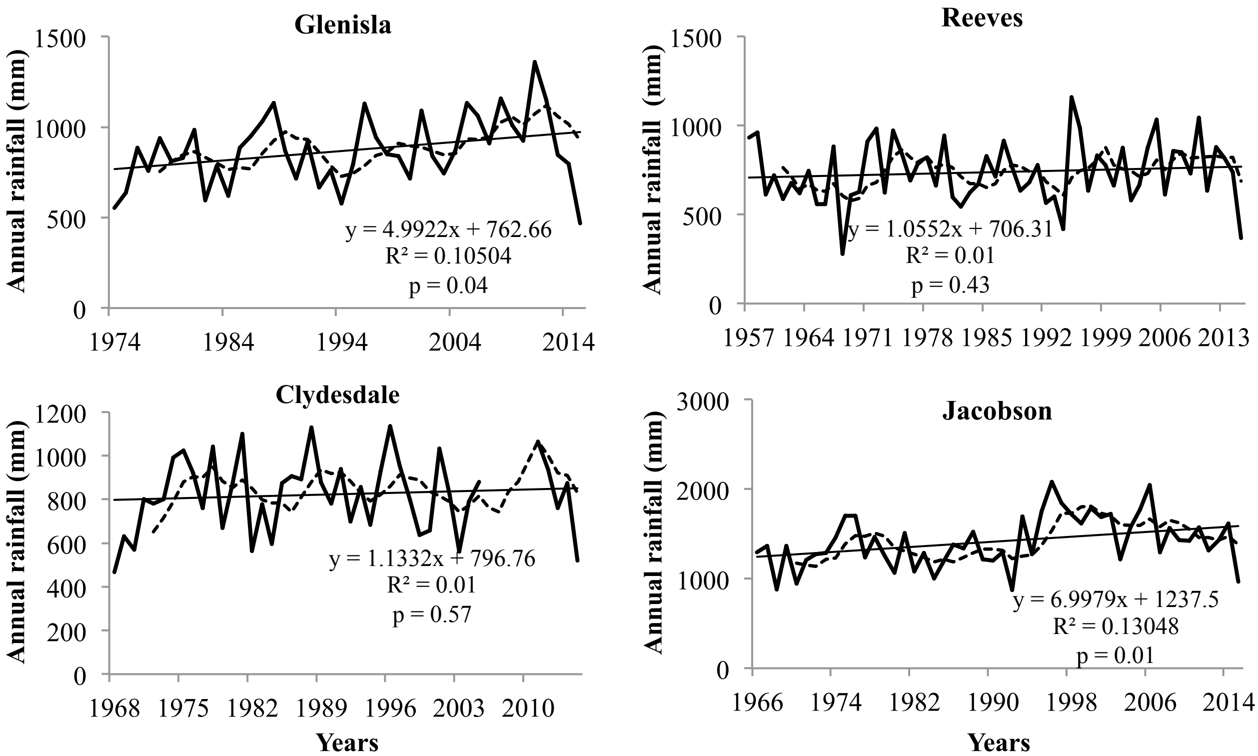


Figure 5.20 Time series of annual rainfall for the four stations in the catchment area with a trendline (solid line) and a 5-year moving average (dashed line).

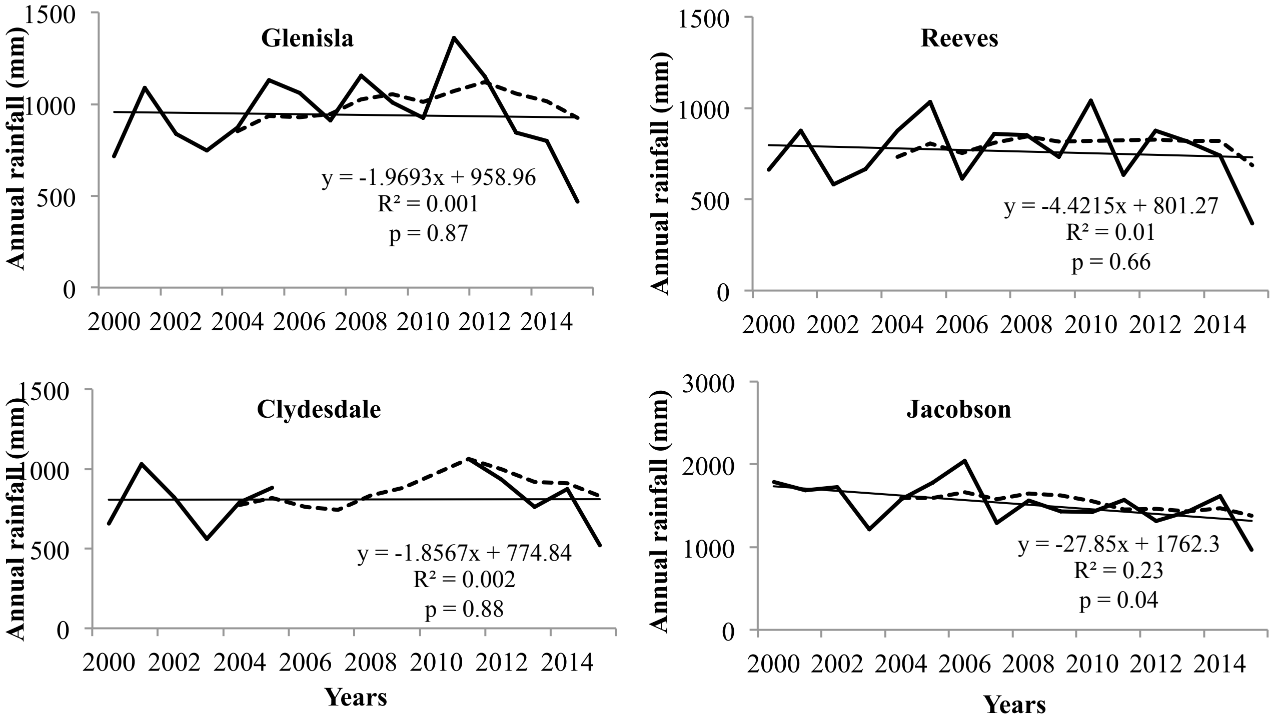


Figure 5.21 Time series of annual rainfall for four stations in the catchment area showing the trendline (solid line) and a 5-year moving average (dashed line) from 2000-2015.

The mean annual rainfall for the combined stations in the catchment area fluctuates over the years (Figure 5.22). The linear trend depicts an increase in rainfall over time; however, this is not statistically significant. Over the 42-year timeline, the lowest rainfall averages occurred in 1982 (696.29 mm), 1992 (700.73 mm) and 2015, which had the lowest recorded rainfall of 579.9 mm. Rainfall peaks occur, on average, every 10 years, namely, 1978-1988, 1988-1996, 1996-2006, and from 2006, the cycle starts to decrease again to 2015. On inspection, it can be seen that small-scale fluctuations occur, on average, every three years. There are three major rainfall troughs that have an average 11-year cyclicity, namely, 1982-1992, 1992-2003 and 2003-2015. An analysis of the post-2000 trends shows a decrease in rainfall over time, but this is not significant to the 95th confidence level (Figure 5.23).

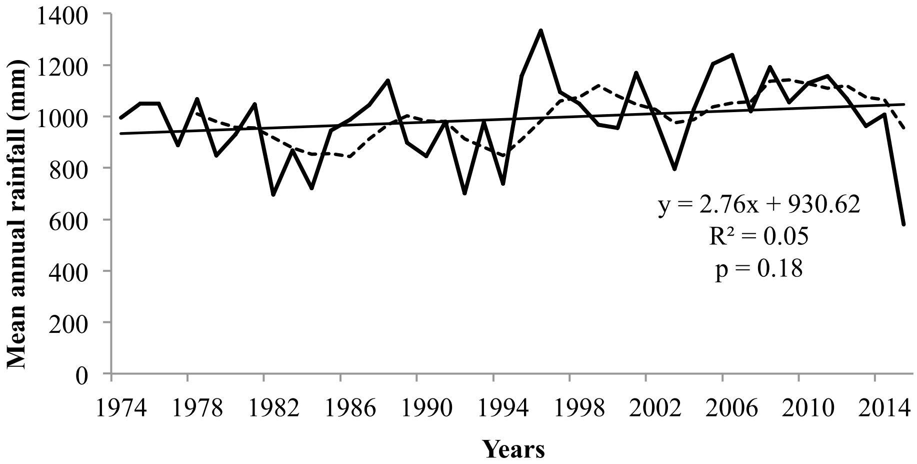


Figure 5.22 Time series of mean annual rainfall for four combined stations showing a trendline (solid line) and the 5-year moving average (dashed line) for the period 1974-2015.

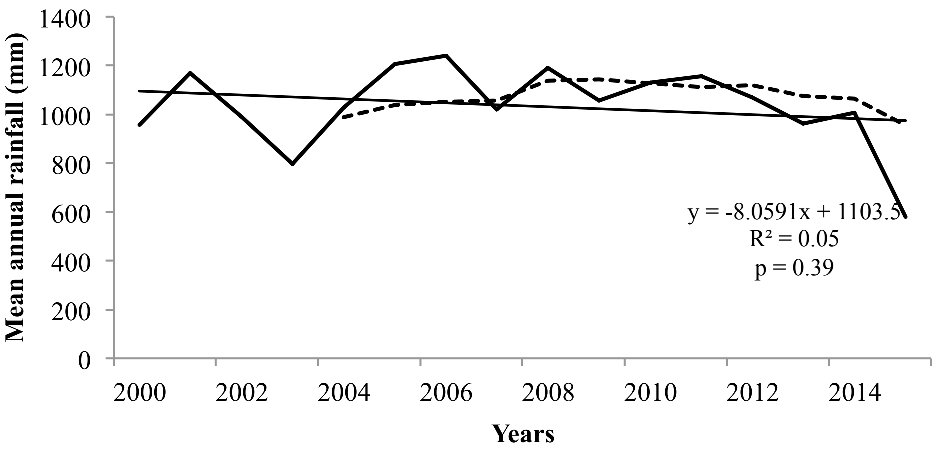
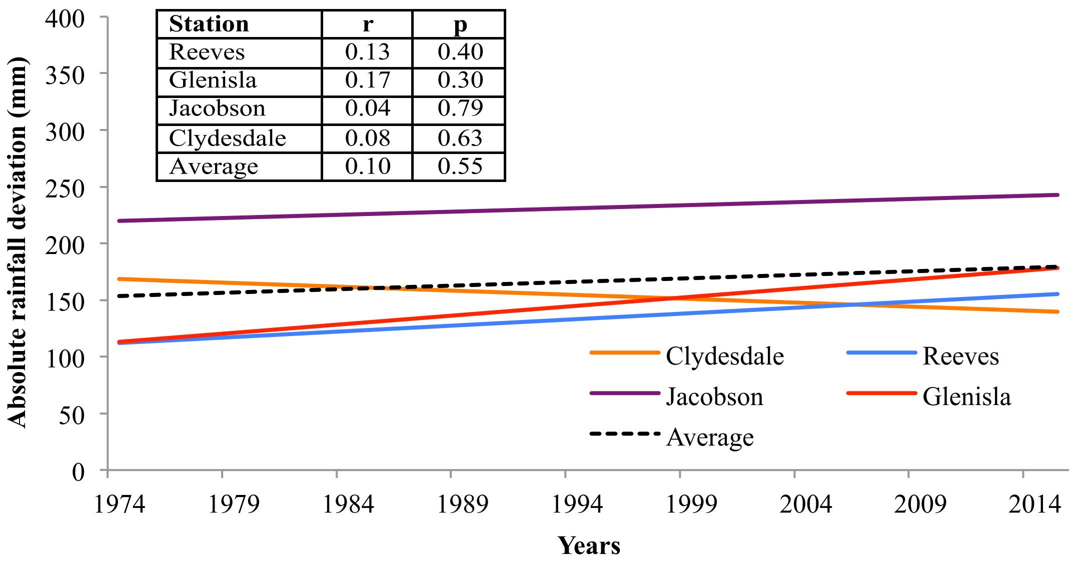


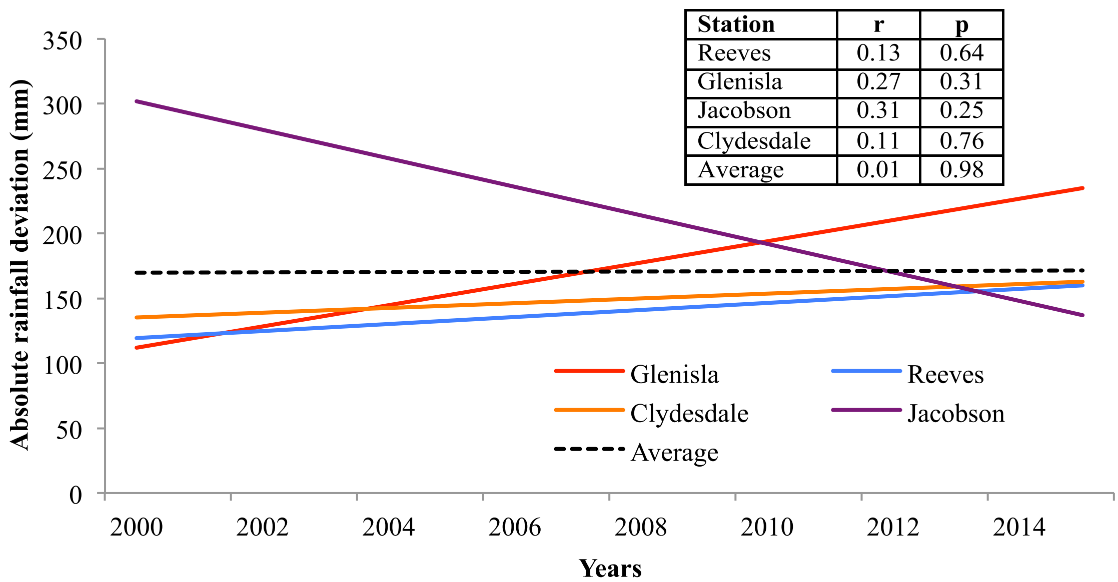
Figure 5.23 Time series of mean annual rainfall for the four combined stations in the catchment showing the trendline (solid line) and 5-year moving average (dashed line) from 2000-2015.

The absolute deviation of rainfall from the mean at all stations, except Clydesdale, increases with time, indicating an increase in variability; however, none of these trends are significant to the 95th confidence level (Figure 5.24). An analysis of the trends from 2000-2015 reveals that the Clydesdale, Reeves and Glenisla stations have increasing liner trends over time, whereas Jacobson displays a decreasing trend (Figure 5.25). For both time periods, the average PCI is increasing, although none of these relationships are significant to the 95th confidence level.



Station	r	p
Reeves	0.13	0.40
Glenisla	0.17	0.30
Jacobson	0.04	0.79
Clydesdale	0.08	0.63
Average	0.10	0.55

Figure 5.24 Linear trends of the absolute deviation of rainfall from the mean annual rainfall measured at each station for the period 1974-2015.



Station	r	p
Reeves	0.13	0.64
Glenisla	0.27	0.31
Jacobson	0.31	0.25
Clydesdale	0.11	0.76
Average	0.01	0.98

Figure 5.25 Linear trends of the absolute deviation of rainfall from the mean annual rainfall measured at each station for the period 2000-2015.

5.4.3 Trends in intra-annual rainfall variability

All stations, except Clydesdale, show an increase in the PCI values over time, indicating an increase in the variability of monthly rainfall, but none of these trends are statistically significant (Figure 5.26). The linear trends running from 2000-2015 reveal that all stations have an increasing PCI over time (Figure 5.27). The average PCI trend (P=0.02), Glenisla (P=0.02) and Jacobson (P=0.03) have significant increasing trends.

