A WATER BALANCE APPROACH TO GROUNDWATER RECHARGE ESTIMATION IN MONTAGU AREA OF THE WESTERN KLEIN KAROO

Xianfeng Sun

A thesis submitted in fulfillment of the requirements for the degree of Master in the Faculty of Natural Sciences, Department of Earth Sciences, University of the Western Cape, Bellville, South Africa.

Supervisor: Prof. Yongxin Xu

Co-supervisor: Dr. Shafick Adams

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ABSTRACT

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Xianfeng Sun
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Department of Earth Sciences

Key words: Recharge estimation, Water Balance, TMG, Montagu, Modeling

The Western Klein Karoo-Montagu area is located in the mid-eastern of the Western Cape Province, South Africa. In most of the study areas within semi-arid climatic zone, groundwater plays an important role in meeting both agricultural and urban water requirements. Developments of agriculture depend on more and more groundwater supply from Table Mountain Group (TMG) sandstone aquifer system in the study area. Groundwater recharge is considered as one of the most important factors governing the sustainable yield of groundwater exploitation. There have been few studies on the recharge estimation of the TMG aquifer system in the Montagu area. Thus accurate and reliable recharge estimation of the TMG aquifer system in the Montagu area is important. The TMG aquifer in the Montagu area comprises approximate 4,000m thick sequence of sandstone with an outcrop area of 3,124 km^2, which is recharge area. The outcrops are characterized by mountainous topography with sparse to dense vegetation, shallow and intermittent diverse soils and mean annual rainfall of 350-450 mm/yr.

Based on detail analysis and interpretation of factors influencing recharge, water balance method is used to estimate recharge rates by using readily available data (rainfall, runoff, temperatures). Other estimate methods are difficult to be applied due to the limited information available in the study area. In this study, the water balance approach based on empirical evapotranspiration and runoff model is employed to determine and analyse long-term average water recharge. The long-term average recharge is modelled as a function of the regional interaction of the site conditions:

climate, soil, geology and topography. Modelling is performed according to the outlined procedure using long-term climatic and physical data from the different rainfall period of different gauge stations. As results, actual evapotranspiration, direct runoff and recharge have been quantified. The recharge ranges vary from 0.1 mm/yr to 38.0 mm/yr in the study area, and the values less than 20.0 mm/yr are predominant. Relatively low recharge rates coincide with low precipitation in most regions. Recharge is less than 5.0 mm/yr if mean annual precipitation (MAP) is less than 400 mm/yr. The ranges of 10.0-20.0 mm/yr of recharge occur in precipitation ranging from 600 mm/yr to 1,200 mm/yr. The recharge rates exceeding 20.0 mm/yr are more related to the precipitation with 800 mm/yr or more. The low recharge rates less than 2.0 mm/yr are related to single high rainfall event in the study area. The total recharge volume of the outcrop of the TMG in the study area is approximately $54.2 \times 10^6 m^3/yr$. Approximately 29.3% of the stream flow may be contributed by recharge in terms of baseflow.

The recharge in the study area increases with increasing precipitation, but recharge percentage is non-linear relationship with the precipitation. Separate high rainfall events mainly contribute recharge if annual precipitation is extremely low in the study area. Spatial distribution of recharge is associated with the variations in precipitation, geological and geomorphologic settings in the study area.

The method used yields a point estimate and then extrapolates to the whole study area. The ranges of recharge may be exaggerated or underestimated due to the finite number of the rainfall stations in the outcrop of the TMG of the study area. After comparison to other recharge estimates from early studies in the area, the estimates are considered as reasonable and reliable. The feasibility of the water balance approach in semi-arid area is confirmed as well. The estimates based on the water balance model should be crosschecked before they are applied for management of groundwater resources.

DECLARATION

I declare the **A WATER BALANCE APPROACH TO GROUNDWATER RECHARGE ESTIMATION IN MONTAGU AREA OF THE WESTERN KLEIN KAROO** is my own work, that it has been submitted for any degree or examination in any other university, and that all the sources I have used or quoted have been indicated and acknowledge by complete references.

Full name: Xianfeng Sun Date: 15th May 2005

Signed: ? ??????????????????????????????

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Abbreviations and Notations

Abbreviations

Abbreviations	Description
CAGE	Cape Artesian Groundwater Exploration
CFB	Cape Fold Belt
CN	Curve number
CNC	Cape Nature Conservation
CGS	Council for geosciences
CRD	Cumulate Rainfall Departure
DWAF	Department of Water Affairs and Forestry
EARTH	Extended model for Aquifer Recharge and Moisture Transport through Unsaturated Hardrock
GM	Groundwater Modelling
MAP	Mean Annual Precipitation
PRMS	Precipitation-Runoff Modelling System
SCS	Soil Conservation Service
SAWS	South Africa Weather Service
SHE	System Hydrology European
Stdev	Standard deviation
SVF	Saturated Volume Fluctuation
TMG	Table Mountain Group
USDA	US Department of Agriculture
WR90	Water Research Commission Flow data in 1990 (published in 1994)
WRC	Water Research Commission
WTF	Water Table Fluctuation
Sn	Nardouw Sub-group
Oc	Cedarberg Formation
Ope	Peninsular Formation

Notations

Notation	Description	Dimension or unit
F	latitude (radians, positive for north negative for south)	Degree
?	latent heat of vaporization	$MJkg^{-1}$
?	density of water	ML^{-3}
? ? (t)	change in volumetric water content in soils	L or L^3
A	recharge area	L^2
C_c	coverage of vegetation	Dimensionless
C_s	soil factor	Dimensionless

C_v	vegetation factor	Dimensionless
EC	electrical conductivity	mS/m
E (t)	evapotranspiration	L or L^3
E_t (t)	evapotranspiration	L or L^3
E_{tr}(t)	evapotranspiration	L or L^3
G_{SC}	the solar constant with a value of 118.1	$MJm^{-2}d^{-1}$
? h_i	water level change during month i	L
H	hydraulic head	L
I	inflow	L^3
i, j	principal coordinate directions	L
J	the number of days since January 1 of the current year	Dimensionless
K	hydraulic conductivity tensor	L/F
K_c (t)	vegetation coefficient	Dimensionless
L_f	lithological factor	Dimensionless
O	outflow	L or L^3
P_i	rainfall for month i	L/T
Pre	precipitation	L
P_t	threshold value representing aquifer boundary conditions	L
P (t)	precipitation	L
Q_a	abstraction during period	L^3/T
Q_{out}	natural outflow	L^3
Q_p	fluid sources or sinks per unit volume	l/T
Q (t)	runoff	L or L^3
Q_v	annual recharge volume	M^3
R_A	extraterrestrial radiation	$MJm^{-2}d^{-1}$
Re	recharge	L or L^3
RE	annual recharge rate	mm/yr
R_f	variable recharge rate	L or L^3
R (t)	recharge	L or L^3
S	aquifer storativity	Dimensionless
S_c	catchment area	[L^2]
S_f	slope factor	Dimensionless
S_s	specific storage	l/L
S_y	specific yield	l/L
T	time	T
T	average monthly temperature	°C
T_D	inference between average monthly maximum and minimum temperatures	°C
?t	time increment	T
?v	change in saturated volume of the aquifer	L^3
x	a space coordinate	L

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CHAPTER 1

INTRODUCTION

1.1 BACKGROUND

Study area is located in the latitude 33°15'S to 34°10'S and longitude 19°30'E to 22°00'E, including Montagu and Barrydale towns of the Western Cape Province, South Africa. Here is denominated as Montagu area. The border of the study area is referred to the hydrogeological unit 9 of the WRC project entitled: " Recharge processes of the Table Mountain Group aquifer system" (Wu, 2005).