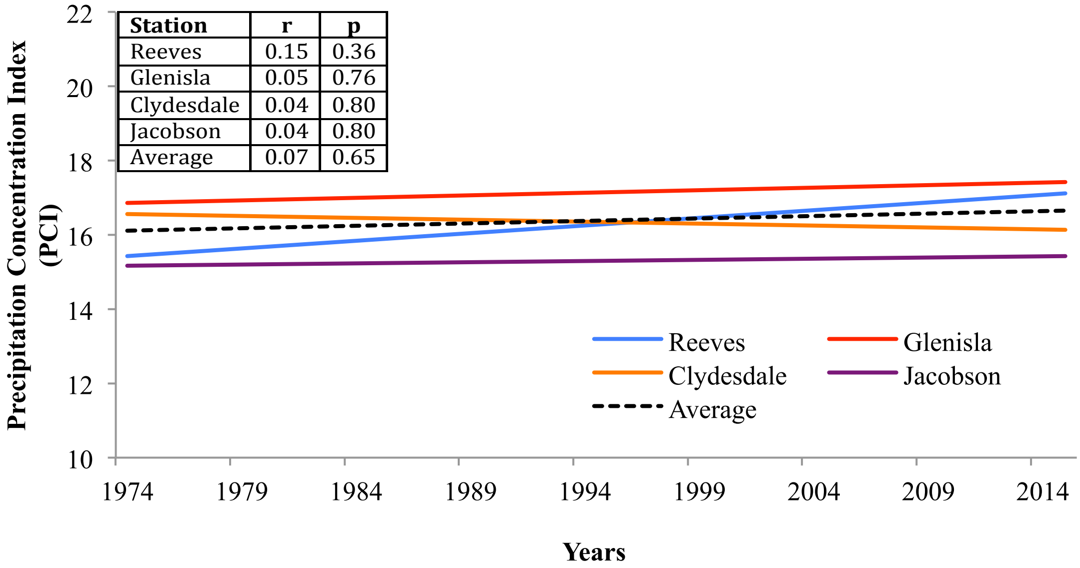


Figure 5.26 Linear trends of PCI measured at each station in the catchment area for the period 1974-2015.

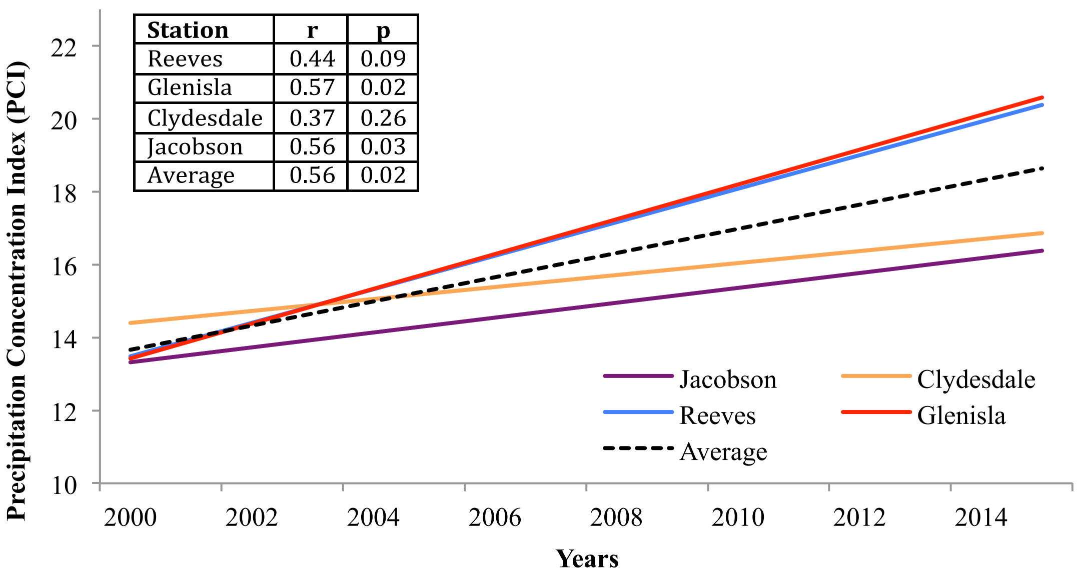


Figure 5.27 Linear trend of PCI measured at each station in the catchment for the period 2000-2015.

5.4.4 Seasonality

The combined station average rainfall linear trend is increasing in all seasons, except in spring (Figure 5.28). Every station, except Reeves, has an increase in summer rainfall over time and only Glenisla has a statistically significant trend (P=0.02) at the 95% confidence level. In autumn, all stations have increasing rainfall over time. Spring rainfall decreases over time for all stations, except Jacobson, which has an increasing rainfall trend. In winter, Jacobson and Reeves have decreasing rainfall trends, whilst Clydesdale and Glenisla have increasing trends over time. None of these trends are significant to the 95% confidence level.

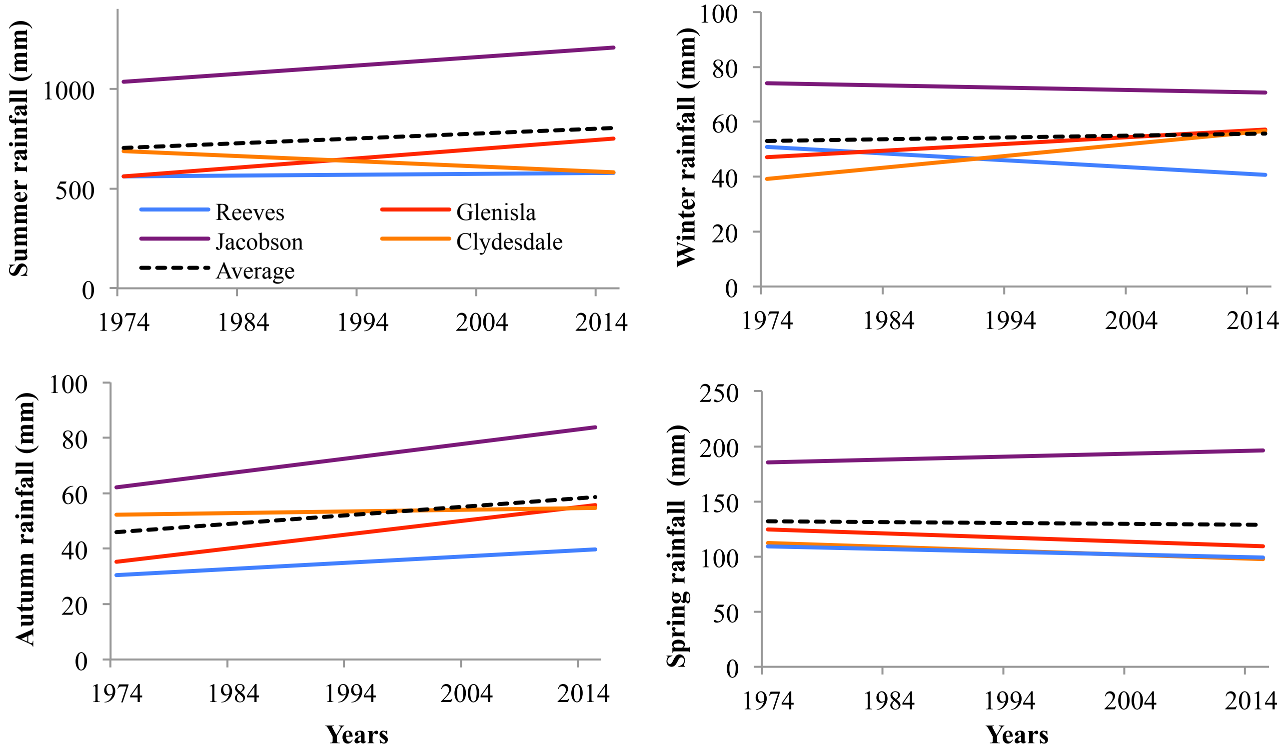


Figure 5.28 Linear trends of annual seasonal rainfall for four stations in the catchment area for the period 1974-2015.

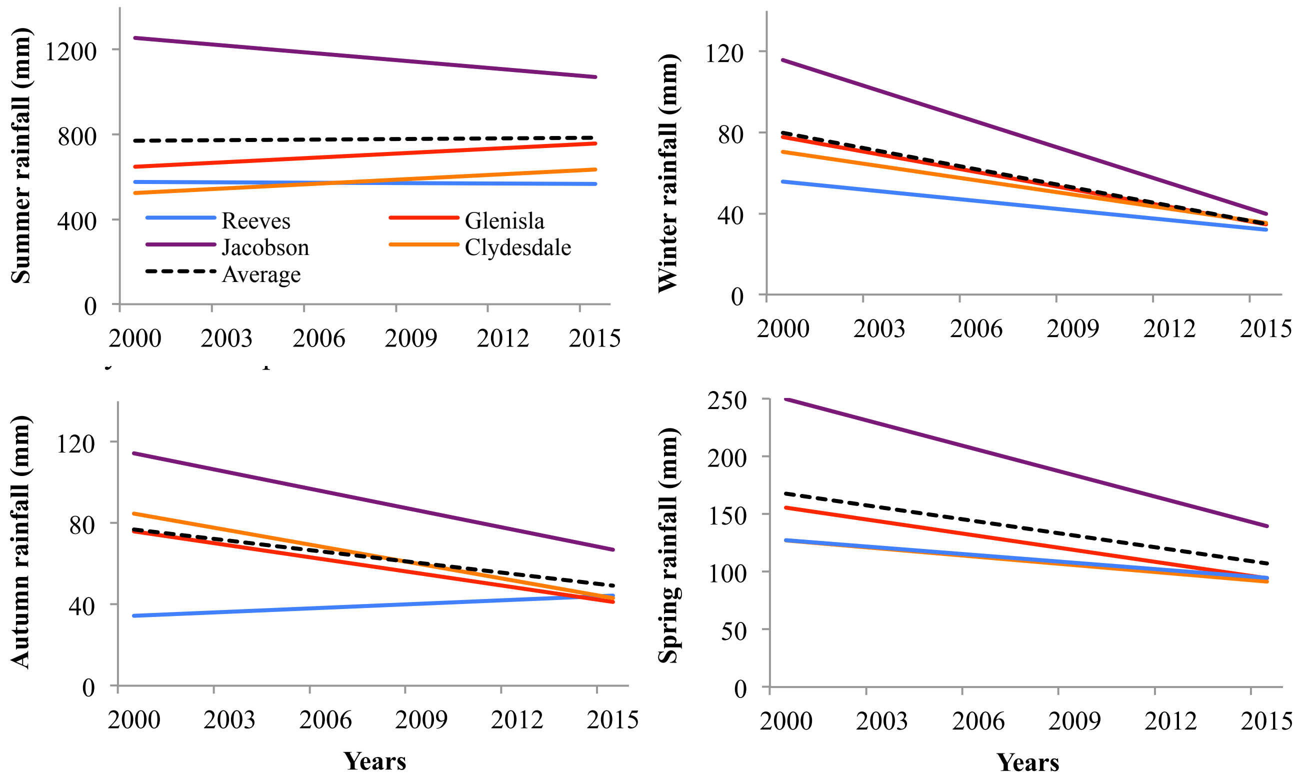


Figure 5.29 Linear trends of annual seasonal rainfall for four stations in the catchment area for the period 2000-2015.

The combined station analysis of the average summer, winter, spring and autumn depicts an increase in the average summer, winter and autumn rainfall and a decrease in average spring rainfall, over the 1974-2015 period (Figure 5.30) . However, none of these trends are significant. The analysis of the post-2000 period reveals that the average winter and autumn rainfall trends are decreasing over time (Figure 5.31). Summer retains an increasing rainfall trend, as does spring decreasing rainfall over time. These trends are not significant to the 95% confidence level.

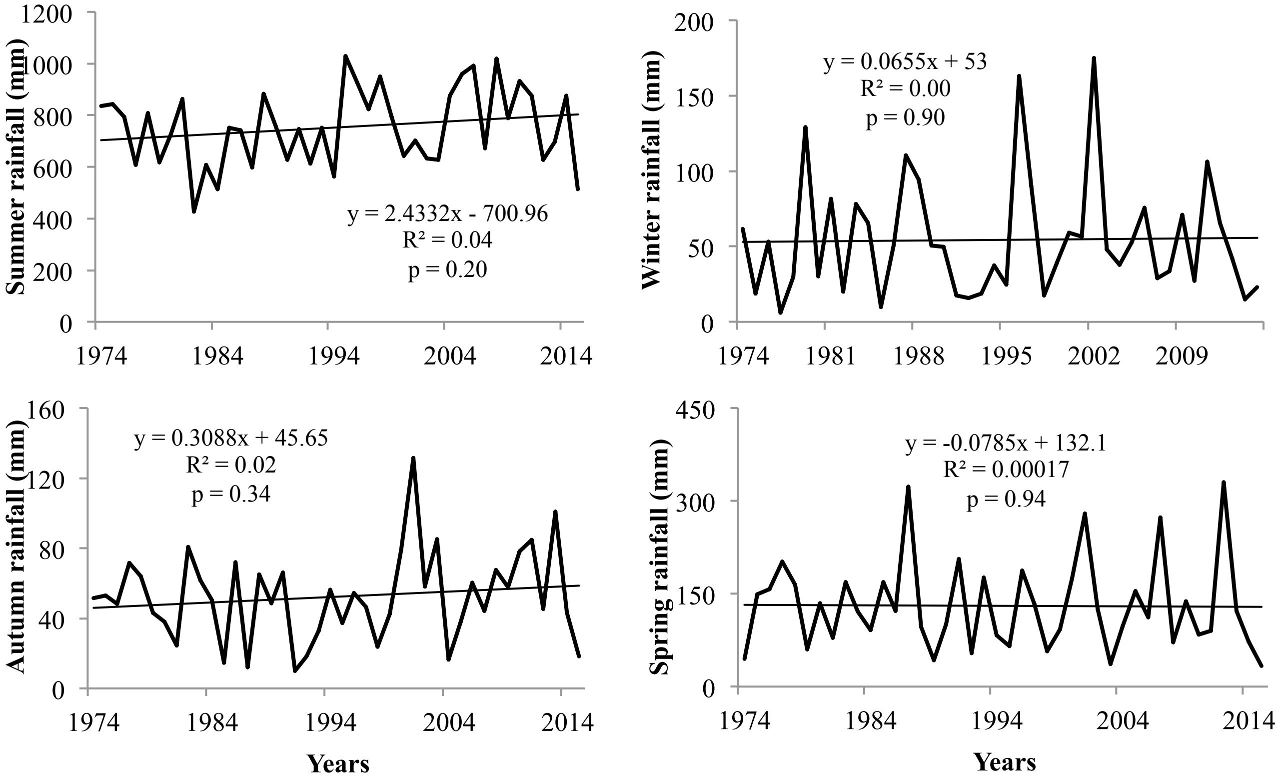


Figure 5.30 Time series and linear trends of summer, winter, autumn and spring rainfall for four combined stations in the catchment for the period 1974-2015.

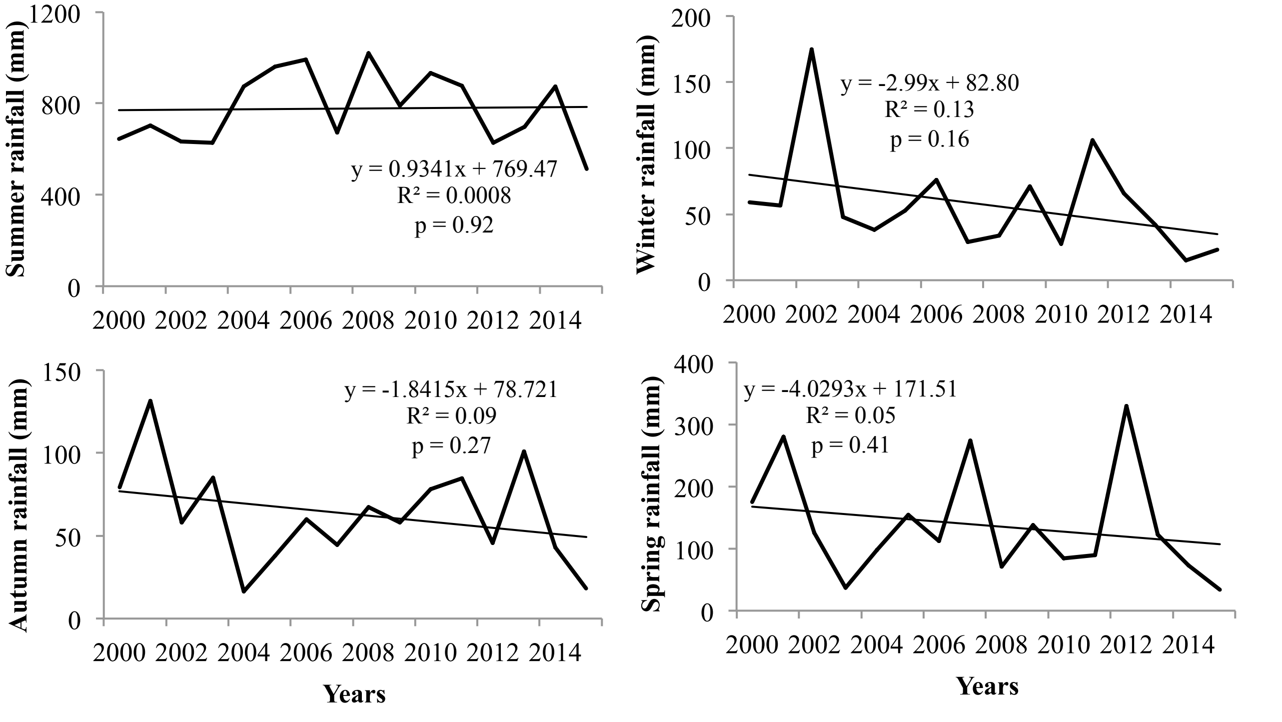


Figure 5.31 Time series and linear trends of summer, winter, autumn and spring rainfall for four combined stations in the catchment for the period 2000-2015.

5.4.5 Southern Oscillation and summer rainfall

A statistically significant correlation exists between summer rainfall in the catchment region and the Southern Oscillation Index (Table 5.5). All significance is computed to the 95% confidence level ($P<0.05$). The correlation coefficients between the summer rainfall and the preceding months are all above 0.3, with the highest lag correlation, $r=0.32$, during the June + July + August + September, June + July + August + September + October and July + August + September summer months.

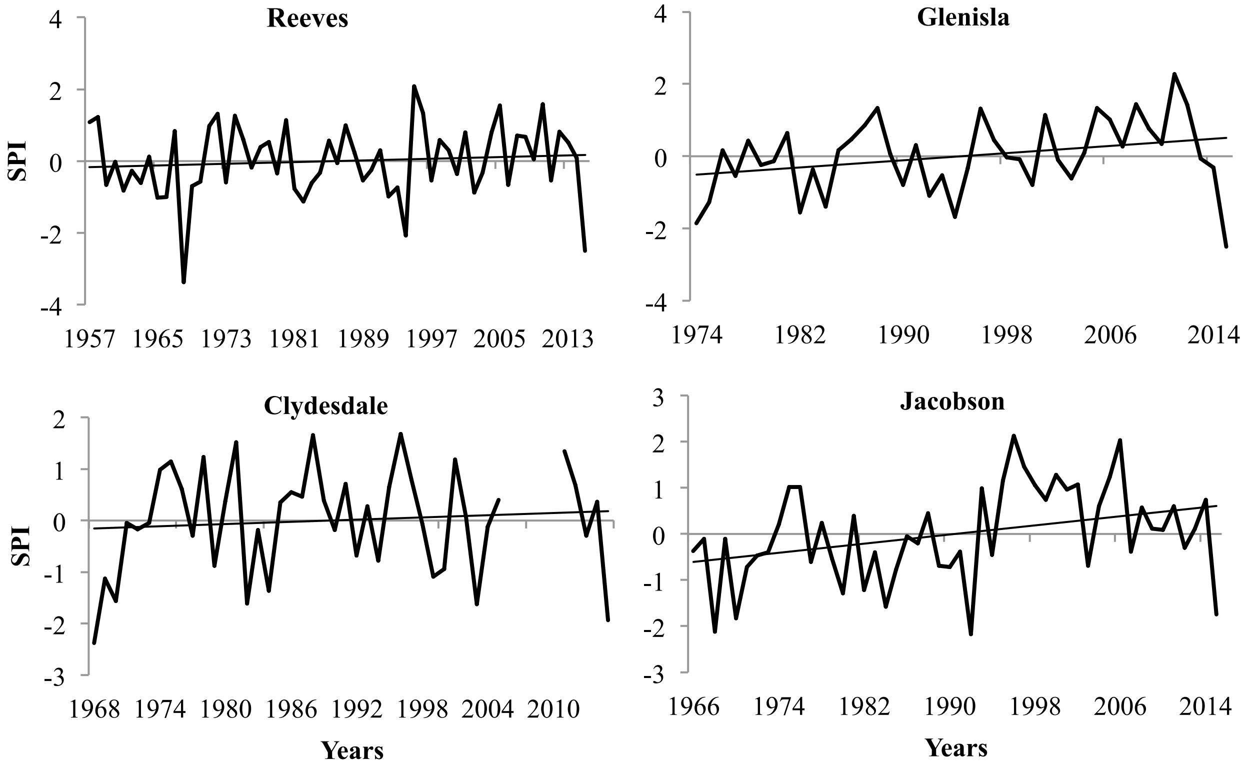
Table 5.5 Correlation coefficient *r* with the relevant level of significance *P* between standardized regional summer rainfall and the mean SOI values for certain periods.

Rainfall Period	Period of SOI values (non-lagged)	r	p
November - March	Nov + Dec + Jan	0.32	0.01
	Nov + Dec + Jan+ Feb + Mar	0.30	0.02
	Period of SOI values (lagged)		
	May + Jun + Jul + Aug + Sep	0.24	0.04
	Jun + Jul + Aug + Sep	0.32	0.01
	Jun + Jul + Aug + Sep + Oct	0.32	0.01
	Jul + Aug + Sep	0.32	0.01
	Jul + Aug + Sep + Oct	0.31	0.01

5.5 Drought Analysis: Sterkspruit Catchment

5.5.1 Standard Precipitation Index

In 2015, all stations had low SPI values with Reeves and Glenisla both showing a very low SPI value (-2.50) resulting in extremely dry conditions (Figure 5.32). Jacobson (-0.74) and Clydesdale (-0.94) had severe dryness in 2015.



Station	r	p
Glenisla	0.09	0.05
Reeves	0.01	0.45
Jacobson	0.13	0.01
Clydesdale	0.33	< 0.001

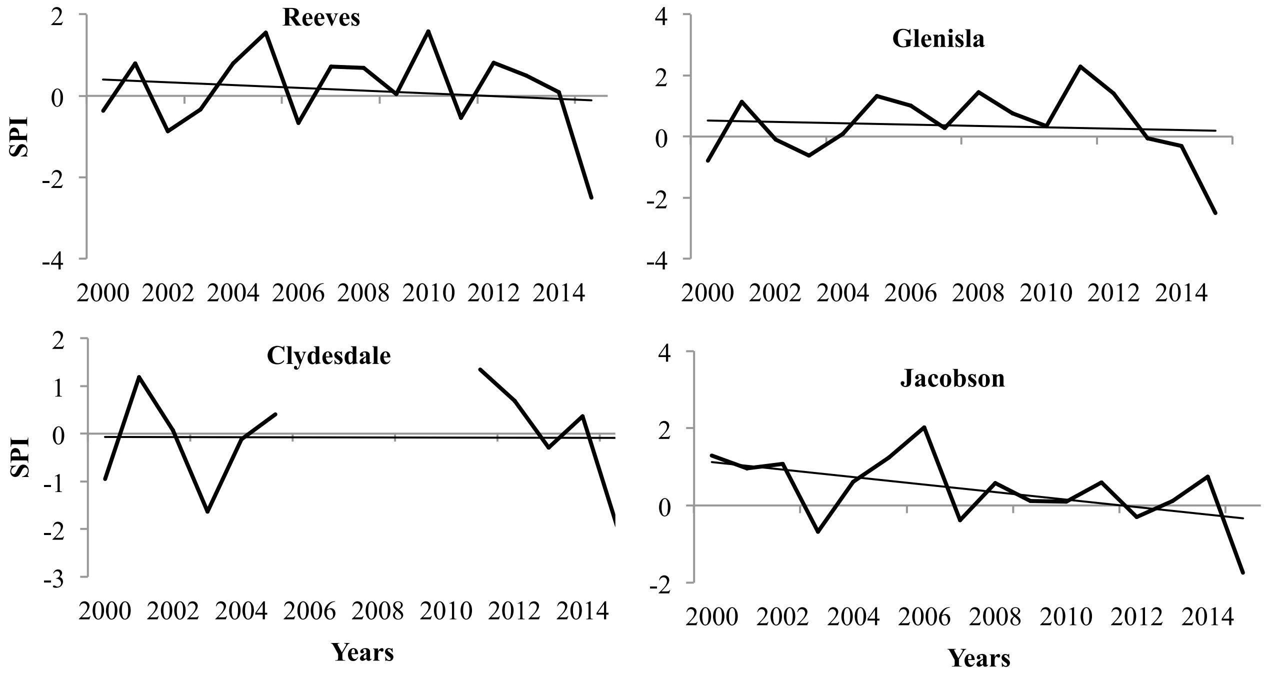
Figure 5.32 Time series analysis of the yearly SPI values for four stations in the catchment for the period 1974-2015.

On average, near normal conditions are dominant across the catchment stations, occurring 66.7% of the time over the 42-year period (Table 5.6). Following the general trend for the Drakensberg, the dry years experienced at the catchment level include 1994, 2003, 2005, 2007 and 2009 as moderately dry years, 1982, 1984 and 1994 as severely dry years and lastly, 2015 was an extremely dry (Table 5.6). The wettest year on record occurred is 1996. Both extremely wet and extremely dry years occurred 2.4% of the time and severely dry years occurred 11.9% of the time, in comparison to very wet years, which did not occur at all (Figure 5.35). Moderately wet years (9.5%) were experienced more than moderately dry years (2.4%).

Table 5.6 Percent occurrence of drought categories for the average Standard Precipitation Index for the four stations in the catchment.

SPI Value	Drought category	% Occurrence 1974 - 2015	% Occurrence 2000 - 2015	Years
> 2.00	Extremely wet	2.4	0.0	1996
1.5 to 1.99	Very wet	0.0	0.0	None
1 to 1.49	Moderately wet	9.5	12.5	1988, 1995, 2001, 2011
- 0.99 to 0.99	Near normal	66.7	56.3	28 of total
- 1.00 to -1.49	Moderately dry	2.44	25.0	1994, 2003, 2005, 2007, 2009
-1.50 to -1.99	Severely dry	11.9	0.0	1982, 1984, 1992
< - 2.00	Extremely dry	2.4	6.3	2015

Analysis of the SPI trends for the period 2000-2015 depicts all stations having decreasing trends over time (Figure 5.33). For this time period, only Jacobson's trend was found to be significant to the 95% confidence level (P = 0.04). All stations had low SPI values in 2003, 2007 and 2015. Over this time period, 25% of the years were moderately dry and only 12.5% were moderately wet (Table 5.6). A higher frequency of dry years occurred, as opposed to wet years, with extremely dry years occurring 6.3% of the time (Figure 5.34),



Station	r	p
Glenisla	0.01	0.73
Reeves	0.03	0.56
Jacobson	0.26	0.04
Clydesdale	0.00	0.98

Figure 5.33 Time series analysis of the average SPI values for four stations in the catchment for the period 2000-2015.

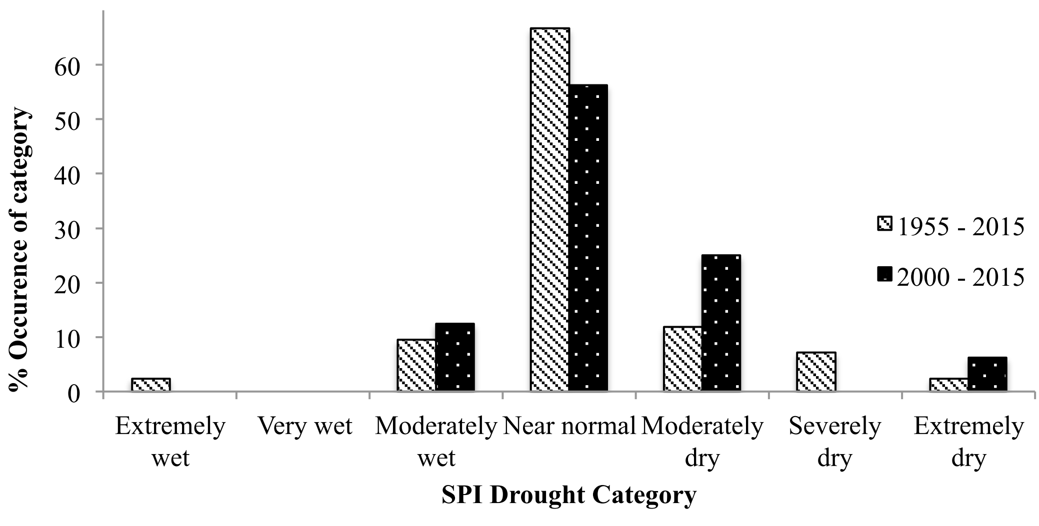


Figure 5.34 Percent occurrence of the SPI drought categories for the periods 1955-2015 and 2000-2015 for the average SPI values of the four stations in the catchment.

5.6 Hydrological Data

5.6.1 Discharge on the Mlambonja River

The Mlambonja River has an overall increasing discharge trend from 1977-2015, however, from 2000-2015 there is an overall decreasing trend in discharge over time (Figure 5.35). Neither of these trends is significant to the 95% confidence level. The discharge time series follows the time series trend of the Drakensberg rainfall (Figure 5.36). The years with the lowest average recorded discharge are 1983 (10.41 m^3/s), 1992 (6.27 m^3/s) and 2015 (9.4 m^3/s) and these years correlate to low rainfall experienced during the same period.

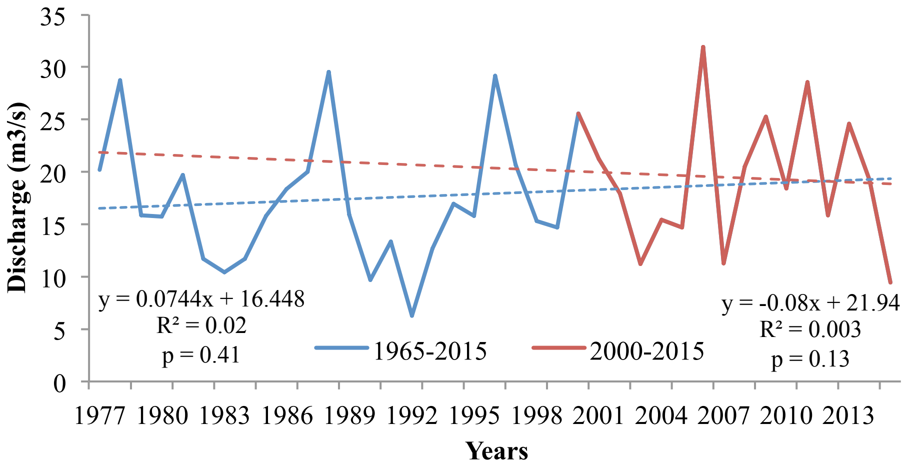


Figure 5.35 Average annual discharge of the Mlambonja River over the period 1977-2015.