Most of the study areas are located in the Western Klein Karoo, which is characterized by a flat topography. In south and west of the study area is Cape Ranges with an elevation of about 1,000 m above mean sea level where outcrops of the Table Mountain Group (TMG) present. The Montagu area forms tectonic basins including the Koo Valley, the Kaiser Valley and parts of the Hex River Valley filled with unconsolidated sandy deposits of Tertiary to Quaternary age and underlain mainly by Ordovician to Carboniferous sedimentary rocks of the Cape Supergroup. The area receives precipitation from less than 200.0 mm/yr in the Little Karoo to more than 1000.0 mm/yr on the mountainous areas. The most part of flat area receives limited rainfall, which is less than 400.0 mm/yr. The Montagu area, except for the Langeberg Mountain and the Hex River Mountain, is thus characterized by absence of surface water.

The most part of the relatively flat sand-covered area supports lots of plantations and agriculture is developed in this area. Groundwater from Bokkeveld Group aquifer and Table Mountain Group (TMG) aquifer plays an important role in meeting both agricultural and urban water requirements in this area. There are approximately 15.46 millions of cubic meters of groundwater used in this area per year (National Groundwater Base, South Africa) and groundwater demands will increase with the development of agriculture. Continued development of agriculture depends on sustainable groundwater supply from the aquifer systems, especially in the TMG aquifer system in a certain degree due to the lack of surface water in this area. Early research works carried out to date indicate that the TMG aquifer system is a regional aquifer with potential to be a major source for future water supply in the study area. The TMG aquifer consists mainly of sandstone, quartzite and a little shale and is exploited extensively for agricultural purposes. These figures highlight importance of not only the necessity to properly explore the available resource but also to critically assess their sustainability over prolonged period of time.

A major challenge in the exploitation of groundwater in the Montagu area is that harmony of water resources and development of agriculture. This is dependent on their proper assessment and management of the groundwater resources. Groundwater recharge is considered as one of the most important factors governing the sustainable yield of groundwater exploitation in this area; however, there have been few studies on the recharge of the TMG aquifer system. Thus accurate and reliable recharge estimation of the TMG system in the Montagu area needs to be done.

The study focuses on estimating recharge rates of the TMG aquifer system in the Montagu area. Recharge is influenced by climatic and hydrogeological factors, which change with space and time. Understanding the factors and processes are important to determine the variability in quantity of groundwater recharge. An analysis of relationship between climatic, hydrogeological and geomorphologic conditions and recharge is the basis of construction of model for recharge estimation. It will improve reliability of estimates in the study area.

1.2 OBJECTIVES

Recharge estimation of the TMG aquifer in semi-arid regions is one of the key factors necessary for effective and rational management of groundwater resources. For the

TMG aquifer, the recharge area is the outcrop of the TMG in the study area. Therefore, this study focuses on the recharge estimation in the outcrop area. The aim of this study is to improve understanding of functioning of the TMG aquifer system and contribute to the sustainable development of this potential source for water supply in the Montagu area. Therefore, specific objectives in this study focus on:

1. Collect and collate information related to the study area and topic ; analyze the factors impact on recharge;
2. Interpret spatial and temporal variability of rainfall patterns ;
3. Construct water balance model of the area based on existing and new geological and climatic information;
4. Estimate recharge , including evaluation of the spatial and temporal variability.

1.3 RESEARCH FRAMEWORK

To quantify the spatial and temporal variability of the recharge, recharge estimation method need to be selected in order to achieve the objectives using limited data sets available in the study area. In attempt to meet the primary and the specific objectives mentioned above, an outline of research process is illustrated in Figure 1.1. The data sets of climate, geomorphology, soil and vegetation are necessary because they influence synthetically the processes of recharge. This approach employs an empirical water balance model, in which the spatial and temporal variability of physical parameters are considered. Spatial distributions of recharge are determined by integrating long-term average recharge rates of each rainfall stations.