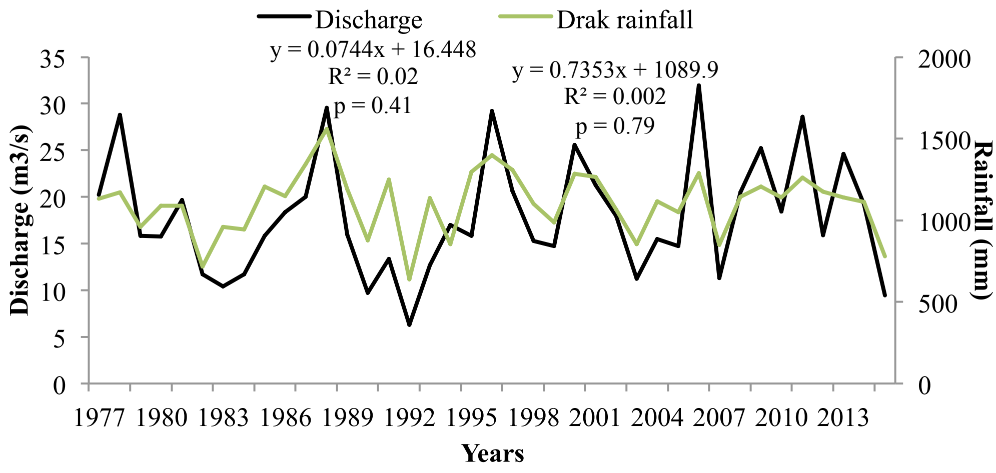


Figure 5.36 Overall discharge from the Mlambonja River in relation to yearly rainfall totals in the Drakensberg for the period 1977-2015.

The seasonal discharge at the Mlambonja River depicts an increasing winter and summer discharge over time, although these trends are not statistically significant (Figure 5.37). The average Drakensberg rainfall correlates highly significantly (P<0.001) to the average discharge and the average summer discharge of the Mlambonja River (Table 5.7). Summer rainfall in the Drakensberg also has a significant relationship to the summer discharge. In winter, the only significant correlation is that of the winter Drakensberg rainfall and the winter streamflow discharge (P=0.001).

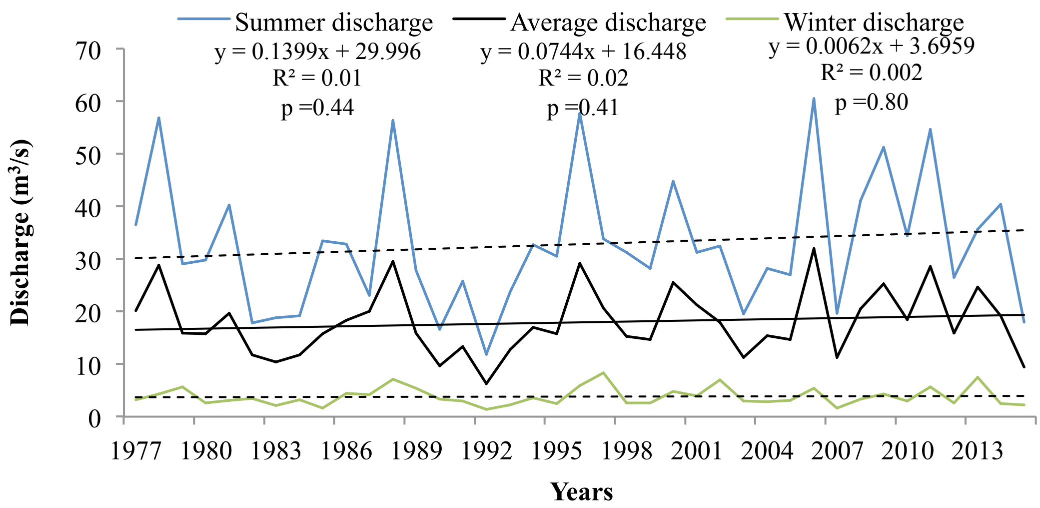


Figure 5.37 Average overall discharge in relation to summer and winter discharge of the Mlambonja for the period 1977-2015.

Table 5.7 The Mlambonja River discharge correlated statistically to the average, summer and winter rainfall in the Drakensberg study region.

Area	**Mlambonja River**	**r**	**p**
Drakensberg	Av Drak rainfall vs Av discharge	0.76	<0.001
	Av Drak rainfall vs Av summer discharge	0.70	<0.001
	Av Drak rainfall vs Av winter discharge	0.51	0.001
	Av Drak summer rain vs Av summer discharge	0.70	<0.001
	Av Drak winter rain vs Av winter discharge	-0.06	0.73

5.6.1.1 Hydrological Drought

The streamflow drought index at Mlambonja River indicates a trend of increasing wet (non-drought) years over time, although this is not significant to the 95th confidence level (Figure 5.38). There is an average cyclicity of 10 years between the prominent drought years, namely, 1983-1995, 1995-2007, 2007-2015. In the 38-year time period, 18 of those years had an SDI value of 0.00 or less, of which 44% are classified as mild droughts and 22% are considered to be moderate droughts (Table 5.8). A high percentage (17%) of years were classified as severe droughts and the prominent years include 1983 (-1.88), 1992 (-1.66) and 1995 (-1.62). From 1995, the years have higher SDI values, on average, but 2007 and 2012 are classified as moderate droughts and 2015 as mild.

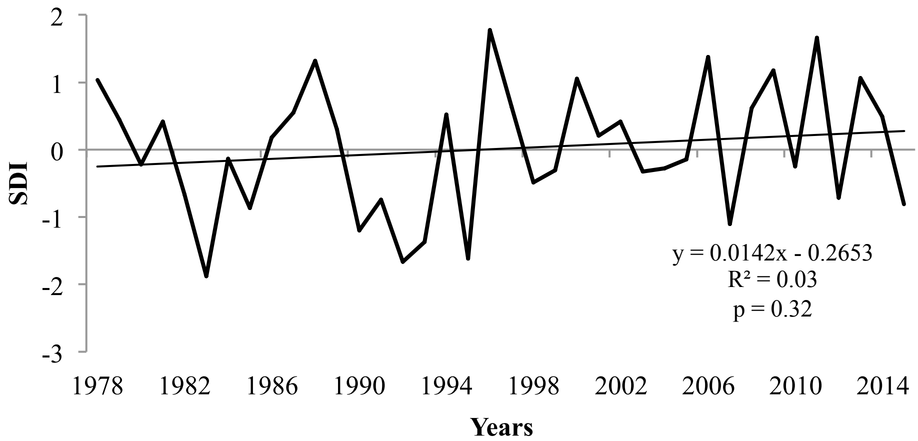


Figure 5.38 Streamflow Drought Index for the Mlambonja River from 1978-2015.

Table 5.8 Streamflow Drought Index categories for the Mlambonja River.

Description	SDI Categories	% Occurrence 1978 - 2015	% Occurrence 2000 - 2015	Years
Mild	0.99 to 0.0	44	71	1980, 1982, 1985, 1991, 1998, 1999, 2003, 2004, 2005, 2010, 2015
Moderate	-1.0 to -1.49	22	29	1990, 1993, 2007, 2012
Severe	-1.5 to -1.99	17	0	1983, 1992, 1995
Extreme	≤ -2.0	0	0	None

The SDI analysed from 2000-2015 shows a decreasing trend over time, indicating an increase in drier years over time, but this trend is not statistically significant (Figure 5.39). Of the 16 years, seven had SDI values above 0.00, of which 71.4% of the years are classified as mild droughts and 28.6% as moderate droughts (Table 5.8). Thus, analyses this century indicates a slight trend towards increased dry years.

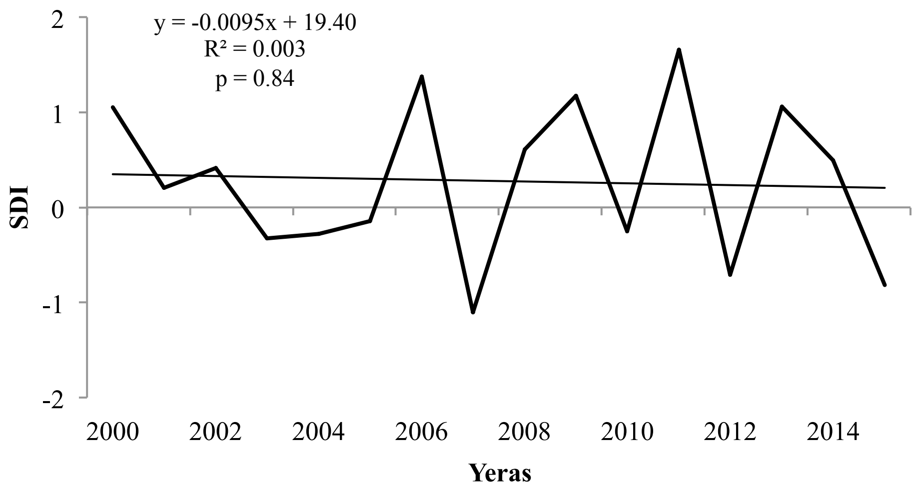


Figure 5.39 Streamflow Drought Index for the Mlambonja River from 2000-2015.

5.6.2 Discharge on the Little Tugela River

The discharge at the Little Tugela River station is decreasing over time, although this trend is not significant to the 95% confidence level (Figure 5.40). The times of low flow occurred during 1984 (7.74 m^3/s), 1992 (5.45 m^3/s), 2003 (6.66 m^3/s) and 2015 (5.62 m^3/s), with 2015 being the lowest recorded discharge over the entire period. The overall discharge linear trend from 1980-2015 indicates an increasing trend of discharge and the trend from 2000-2015 shows a decreasing trend in discharge, but neither are significant to the 95% confidence level.

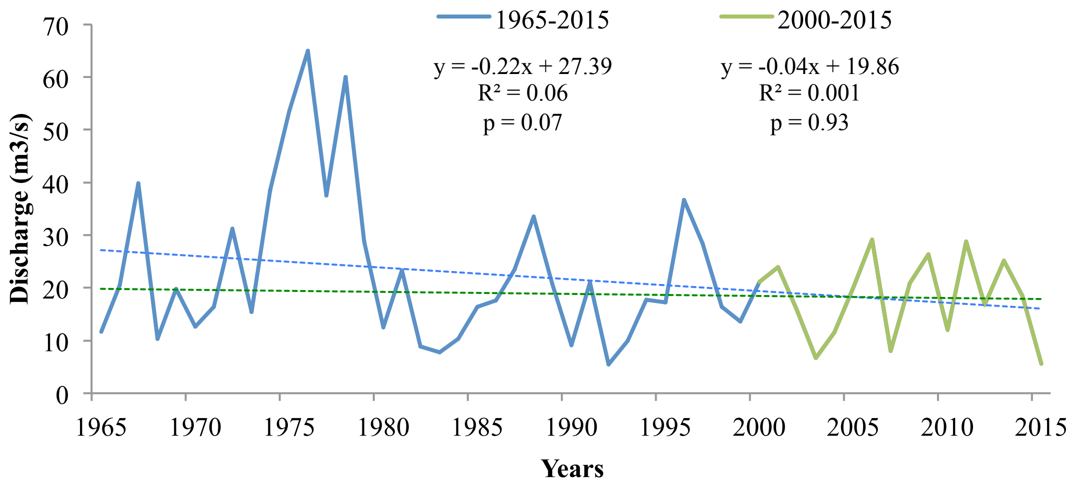


Figure 5.40 Average annual discharge of the Little Tugela River for the period 1965-2015.

The discharge of the Little Tugela follows the rainfall trends of the Drakensberg and the catchment area over time (Figure 5.41). The times of low rainfall and high rainfall are reflected by the discharge. High rainfall resulted in increased discharge flows and this occurred particularly in 1976 (65 m^3/s) and 1978 (60 m^3/s), where the discharge experienced was high.

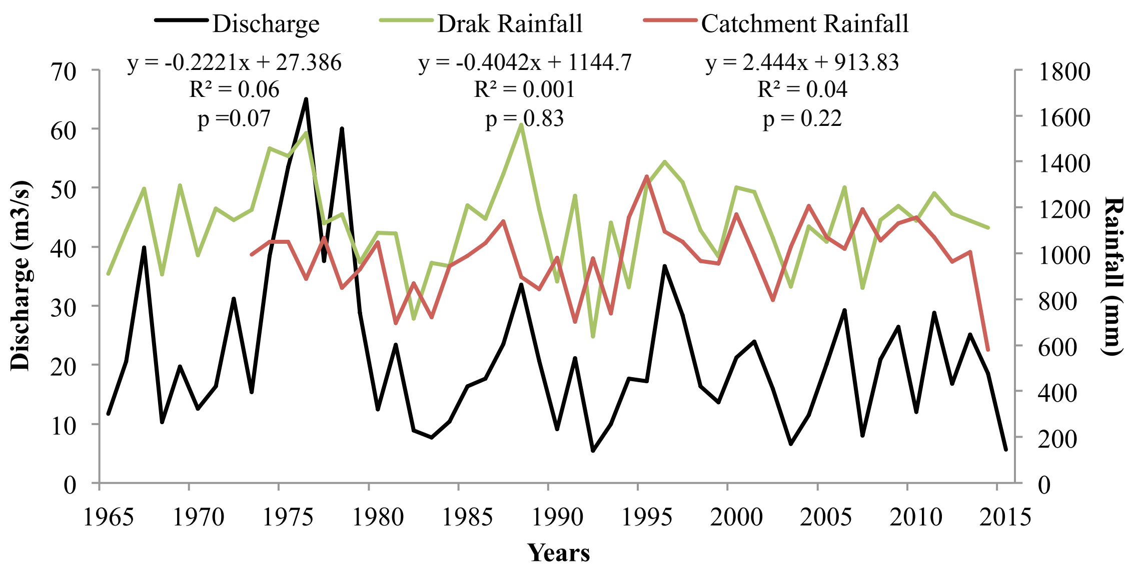


Figure 5.41 Overall discharge from the Little Tugela in relation to yearly rainfall totals in the catchment and Drakensberg for the period 1974-2015.

The overall summer discharge has a significant decreasing trend over time (P=0.05) (Figure 5.42). On inspection, both the summer and winter discharge patterns reflect the same trend as the discharge, except for 2011, where the winter discharge (62 m^3/s) was above the average discharge and the summer discharge (7.4 m^3/s) was below the average discharge of 29 m^3/s. that year.

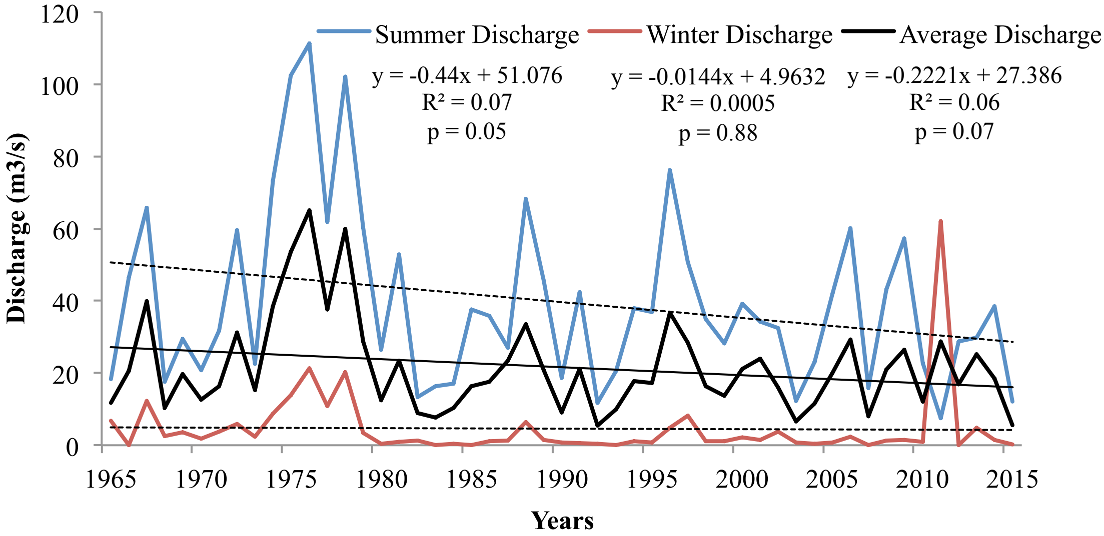


Figure 5.42 Average overall discharge in relation to summer and winter discharge of the Little Tugela for the period 1965-2015.

The overall combined stations' average rainfall for the Drakensberg correlates significantly to the average overall discharge and average summer discharge of the Little Tugela River ($P<0.001$) (Table 5.9). The catchment rainfall also correlates with the average overall discharge and the summer discharge of the Little Tugela, with a significance of P=0.002 and P=0.003, respectively. There is a highly significant relationship between the Drakensberg and catchment summer rainfall and the overall summer discharge ($P < 0.001$). In winter, the average Drakensberg rainfall and winter discharge of the Little Tugela River correlate significantly to the 95% confidence level ($P = 0.02$) (Table 5.10).

Table 5.9 The Little Tugela discharge correlated statistically to the average, summer and winter rainfall in the Drakensberg study area and the catchment study area.

Area	Little Tugela	r	p
Drakensberg	Av Drak rainfall vs Av discharge	0.69	<0.001
	Av Drak rainfall vs Av summer discharge	0.65	<0.001
	Av Drak rainfall vs Av winter discharge	0.32	0.02
	Av Drak summer rain vs Av summer discharge	0.68	<0.001
	Av Drak winter rain vs Av winter discharge	0.19	0.23
Catchment	Av Catchment rainfall vs Av discharge	0.47	0.002
	Av Catchment rainfall vs Av summer discharge	0.45	0.003
	Av Catchment rainfall vs Av winter discharge	0.27	0.09
	Av Catchment summer rain vs Summer discharge	0.68	<0.001
	Av Catchment winter rain vs Winter discharge	0.13	0.37

5.6.2.1 Hydrological Drought

In order to ascertain the years that are classified as hydrological drought, the streamflow drought index (SDI) was plotted against time. Of the 50 years analysed, only 24 years had an SDI of 0.00 or less. The overall trend shows decreasing SDI values over time, indicating an increase in drier years, however this trend is not significant (Figure 5.43). The majority of the dry years (70.8%) were classified as mild droughts (Table 5.10). Moderate drought years occurred 20.8% of the time and both severe and extreme droughts occurred 4.2% of the time. Of the moderate droughts, 1992 (-1.18), 1993 (-1.47), 2003 (-1.31), 2007 (-1.44) and 2015 (-1.43) are consistent with the dry years identified in the previous analysis. There is an average cyclicity of 11 years between the major dry periods, namely, 1968-1980; 1983-1995, 1995-2007 and 2007-2015 (Table 5.11).

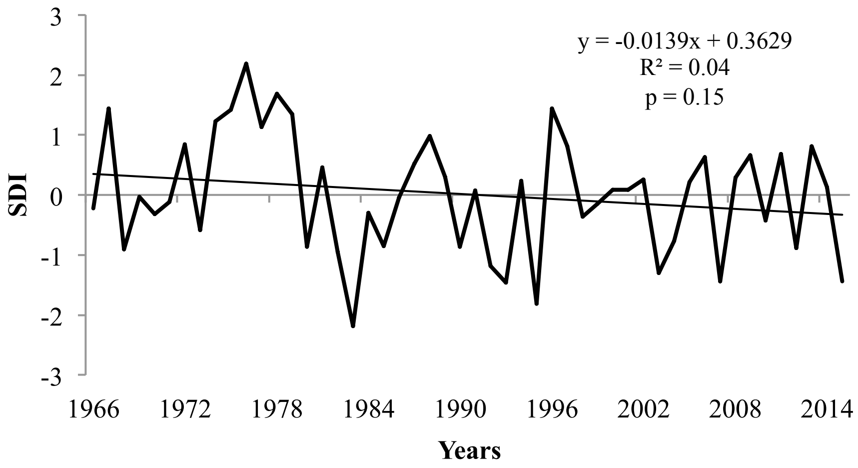


Figure 5.43 Streamflow Drought Index for the Little Tugela River from 1966-2015.

Table 5.10 Streamflow Drought Index categories, occurrence and years.

Description	SDI Categories	% Occurrence 1966 - 2015	% Occurrence 2000 - 2015	Years
Mild	0.99 to 0.0	71	50	1966/68/69/70/71/73, 1980/82/84/85/86, 1990/98/99 2004/2010/2012
Moderate	-1.0 to - 1.49	21	50	1992, 1993, 2003, 2007, 2015
Severe	-1.5 to - 1.99	4	0	1995
Extreme	≤ -2.0	4	0	1983

The SDI analysed from 2000-2015 shows a decreasing trend over time, but this is not significant to the 95th confidence interval (Figure 5.44). Of the 16 years, only six years were classified as droughts, with 50% being found to be mild and 50% being considered as moderate droughts (Table 5.11).

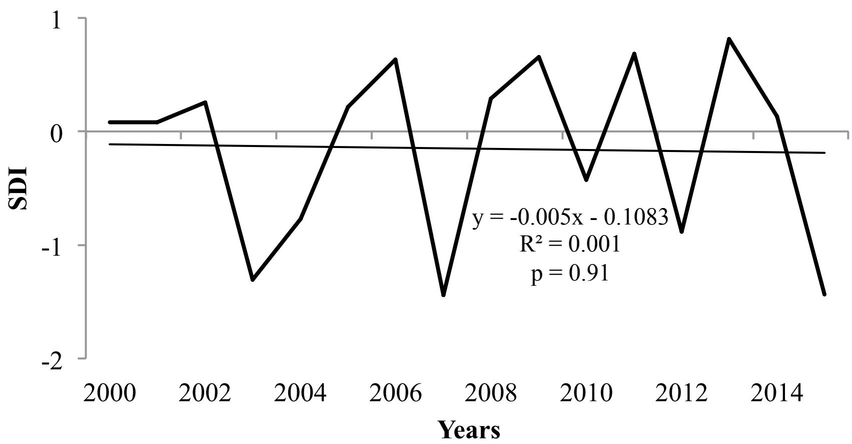


Figure 5.44 Streamflow Drought Index for the Little Tugela River from 2000-2015.

6. DISCUSSION

6.1 Introduction

Long-term rainfall trends for in the KwaZulu-Natal Drakensberg are discussed in the setting of the central and northern Drakensberg study region. Comparisons are then drawn against the Sterkspruit Catchment scale rainfall trends, in order to distinguish the potential localized, valley-scale trends in rainfall variability in the Drakensberg. Temporal trends of intra- and inter-annual rainfall variability of both annual and seasonal rainfall trends are explored and the relationship between summer rainfall in the Drakensberg and its correlation to the El Nino Southern Oscillation is explained. The research further contextualizes the results of rainfall trends on river discharge and on the occurrence and likelihood of meteorological and hydrologcal drought in the region.

6.2 Temporal Trends in Rainfall Variability

6.2.1 Interannual rainfall variability

The mean annual rainfall from 1955-2015 in the central and northern Drakensberg region ranges from 757 mm to 1338 mm. A statistically significant decrease in the mean annual rainfall, averaging to ±150 mm over this time period, is found. Mean annual rainfall of the combined stations has been decreasing consistently from the last half of the 20th century, to present. This trend is reflected across all individual stations, except Bergville, which only shows a decrease in rainfall post-2000. The decrease in rainfall at Giant's Castle, Cathedral Peak and Royal Natal stations are statistically significant. This decrease in overall MAR over time is in contrast to an earlier finding in this region of the Drakensberg by Nel (2009), where a non-significant increasing linear trend was evident in the last half of the 20th century in the Drakensberg. It could thus be assumed that the decreasing rainfall trends, identified this century, have resulted in the overall decrease in rainfall from 1955-2015, consequently highlighting the increased interannual rainfall variability occurring over time, but notably during the beginning of the 21st century. Nel's (2009) work provides a framework for comparison in the study area. However, a direct comparison of the studies is not made, owing to differences in the years and rainfall stations used. Thus the general results found in this study, will differ from the overall results found by Nel (2009).

The declining rainfall in the central and northern Drakensberg compares well to rainfall trends measured across South Africa. Kruger and Nxumalo (2017) saw an increase in rainfall in the western interior and a decrease in rainfall in the north-eastern parts of South Africa from 1921-2015. Decreases in the east are more pronounced in the study by MacKellar et al. (2014), which could indicate that decreases in the rainfall in this particular area have become more pronounced and significant recently, since the latter study focuses on more recent decades (1960-2010). It is evident that regional variability is present across the broader Drakensberg region. MacKellar et al. (2014) found an increase in rain days in the summer for the southern Drakensberg (1960-2010) and Nel and Sumner (2006) reported that there is higher interannual rainfall variability in the northern Drakensberg (compared to the south).

Rainfall trends measured across the central and northern Drakensberg are not reflected in the Sterkspruit Catchment (over the full time period), demonstrating that there is a localized effect of rainfall variability in the region. The Sterkspruit Catchment shows a trend towards decreasing rainfall this century, as opposed to the full period of analysis (1974-2015). It is thus evident that a decrease in rainfall, from the late 20th century and early 21st century, is occurring. Previous research conducted during the 20th century (1946-1988) over southern Africa, found no significant trend towards decreases in summer rainfall (Richard et al., 2001), whereas more recent research identified declining mean annual precipitation for the latter part of the 20th century (1980-2005), as opposed to the former (1950-1975) (Gaughan and Waylen, 2012). The year 2015 had the lowest MAR, in the Sterkspruit Catchment, over the whole time period of analysis and it was one of the driest years experienced over time across the Drakensberg. With regard to the annual rainfall total in the summer rainfall season of South Africa, it was found that 2015 was the year with the lowest annual total rainfall in 112 years for South Africa, resulting in the development of drought conditions across the country (SAWS, 2016).

Increasing interannual rainfall variability across the Drakensberg is further pronounced through stations having a statistically significant absolute rainfall deviation and rainfall deviation from the mean values (the latter of which is found in Appendix A). Giant's

Castle has experienced high variability from 1955-2015, which is shown by the statistically significant absolute rainfall deviation and rainfall deviation values. Cathedral Peak and Royal Natal also show increased interannual variability through significant rainfall deviation values from 1955-2015, as well as two stations in the Sterkspruit Catchment, Jacobson and Glenisla (Appendix A). This increase in variability is in contrast to the previous work in the Drakensberg area, where no stations (including Giant's Castle) were found to have significant interannual variability from absolute rainfall deviation (Nel, 2009). Although the absolute deviation trends this century have been increasing at Giant's Castle, they are not statistically signifcant. However, it could still be assumed that trends occurring this century have influenced the variability of rainfall trends from 1955-2015. This further indicates that increased interannual variability in this area is a recent phenomenon. Historical records across South Africa also indicate that interannual rainfall variability is increasing (Mason, 1996; Nicolson, 2000; Fauchereau et al., 2003).

The mean annual rainfall across the Drakensberg stations was found to increase with altitude and this is pronounced at the catchment scale, where there is a statistically significant relationship of increasing rainfall with altitude. The Reeves farm is situated at the lowest altitude and receives the least rain. The Jacobson household has the highest altitude and receives the most amount of rain, however, there are no farms at this altitude. This explains why the farms, which are located at lower altitudes in the catchment, are heavily reliant on irrigation. Previous work in the area established that for areas below 2100 m a.s.l. in the Drakensberg, the mean annual rainfall is strongly related to altitude and the location with respect to the escarpment (Nel and Sumner, 2006). This highlights the importance and need for water storage units, such as farm dams or reservoirs, in order for farms in low altitude areas to irrigate their land.

6.2.2 Cyclicity of rainfall

Much of the summer rainfall area of South Africa experiences a quasi-20-year rainfall oscillation (Vines, 1980; Tyson et al., 2002; Rouault and Richard 2003, 2005; du Toit and O'Connor, 2014). Previous research in the Drakensberg indicates rainfall oscillations between 16 and 20 years (Tyson et al., 1975) and, more recently, 10 and 20 year oscillations (Nel, 2009). Average annual rainfall across the central and northern Drakensberg, as well as at the catchment scale, shows an average cyclic oscillation of 10-

12 years, which compares more closely to the findings from Nel (2009) than with earlier findings (Tyson et al., 1975). This cycle also concurs with recent research, which found South African rainfall to exhibits an approximately 12-year oscillation over the period 1960–2010 (Jury, 2015). The years between rainfall cycles (peaks and troughs) in the Drakensberg thus appear to be becoming increasingly smaller and more unpredictable over time, which subsequently has implications for regional water management strategies and local farmers, who need to prepare for times of low rainfall. This 10-12 year rainfall cycle needs to be confirmed and investigated in further research.

Small-scale fluctuations in rainfall show that minor cyclicity occurs in the Drakensberg and in the Sterkspruit Catchment area approximately every three years. Tyson et al. (1976) also found minor rainfall cycles in the Drakensberg between 2-3 and 3-4 years and, in a broader context, rainfall cycles of 2-7 years in southern and South Africa have also been identified (Vines, 1980; Mason and Jury, 1997; Jury and Levey, 1993; Kane, 2009; Malherbe et al., 2015). Thus, short-term cyclicitiy seems to have been constant over time, but the long-term cyclicity is becoming increasingly variable and thus requires further analysis in future studies.

6.2.3 Intra-annual rainfall and seasonality variability

The mean annual rainfall in the Drakensberg and at the catchment scale is strongly seasonal, as noted by the PCI, with 75% of the rainfall occurring in the summer months (November–March). This is in agreement with Nel (2009), but it differs marginally from the 70% value calculated by Tyson et al. (1976) for the period up to 1960. Along with a change in annual rainfall, an increase in the variability of the distribution of monthly rainfall can be seen. An increase in the seasonality of monthly rainfall in the Drakensberg region is evident from the statistically significant (95% confidence level) increase in PCI at two stations. Seasonality is strengthening in the Drakensberg region from 2000-2015, which is shown by a highly significant increase in the average station PCI, as well as four stations having statistically significant increases in PCI. An increase in the seasonality of monthly rainfall has thus strengthened this century. In the catchment area, this phenomenon is duplicated, where two stations and the average have statistically significant increases in PCI. These findings are in agreement with previous research conducted in the area for the second half of the 20th century (Nel, 2009), indicating a continuing trend of increasing seasonality of monthly rainfall in the Drakensberg.

A trend analysis of the four seasons in the Drakensberg region indicates a decrease in the average station rainfall over time for all seasons, with the autumn decline being statistically significant. This is further established by a lower percentage of rainfall, over time, occurring in autumn, rather than in winter, across the Drakensberg and the catchment area. In contrast, previous research in the region found that only winter and autumn rainfall was decreasing (Nel, 2009), again highlighting the overall decline in rainfall over time. A decline in autumn rainfall has previously been established across South Africa (MacKellar et al., 2014). All stations, except Bergville, have decreasing rainfall in autumn and Cathedral Peak and Giant's Castle have a significant decreasing trend to the 95% confidence level. In the KwaZulu-Natal Drakensberg, the active crop-growing season is from October to April (Schulze, 1979; Nel, 2009) thus, if this decrease in autumn rainfall persists, it could affect the late crop-growing season in this crucial area. A decrease in summer rainfall is strong evidence for the increased number of dry years that have occurred this century in the Drakensberg. A decrease in the autumn and winter rainfall is continuing to result in a shorter wet season and a more pronounced seasonal cycle.

The Sterkspruit Catchment displays localized variability, differing slightly from the Drakensberg data, where seasonal trends are increasing (except for spring) across the stations (not statistically significant). Glenisla is the only station with significant increasing summer rainfall from 1974-2015. This increasing summer rainfall suggests that there is an increase in the seasonality of rainfall in the area, which again emphasizes the presence of localized variability. There is, however, a trend towards decreased seasonal rainfall occurring this century, where there is a non-significant decrease in rainfall in all seasons across the catchment, except for summer. Jacobson station has had decreasing trends this century for all seasons, except winter, as opposed to the 1974-2015 trends. All stations, except for Reeves, have had declining autumn rainfall post-2000.

Despite the increases in summer rainfall over time, farming in the Sterkspruit Catchment area is still reliant on irrigation in summer. For example, Glenisla Farm, where one of the stations is situated, is a dairy farm and dairy farmers in the catchment area generally only rely on dryland crops (rainfed) for young animals in summer, but the feed is produced mainly off irrigation, as the rainfall is known to be very unreliable (Stockil, pers.comm.). Maize silage is ideally grown in summer and is used in autumn, winter and spring to balance the shortfall of feed grown in the summer. However, despit growing in the summer, most silage is also grown under irrigation from the Bell Park Dam (83 of 94 ha at Glenisla farm) (Stockil, pers.comm.).

Crop farmers in the area plant wheat in winter, with water from the Bell Park Dam, and their lands lie fallow in autumn (Stockil, pers.comm.). The best time to plant soya beans and maize (the main crops in the catchment area) is October and November. Therefore if the early seasonal rain does not fall (or not enough falls) the farmers plant their irrigated crop first, using the Bell Park Dam water, and the dryland crops last, when it eventually rains. Thus, using irrigation from the Bell Park Dam mitigates reduced rainfall in autumn and spring. Reliance on the Bell Park Dam is paramount, which is why, if it runs dry, as it did in 2015, the consequences are severe. Farmers would probably not change crop mixes or planting dates, if it rained more, but they would be able to produce more kilograms on feed on the dryland hectares, as the grass does not grow if it is dry (Stockil, pers.comm.). Farmers would also save on electircity payments that is costly in the irrigation process.

6.3 Effects of Rainfall Variability on River Discharge

The variability of rainfall in the Drakensberg affects the streamflow of rivers, with wet years reflecting a higher discharge and dry years reflecting a low discharge, which is in agreement with previous research in the area (Everson, 2001; Nel and Sumner, 2006). Discharge at the two rivers in the Drakensberg correlates statistically significantly to the rainfall in the region. Years of low discharge reflect those years identified with low mean annual rainfall, namely, 1983, 1992, 1994, 2003, 2007 and 2015. Years of low discharge have become increasingly prominent in the Drakensberg this century. This is suggested from the discharge trends, post-2000, at the Mlambonja River having an opposing decreasing discharge trend and a higher occurrance of hydrological drought years this century. The seasonality of rainfall is reflected in the discharge at the two rivers and the flow patterns are consistent with the summer rainfall regime, where summer discharge is above average and winter discharge is low and below average. Similarly, a study of streamflow characterisitcs on rivers in the Incomati River Basin, reported that the discharge patterns follow those of the summer rainfall regime (Okello, 2015).

The Little Tugela is found at the end of the Sterkspruit Catchment and therefore reflects the effects of rainfall variability on streamflow discharge at a valley-scale in the Drakensberg. Streamflow discharge at the Little Tugela River has been continually decreasing over time and the lowest discharge occurred in 2015, mirroring the national status of rainfall deficit that year (SAWS, 2016). This also highlights why the Bell Park Dam, situated on a tributary of the Sterkspruit River that feeds the Little Tugela, ran dry in 2015.

6.4 The Southern Oscillation and Summer Rainfall

The Southern Oscillation influences the summer rainfall variability of the KwaZulu-Natal Drakensberg with a strong statistically significant correlation existing between the summer rainfall in the Drakensberg and the SOI. This correlation between summer rainfall and ENSO events suggests that an increase in the frequency and intensity of ENSO will negatively affect the rainfall in the Drakensberg. Due to the persistence of the SOI, a statistically significant correlation between summer rainfall also exists and the SOI for preceding periods lagged at least one month. The highest lagged correlation ($r=0.45$) between summer rainfall and SOI is obtained from the SOI values for July to October and this correlates with work done by Nel (2009) for 1955-2000. Hydén and Sekoli (2000) also established a correlation between the November and December rainfall in the Lesotho Lowlands and the SOI values of July to September ($r=0.51$). Over southern Africa, the corrrelation between the Southern Oscillation and the summer rainfall zone was significantly positive from 1867-2006 (Neukom et al., 2014).

Studies in South Africa have found that droughts are caused by El Nino events (Rouault and Richard, 2005; Edossa et al., 2014; Malherbe et al., 2015). ENSO is not the only driver of rainfall variability, as only 30% of rainfall variability in South Africa can be explained by changes in ENSO (Tyson and Preston-Whyte, 2000). Notwithstanding this, the data cleary shows that a lagged correlation between summer rainfall in the Drakensberg and the SOI exsits that can be used as an indicator for seasonal forecasting. This phenomenon is also reflected at the catchment level, indicating that ENSO contributes to localized rainfall variability and is therefore a dominating feature behind the decrease in summer rainfall across the central and northern Drakensberg.

6.5 Drought

6.5.1 Identification and frequency of drought events

Droughts are recurrent features in the central and northern Drakensberg. During the period of analysis, the dry-spells experienced across the central and northern Drakensberg occurred from 1965-1973, 1982-1994, 2003-2007 and 2007-2015. According to the SPI, the largest meteorological droughts, in the severe to extreme categories, occurred in 1982, 1992, 2003, 2007 and 2015. For the period ending in 2000, these agree with earlier findings in South Africa (Rouault and Richard, 2003), southern Africa (DEWFORA, 2012) and on the African continent (Masih et al., 2014), where droughts in 1982/1983 and 1991/1992 and 1994/1995 were considered to be the most severe. The more recent major droughts in 2003 and 2007 correlate to those identified by Mussa et al. (2014), and including 2015, by Malherbe et al. (2015) for different parts of South Africa, with the years identified by DEWFORA (2012) for southern Africa, namely, 2002/2003/2004 and 2005/2006.

Similarly, the catchment area experienced severe droughts in 1982, 1984, 1992 and an extreme drought in 2015. The early 1980s drought is what prompted the farmers in the Sterkspruit Catchment to build the Bell Park Dam, however, the recent drought in South Africa in 2015, considered the worst since 1982 (Corke and Whittles, 2015), casued the Bell Park Dam to run dry. This phenomenon led to farmers having to make alternative plans for accessing water for the irrigation of their crops, such as locating groundwater and building other farm dams. Rainfall trends in the Sterkspruit Catchment vary slightly from the Drakensberg, exhibiting the presence of localized variability, where moderate droughts were felt in 1994, 2003, 2005, 2007 and 2009, and small-scale fluctuations in drought cyclicity are present in the early 2000s.

These major drought years correlate to those of low rainfall found across the stations over the Drakensberg region, therefore also indicating an average drought cyclicity of 10-years and small-scale cyclic variations occurring every ±3 years. This 10-12-year drought cycle concurs with cyclic variations found by Malherbe et al. (2015), based on spectral peaks, using the SPI found for the south-eastern parts of South Africa. Although, overall, the 17-20-year cyclic variation is dominant across South Africa and agrees with the ±18-year rainfall pattern (Tyson et al., 2002), an analysis by Malherbe et al. (2015) found an increase in the 10-12 year variation, specifically in the south-eastern parts of the country. This could explain the tendency towards smaller cycles in dry years occurring over time.

A correlation between low rainfall and meteorological drought is thus evident and an analysis of rainfall trends could seemingly be used to identify when drought will occur. Trends in drought cyclicity have been found to be consistent with those of the rainfall trends. Previous research has found that multi-decadal rainfall trends in southern Africa suggests a declining mean annual precipitation, increasing variability and drier conditions (Nicholson and Entekhabi, 1986; Mason and Jury, 1997; Mason, 2001; Giannini et al., 2008) which have been in part attributed to ENSO (Gaughan et al., 2016).

The SPI is linked to the probability occurrence of dry or wet events. Post-2000, a higher percentage of dry years than wet years is evident, on average, across the Drakensberg and catchment scale study region, indicating an increase in the number of dry years occurring this century. Furthermore, this is substantiated by an increase in the number of dry years over time, for all stations in the catchment area (no significant trends measured except at Jacobson) post-2000 and at one station in the Drakensberg (Bergville) region, where trends were previously increasing for the respective full timescales of analysis. Therefore, the increasing variability of rainfall this century is contributing to the increased number of dry years experienced over time. An increase in dry years over time has been previously identified over southern Africa, where there was an observed decrease in annual rainfall in the last quarter of the 20th century, stretching into the early 2000s (Gaughan and Waylan, 2012; Gaughan et al., 2016).

6.5.2 Hydrological drought

It is important to understand the hydrological impacts of decreased rainfall in an area such as the Drakensberg, especially the impacts on the features at the lower end of the hydrological spectrum, such as reservoirs, dams and streams. Agricultural activities are not only adversely affected by the direct impact of low rainfall, but also by water scarcity, from low reservoir and river levels or within the soil as a lowered water table (Malherbe et al., 2015). Farmers in the region rely on runoff and streams to refill their farm dams that are used for irrigation.

The years of hydrological drought identified from the Streamflow Drought Index, for the two stations in the Drakensberg match those of the low rainfall and meteorological droughts years. Severe droughts that occurred in 1983, 1992, 1993, 1995, 2003, 2007, 2015, correlate to those found by Trambauer et al. (2014) in the Limpopo River Basin and Malherbe et al. (2015) for South Africa. An average cyclicity of 10 years between the prominent drought years is evident at both discharge stations. The drought periods experienced at the discharge stations, reflect, to a close extent, the dry periods found by Malherbe et al. (2015) for South Africa, namely, 1965-1973, 1983-1994, and 2003-2007.

The Little Tugela River experienced drought conditions in the late 1960s to early 1970s, the early 1980s to the mid-to-late 1990s and early 2000s. An increase in the severity of drought at the Little Tugela is evident this century, where the occurrence of mild droughts decreased and the occurracne of moderate droughts increased, post-2000. The effects of the 2003 and 2015 droughts were also felt more intensely at the Little Tugela River than the Mlambonja River, therefore, the presence of localized variability is evident and it is felt specifically at the catchment scale. Streamflow at the Mlambonja River did not experience dry conditions in the early 1960s and 1970s, however, more small-scale fluctuations in the dry conditions from 2000-2015 was experienced. At the Mlambonja River, a higher occurrance of mild and moderate droughts were expereinced post-2000, indicating the increase the incident of droughts occurring this century. This follows the work by Prudhomme et al. (2014), who showed that a likely increase in the global severity of hydrological droughts will occur at the end of the 21st century.

6.5.3 Link between drought and ENSO events

In recent years (the latter half of the 20th century), there has been an increase in the number of warm phase ENSO events (Mason, 2001), which are often associated with a declining mean annual precipitation and increasing variability (Gaughan and Waylan, 2012). One possible explanation for the change in precipitation pattern and the possible ENSO phase frequency is a shift in the global coupled ocean-atmosphere system in the tropical Pacific, during the late 1970s, that saw increased sea surface temperatures and an eastward shift in Pacific convection (Graham, 1994; Mason, 2001; Gaughan and Waylan, 2012)

The mechanisms behind the decrease in summer rainfall can thus also be the cause of drought conditions. The El Nino phenomenon is therefore a causal link and contributor to the drought conditions experienced across the central and northern Drakensberg. All the drought years correlate to the years when southern Africa experienced the effects of El Nino (Climate Protection Centre, 2017). These findings are in agreement with previous work on droughts in South Africa. Rouault and Richard (2005) demonstrate that the ENSO conditions attributed to 8 out of the 12 droughts in South Africa occurring during the 20th century. Edossa et al. (2014) found 7 of 17 drought events identified succeeded El Nino events in central South Africa. Gaughan and Waylen (2012) found the El Nino event of 1982-1983 bought widespread drought during the summer rainfall season.

7. CONCLUSION

7.1 Introduction

The central and northern KwaZulu-Natal Drakensberg area is known to produce twice as much runoff per unit rainfall across South Africa and it is thus crucial for runoff generation, where the water that is captured feeds the water transfer schemes and supplies water to farm dams in the area. The study aimed to present findings on annual rainfall totals and intra- and inter-annual rainfall variability trends across five stations from 1955-2015 in the central and northern KwaZulu-Natal Drakensberg and hence to update and extend knowledge of rainfall characteristics in the area. In addition, rainfall variability was linked to the identification and frequency of meteorological and hydrological droughts in the area. The driving mechanisms behind low rainfall and subsequent droughts were explored, with the focus being on the El Nino Southern Oscillation. Also presented in this study are findings on four stations in the Sterkspruit Catchment area, in order establish the characterisitcs and presence of localized rainfall variability occurring within the greater Drakensberg region.

The Drakensberg is a summer rainfall dominated area but farmers in the area do not receive enough rainfall in summer to meet their desired agricultural needs. In response to this phenomenon, irrigation-fed farms, that rely on water from farm dams, have been established, as a farmers' adaption initiative to water shortages in times of drought. The recent drought of 2015 left the Bell Park Dam dry, which subsequently caused major consequences for the farmers that are reliant on the water. Thus the increasing variability of rainfall and the occurrence of drought in the Drakensberg, pose a problem to agriculture in the area.

7.2 Key Findings

Several key findings are presented on rainfall variability and drought in the KwaZulu-Natal Drakensberg and they are presented below:

- **The central and northern KwaZulu-Natal Drakensberg is a summer rainfall dominated area that is experiencing a statistically significant decline in mean annual rainfall over time (1955 – 2015).** This decreasing rainfall over time is reflected significantly at the Giant's Castle, Cathedral Peak and Royal Natal stations. Decreasing rainfall in the eastern and north-eastern part of South Africa are pronounced in studies by MacKellar et al. (2014) and Kruger and Nxumalo (2017). Increasing interannual variability is increasing over time, where Giant's Castle has experienced significant increasing variability over time, which is in contrast to previous research in the area from 1955-2000 (Nel, 2009). The SPI drought analysis concurs with the rainfall analysis, where these three abovementioned stations have statistically decreasing SPI significant trends, indicating that there is an increase (decrease) in the number of dry (wet) years occurring over time in the central and northern Drakensberg.

- **An increase in the variability of the distribution of monthly rainfall is present across the central and northern Drakensberg**. The seasonal rainfall trends are consistent with the annual rainfall trends and show decreasing trends across both time periods, with a significant decrease in autumn rainfall, from 1955-2015, at Giant's Castle and Cathedral Peak. The seasonality of Drakensberg rainfall this century is strengthening, where the analysis found a statistically significant increase in PCI post-2000, on average, at all stations, except at Giant's Castle. Analysis for 1955-2015 found Giant's Castle and Cathedral Peak showing an increasing seasonal variability. An increase in seasonality over time has previosuly been identified in the Drakensberg (Nel, 2009) and is thus continuing to increase in variability.

- **The Sterkspruit Catchment area highlights the localized variability of rainfall that occurs within the Drakensberg**. All stations have increasing annual rainfall from 1974-2015, with Glenisla and Jacobson being statistically significant. However, it is clear that a decrease in rainfall has been seen this century, as all stations have decreasing annual rainfall post-2000, with Jacobson being statistically significant. This trend is reflected in the SPI drought analysis data, consolidating the narrative of a pattern towards an increase in the number of dry years (less rainfall) this century. Gaughan and Waylen (2012) and Richard et al. (2001) show evidence supporting the declining mean annual precipitation trends and increased rainfall variability in the 21st century across southern and South Africa respectively. Although not significant, the average seasonal rainfall for all seasons over time reflects this trend. Furthermore, and similar to the Drakensberg data, the seasonality post-2000 is increasing at the Glenisla and Jacobson stations and this is expressed through statistically significant PCI values, indicating an increase in the variability of the distribution of monthly rainfall.

- **A decline in rainfall and an increase in the variability and seasonality of rainfall pose a threat to farmers in the Sterkspruit Catchment**. Glenisla farm has had a statistically significant increase in summer rainfall over time, but the farm is heavily reliant on irrigation during summer. The summer rainfall totals are not enough to fulfil the needs of the farm and the timing of the rainfall is unreliable (Stockil, pers. comm.). A decline in autumn and spring rainfall presents a problem to crop farmers who plant in October and November. If the early seasonal rains do not fall, they have to plant irrigation crops first, and when the rain finally comes, they plant dryland crops, which is a temperamental and uncertain approach to farming (Stockil, pers. comm.). Thus, using irrigation from the Bell Park Dam combats the reduced rainfall in autumn and spring, and should more rain fall in summer, farmers would be able to produce more dryland crops, such as feed for the animals. A decrease in autumn rainfall has previously been established across South Africa (MacKellar et al., 2014).

- **El Nino Southern Oscillation events affects the rainfall in the central and northern Drakensberg**. A strong correlation between the summer rainfall and the ENSO, and an increase in the frequency of ENSO, negatively affects rainfall in the Drakensberg. This phenomenon is linked to the decrease in rainfall and drier conditions experienced in the Drakensberg region over time as found by Nel (2009). Research in South Africa has found that droughts are driven by El Nino events (Rouault and Richard, 2005; Edossa et al,. 2014; Malherbe et al., 2015).

- **Droughts are recurrent features in the central and northern Drakensberg.** The largest meteorological droughts, according to the SPI, occurred in 1982, 1992, 203, 2007 and 2015. For the period ending in 2000, this is in agreement with Roualt and Richard (2003) and DEWFORA (2012), and the droughts experinced post-2000 agree with work by Mussa et al. (2014) and Malherbe et al. (2015) for different parts of South Africa.

- **The cyclicity of low rainfall events in the Drakensberg and the Sterkspruit Catchment occurs approximately every 10-12 years.** In comparison to previous research, which identified a 10-20-year cycle (Nel, 2009), the 10-12 year cyclicity found is shorter. The years of low rainfall correlate to the years of meteorological and hydrological drought identified by the SPI and SDI are, namely, 1982, 1992, 2003, 2007 and 2015. Small-scale cycles of ±3 years have remained consistent with past research (Vines, 1980; Mason and Jury, 1997; Jury and Levey, 1993; Kane, 2009; Malherbe et al., 2015), thus changes in the cyclicity are occurring over a greater timescale. Cycles between the low rainfall/drought years are therefore becoming shorter and more variable over time, advancing the notion of increased rainfall variability being experienced in the central and northern Drakensberg over time.

- **The streamflow in the Drakensberg follows the general rainfall and seasonal trends, where years of low rainfall reflect years of low discharge.** A trend towards decreasing discharge this century is evident at the Mlambonja River. The SDI reflects this trend, where there is an increase in the number dry years, shown by a higher occurrence of mild and moderate droughts exisitng from 2000-2015, compared to the full period of analysis. Streamflow in the catchment, as expressed by the streamflow discharge of the Little Tugela, shows a decreased discharge over time with the lowest discharge occurring in 2015, which explains why the Bell Park Dam ran dry. There is localized variability, expressed through the comparison of the two rivers' discharge data, where the overall discharge trends do not correlate, as well as some of the drought years identified by the SDI.

7.3 Implications

The study specifically highlights the importance of understanding the temporal patterns of rainfall, especially for decisions concerning future land use and water availability. Increased variability or persistently drier conditions demand effective cropping and adaption strategies (Thorton et al., 2004; Vetter, 2009) because changes in the current climate can impact the length of the growing season and the potential yield rates for various agricultural crops (Kotir, 2011). Areas with higher precipitation variability, such as the KwaZulu-Natal Drakensberg, may experience a more unpredictable yield production, due to inconsistency in the annual rainfall. The effects of ENSO on precipitation variability will also influence favourable cropping seasons, as warm and cold phases of ENSO will exacerbate existing dry (warm phase) and wet (cold phase) years of rainfall.

With the Drakensberg being an important primary water source, it is vital that trends in the variability of rainfall are understood. This is significant at a regional level where transboundary initiatives, such as the Tugela–Vaal transfer tunnel (TUVA) and the Lesotho Highlands Water Project (LHWP), were developed to transfer water from KwaZulu-Natal to Gauteng. Local farmers need assurance of water availability, as they are reliant on irrigation throughout the year. This research helps to provide feedback to farmers, which will allow for them to be cognisant of the historical shifts in rainfall, and by implication streamflow, in order to aid adaptive strategy decision-making, such as whether to build more farm dams or not, as rainfall is unreliable.

7.4 Recommendations

This work sets the background for future studies on rainfall variability in the KwaZulu-Natal Drakensberg. These strategies rely on a knowledge of temporal patterns of past and present precipitation and how shifts in climate may influence the timing, distribution or frequency of rainfall. Comprehensive studies on historic drought events could significantly guide better planning and mitigation strategies with regards to droughts. More research is also necessary to determine how short-term variability in rainfall patterns fit within the longer term changes of the climate. The prediction of the recurrence of rainfall cycles could also be explored further, in order to asses if the shortened cyclicitiy of 10-12 years can be expected in the future.

Additionally, it is pointed out that temperature regimes in the Drakensberg area, alongside drought, were not considered. For the purposes of this study, however, the definition of meterological and hydrological drought used was evaluated mathematically and is temperature independent. It is acknowledged that future studies could involve the consideration of temperature regimes in drought analysis. The meteorological conditions and weather systems associated with drought have not been analysed further than the causal association with ENSO. This is beyond the scope of the study, however, it is acknowledged that it needs to be analysed further in future studies.

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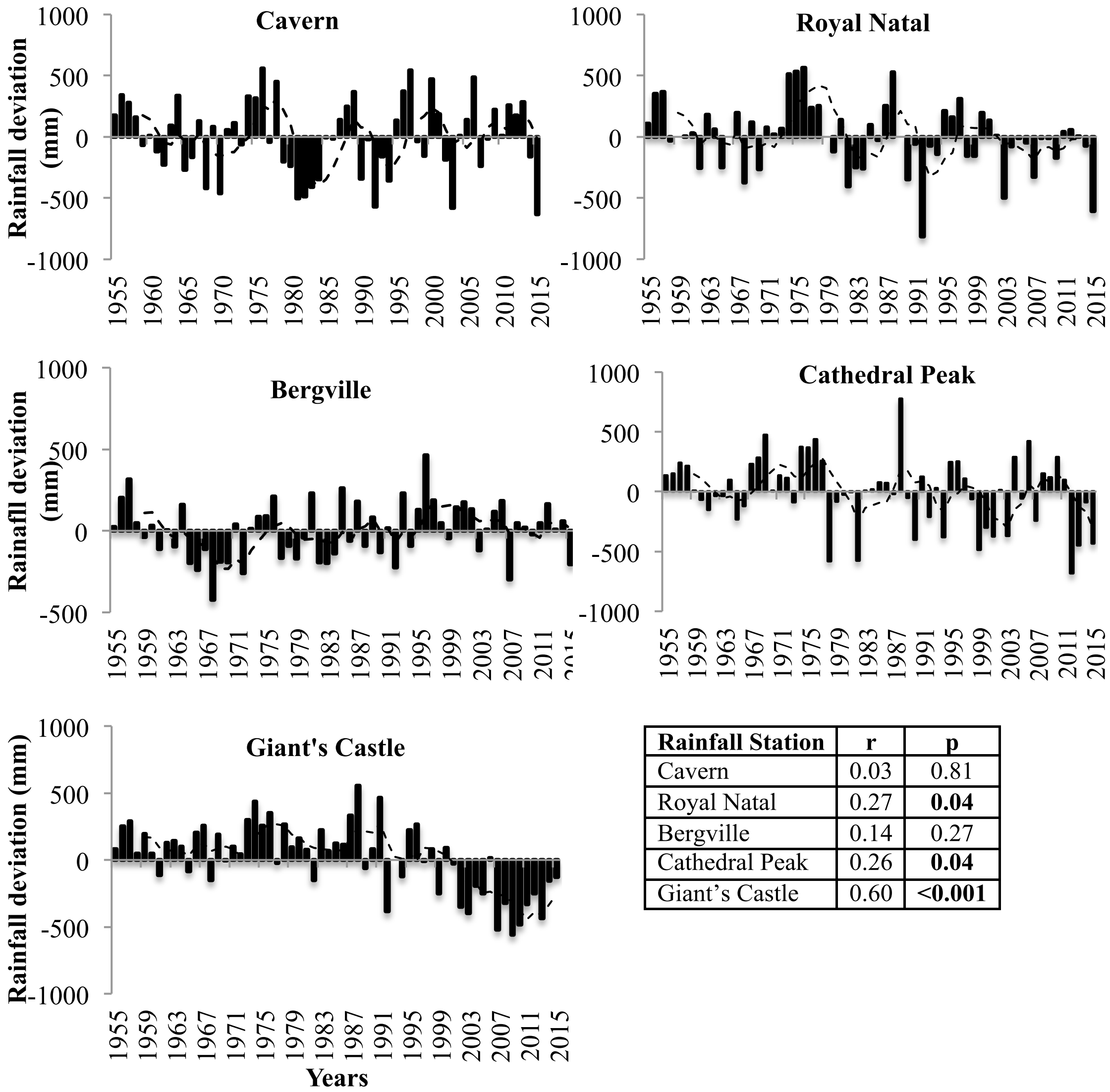
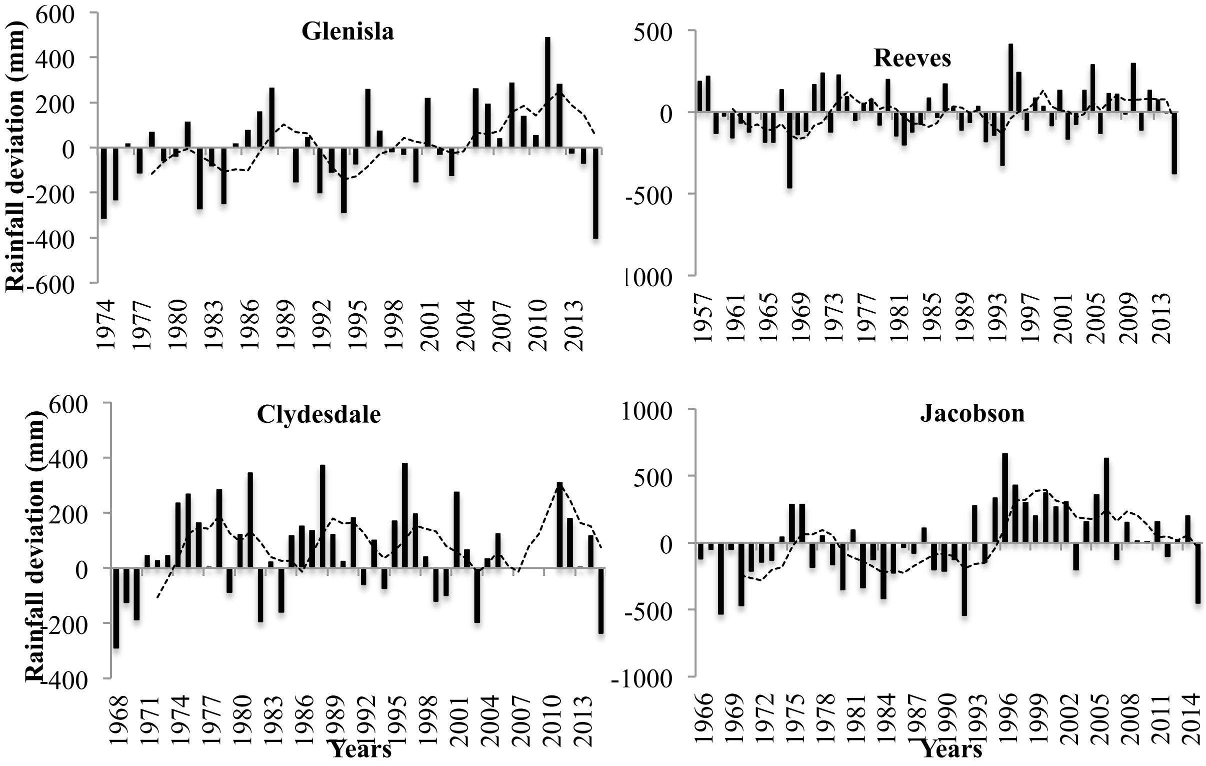
9. APPENDIX

Appendix A



Rainfall Station	r	p
Cavern	0.03	0.81
Royal Natal	0.27	**0.04**
Bergville	0.14	0.27
Cathedral Peak	0.26	**0.04**
Giant’s Castle	0.60	**<0.001**

Figure 9.1 Rainfall deviation from the mean at the five rainfall stations with 5-year moving average over the period 1955-2015.



Station	r	p
Reeves	0.11	0.43
Glenisla	0.32	**0.04**
Clydesdale	0.09	0.57
Jacobson	0.36	**0.01**

Figure 9.2 Rainfall deviation from the mean for four individual stations in the Sterkspruit Catchment with a five-year running average.

Appendix B

Methodology for the calculation of the SPI taken from Edwards (1997).

3.0 METHODOLOGY

3.1 SPI Defined

McKee *et al.* (1993) developed the Standardized Precipitation Index (SPI) for the purpose of defining and monitoring drought. Among others, the Colorado Climate Center, the Western Regional Climate Center, and the National Drought Mitigation Center use the SPI to monitor current states of drought in the United States. The nature of the SPI allows an analyst to determine the rarity of a drought or an anomalously wet event at a particular time scale for any location in the world that has a precipitation record.

Thom (1966) found the gamma distribution to fit climatological precipitation time series well. The gamma distribution is defined by its frequency or probability density function:

$$g(x) = \frac{1}{\beta^{\alpha}\Gamma(\alpha)} x^{\alpha-1} e^{-x/\beta} \qquad \text{for } x>0 \tag{3.1}$$

where:

$\alpha > 0$ — α is a shape parameter (3.2)

$\beta > 0$ — β is a scale parameter

(3.3)

$x > 0$ — x is the precipitation amount (3.4)

$\Gamma(\alpha) = \int_0^{\infty} y^{\alpha-1} e^{-y} dy$ — $\Gamma(\alpha)$ is the gamma function (3.5)

For example, figure 3.1 shows the gamma distribution with parameters α=2 and β=1. This distribution is skewed to the right with a lower bound of zero much like a precipitation frequency distribution.

Computation of the SPI involves fitting a gamma probability density function to a given frequency distribution of precipitation totals for a station. The alpha and beta parameters of the gamma probability density function are estimated for each station, for each time scale of interest (3 months, 12 months, 48 months, etc.), and for each month of the year. From Thom (1966), the maximum likelihood solutions are used to optimally estimate α and β:

$$\hat{\alpha} = \frac{1}{4A}\left(1 + \sqrt{1 + \frac{4A}{3}}\right) \tag{3.6}$$

$$\hat{\beta} = \frac{\overline{x}}{\hat{\alpha}} \tag{3.7}$$

where:

$$A = \ln(\overline{x}) - \frac{\sum \ln(x)}{n} \tag{3.8}$$

n = number of precipitation observations (3.9)

The resulting parameters are then used to find the cumulative probability of an observed precipitation event for the given month and time scale for the station in question. The cumulative probability is given by:

$$G(x) = \int_0^x g(x)dx = \frac{1}{\hat{\beta}^{\hat{\alpha}}\Gamma(\hat{\alpha})}\int_0^x x^{\hat{\alpha}-1}e^{-x/\hat{\beta}}dx \tag{3.10}$$

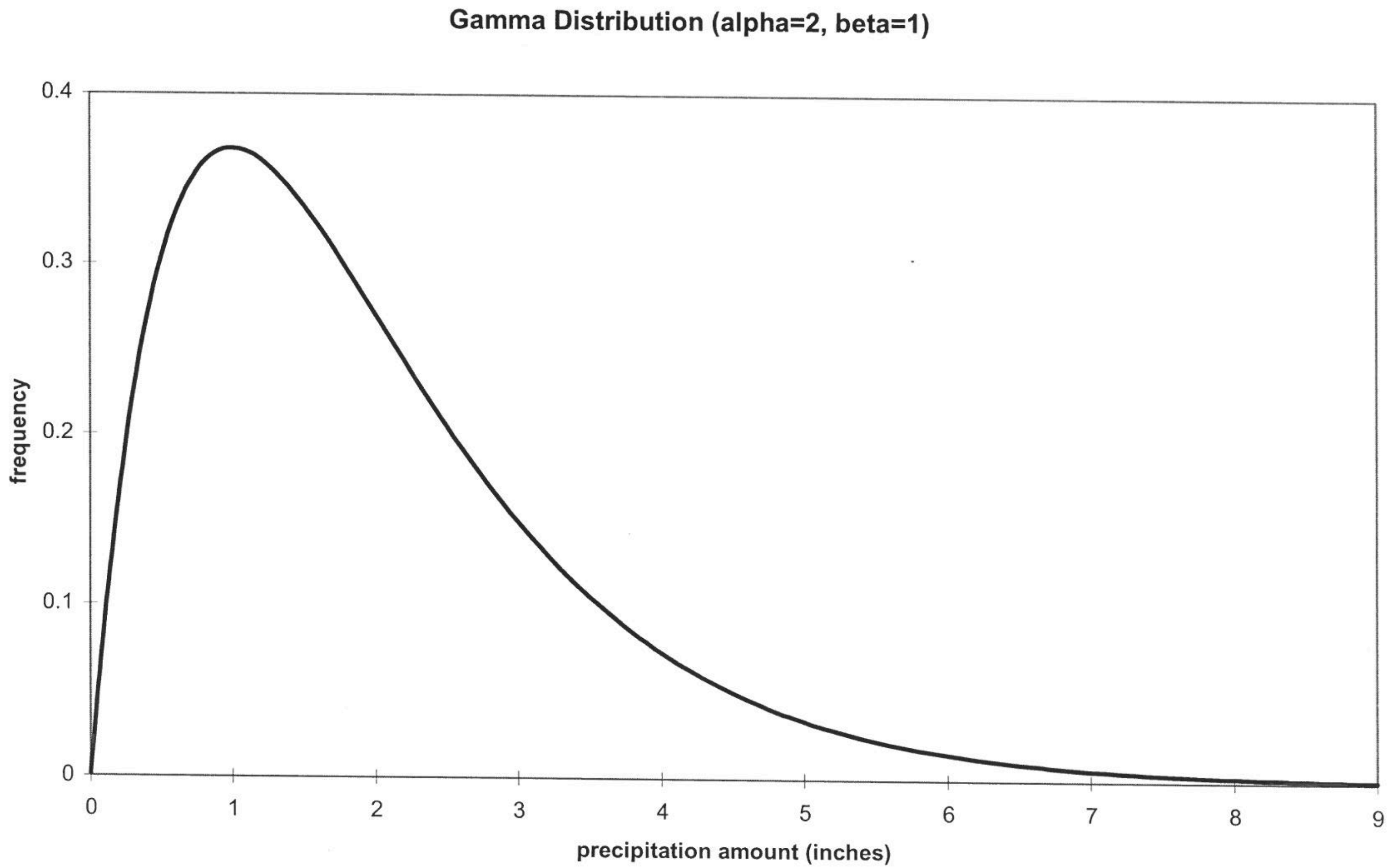


Fig. 3.1 Gamma frequency distribution with parameters alpha=2 and beta=1.

Letting $t = x / \hat{\beta}$, this equation becomes the incomplete gamma function:

$$G(x) = \frac{1}{\Gamma(\hat{\alpha})} \int_0^x t^{\hat{\alpha}-1} e^{-t} dt \tag{3.11}$$

Since the gamma function is undefined for x=0 and a precipitation distribution may contain zeros, the cumulative probability becomes:

$$H(x) = q + (1 - q)G(x) \tag{3.12}$$

where q is the probability of a zero. If m is the number of zeros in a precipitation time series, Thom (1966) states that q can be estimated by m/n. Thom (1966) uses tables of the incomplete gamma function to determine the cumulative probability G(x). McKee *et al.* (1993) use an analytic method along with suggested software code from Press *et al.* (1988) to determine the cumulative probability.

The cumulative probability, $H(x)$, is then transformed to the standard normal random variable Z with mean zero and variance of one, which is the value of the SPI. This is an equiprobability transformation which Panofsky and Brier (1958) state has the essential feature of transforming a variate from one distribution (*ie.* gamma) to a variate with a distribution of prescribed form (*ie.* standard normal) such that the probability of being less than a given value of the variate shall be the same as the probability of being less than the corresponding value of the transformed variate. This method is illustrated in figure 3.2. In this figure, a 3 month precipitation amount (January through March) is converted to a SPI value with mean of zero and variance of one. The left side of figure 3.2 contains a broken line with horizontal hash marks that designate actual values of 3 month precipitation amounts (x-axis) for Fort Collins, Colorado for the months of January

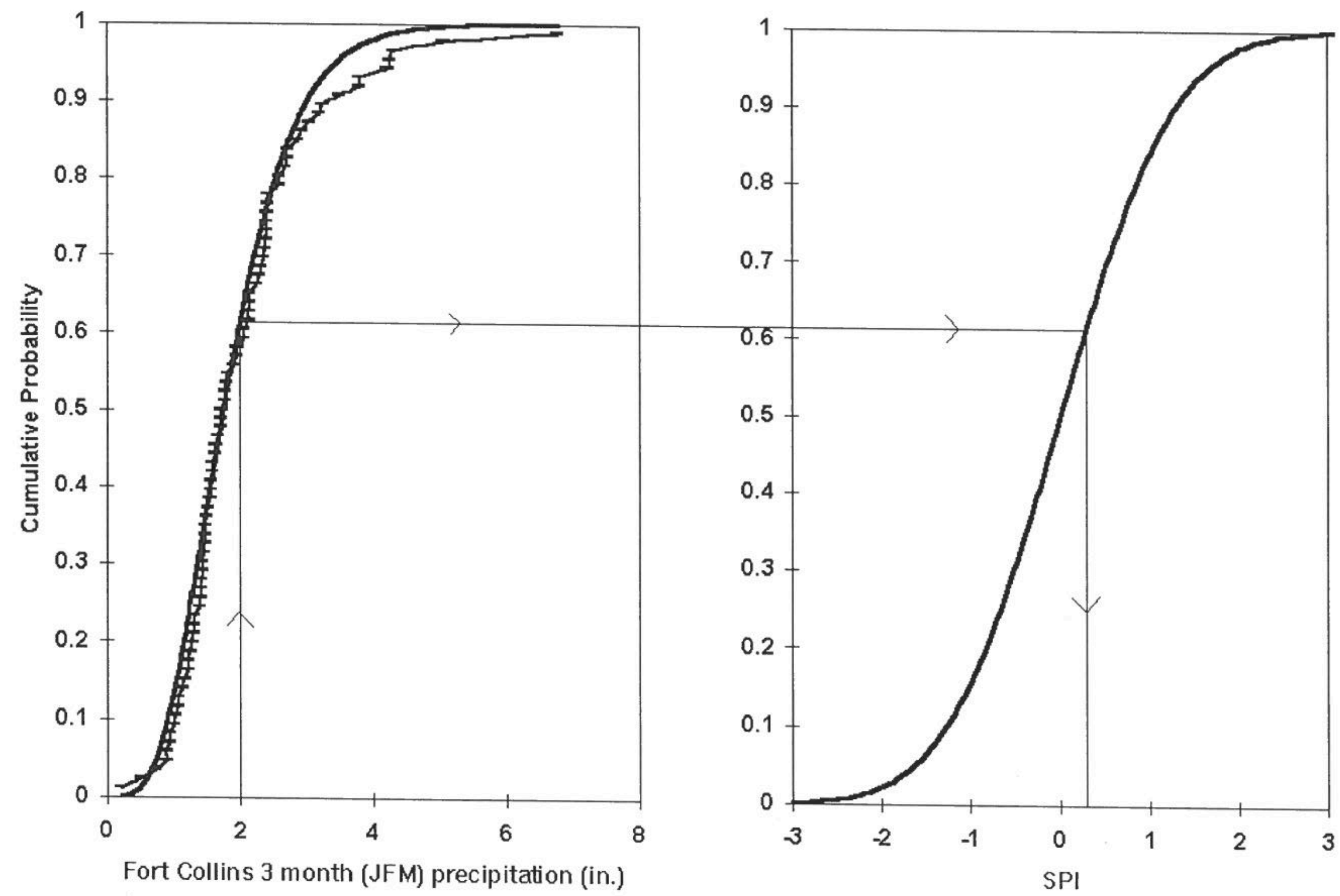


Fig. 3.2 Example of equiprobability transformation from fitted gamma distribution to the standard normal distribution.

through March for the period 1911 through 1995. The broken line also denotes the empirical cumulative probability distribution (y-axis) for the period of record. The empirical cumulative probabilities were found optimally as suggested by Panofsky and Brier (1958) where the precipitation data is sorted in increasing order of magnitude so that the kth value is k-1 values from the lowest and where n is the sample size:

$$\text{empirical cumulative probability} = \frac{k}{n+1} \tag{3.13}$$

The smooth curve on the left hand side of figure 3.2 denotes the cumulative probability distribution of the fitted gamma distribution to the precipitation data. The smooth curve on the right hand side of figure 3.2 denotes the cumulative probability distribution of the standard normal random variable Z using the same cumulative probability scale of the empirical distribution and the fitted gamma distribution on the left hand side of the figure. The standard normal variable Z (or the SPI value) is denoted on the x-axis on the right hand side of the figure. Hence, this figure can be used to transform a given 3 month (January through March) precipitation observation from Fort Collins, Colorado to a SPI value. For example, to find the SPI value for a 2 inch precipitation observation, simply go vertically upwards from the 2 inch mark on the x-axis on the left hand side of figure 3.2 until the fitted gamma cumulative probability distribution curve is intersected. Then go horizontally (maintaining equal cumulative probability) to the right until the curve of the standard normal cumulative probability distribution is intersected. Then proceed vertically downward to the x-axis on the right hand side of figure 3.2 in order to determine the SPI value. In this case, the SPI value is approximately +0.3.

Since it would be cumbersome to produce these types of figures for all stations at all time scales and for each month of the year, the Z or SPI value is more easily obtained computationally using an approximation provided by Abramowitz and Stegun (1965) that converts cumulative probability to the standard normal random variable Z:

$$Z = SPI = -\left(t - \frac{c_0 + c_1 t + c_2 t^2}{1 + d_1 t + d_2 t^2 + d_3 t^3} \right) \qquad \text{for } 0 < H(x) \leq 0.5 \qquad (3.14)$$

$$Z = SPI = +\left(t - \frac{c_0 + c_1 t + c_2 t^2}{1 + d_1 t + d_2 t^2 + d_3 t^3} \right) \qquad \text{for } 0.5 < H(x) < 1.0 \qquad (3.15)$$

where:

$$t = \sqrt{\ln\left(\frac{1}{(H(x))^2} \right)} \qquad \text{for } 0 < H(x) \leq 0.5 \qquad (3.16)$$

$$t = \sqrt{\ln\left(\frac{1}{(1.0 - H(x))^2} \right)} \qquad \text{for } 0.5 < H(x) < 1.0 \qquad (3.17)$$

$$\begin{aligned} c_0 &= 2.515517 \\ c_1 &= 0.802853 \\ c_2 &= 0.010328 \\ d_1 &= 1.432788 \\ d_2 &= 0.189269 \\ d_3 &= 0.001308 \end{aligned} \qquad (3.18)$$

Conceptually, the SPI represents a z-score, or the number of standard deviations above or below that an event is from the mean. However, this is not exactly true for short time scales since the original precipitation distribution is skewed. Nevertheless, figure 3.3 shows that during the base period for which the gamma parameters are estimated, the SPI will have a standard normal distribution with an expected value of zero and a variance of one. Katz and Glantz (1986) state that requiring an index to have a fixed expected value and variance is desirable in order to make comparisons of index values among different stations and regions meaningful.

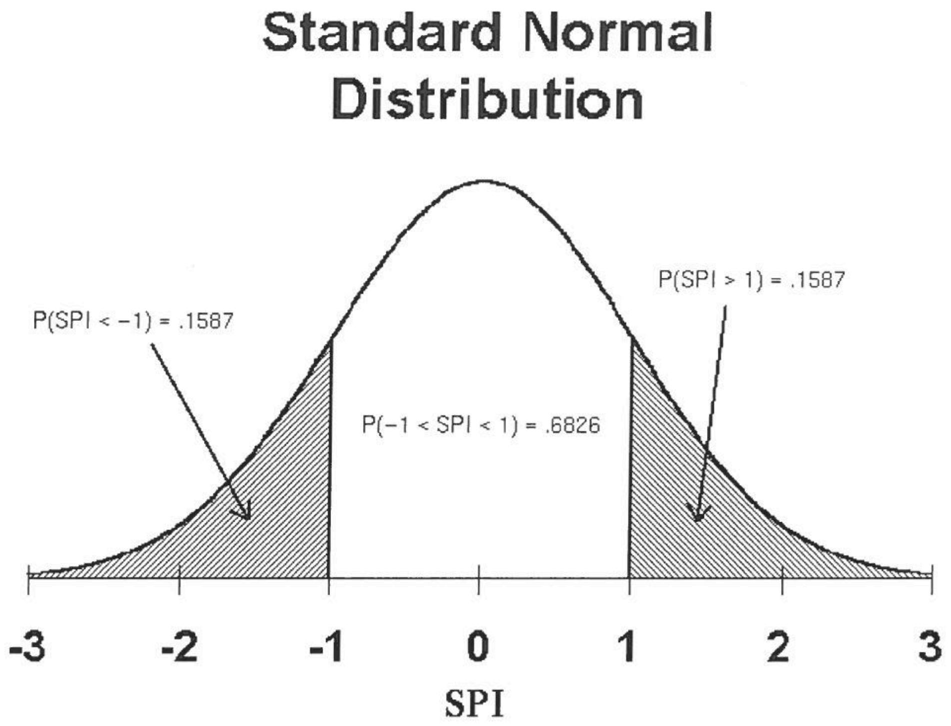


Fig. 3.3 Standard normal distribution with the SPI having a mean of zero and a variance of one.

Tannehill (1947) states that rainfall in the worst drought ever experienced in Ohio would be abundant rainfall in Utah. Akinremi *et al.* (1996) state that the spatial and temporal dimensions of drought create problems in generating a drought index because not only must an anomaly be normalized with respect to location, but the anomaly must also be normalized in time if it is to produce a meaningful estimate of drought. The SPI accomplishes both. The SPI is normalized to a station location because it accounts for the frequency distribution of precipitation as well as the accompanying variation at the station. Additionally, the SPI is normalized in time because it can be computed at any number of time scales, depending upon the impacts of interest to the analyst. Additionally, no matter the location or time scale, the SPI represents a cumulative probability in relation to the base period for which the gamma parameters were estimated. Table 3.1 is a table of SPI and its corresponding cumulative probability.

Table 3.1: SPI and Corresponding Cumulative Probability in Relation to the Base Period

SPI	Cumulative Probability
-3.0	0.0014
-2.5	0.0062
-2.0	0.0228
-1.5	0.0668
-1.0	0.1587
-0.5	0.3085
0.0	0.5000
+0.5	0.6915
+1.0	0.8413
+1.5	0.9332
+2.0	0.9772
+2.5	0.9938
+3.0	0.9986

An analyst with a time series of monthly precipitation data for a location can calculate the SPI for any month in the record for the previous *i* months where *i*=1, 2, 3, ..., 12, ..., 24, ..., 48, ... depending upon the time scale of interest. Hence, the SPI can be computed for an observation of a 3 month total of precipitation as well as a 48 month total of precipitation. For this study, a 3 month SPI is used for a short-term or seasonal drought index, a 12 month SPI is used for an intermediate-term drought index, and a 48 month SPI is used for a long-term drought index. Therefore, the SPI for a month/year in the period of record is dependent upon the time scale. For example, the 3 month SPI calculated for January, 1943 would have utilized the precipitation total of November, 1942 through January, 1943 in order to calculate the index. Likewise, the 12 month SPI for January, 1943 would have utilized the precipitation total for February, 1942 through January, 1943 while the 48 month SPI would have utilized the precipitation total for February, 1939 through January, 1943.