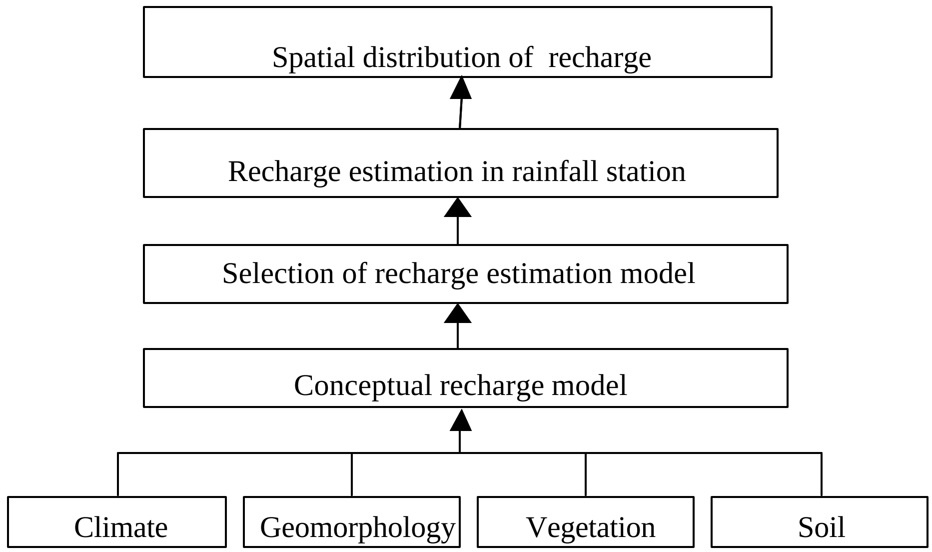


Figure 1.1 Flow diagram for study of recharge estimation in the Montagu area

1.4 REPORT LAYOUT

The thesis comprises seven chapters, and the various chapters are set-up as follows:

- Chapter one outlines the study background and motivation for undertaking the study as well as the aims and objectives of the study. An introduction of the study is addressed.

- Chapter two discusses the applied methods and theoretical aspects of groundwater recharge estimation. An overview of water balance methods commonly used for groundwater recharge estimation is discussed; particular emphasis is laid on the use of models in semi-arid study area. The previous studies and research undertaken, and its relevance to this study are detailed in this chapter.

- Chapter three gives the outline of the study area in terms of the climate, geomorphology, vegetation, soils, geology and hydrogeology. The topographic and vegetation features are figured out through field photos. The climatic patterns are discussed with precipitation, temperature and potential evapotranspiration. The geological characteristics are described through lithology, statigraphy and structure. The hydrogeological setting are generalized by aquifer property, flow pattern and hot springs.

- Chapter four outlines the methodology of the study. Principle of water balance and the empirical model proposed in the study area are discussed The data and error analysis, calculation method and model calibration are discussed in detail.

- Chapter five presents the influence factors of recharge, including the precipitation, topography, geology, soils and vegetation. Special aim focuses on the determination of spatial and temporal variability and tendency of rainfall in the Montagu area in order to facilitate the evaluation of the recharge in response to change of rainfall. The impacts of precipitation characteristics such as precipitation type, intensity, duration and distribution on actual evapotranspiration and direct runoff are discussed in detail.

- Chapter six presents the results of the study. Recharge is considered as the residual between precipitation and direct runoff and actual evapotranspirtation at year-round scale in this study. The estimates from the water balance method are evaluated. The recharge volume is calculated for the management of the TMG aquifer in the study area. The relationship between groundwater recharge and precipitation is discussed. The value of recharge obtained from early studies within the Montagu area and neighboring area in the Klein Karoo are presented and compared with the results of this study. The recharge in the study area is compared with the separation of stream flow, in terms of baseflow.

- Finally, a brief summary of the main findings from this research programme is given and some suggestions for future research work in the semi-arid environment of the Montagu area are presented.

All scientific papers prepared during this study, which provide additional information, are listed as references. The results of rainfall pattern, recharge rate, runoff and evapotranspiration calculated of selected rainfall stations are listed as appendix to this thesis.

CHAPTER 2
LITERATURE REVIEW

2.1 INTRODUCTION

In order to understanding current research into recharge estimation, it is necessary to do a review of available literatures. It will contribute to an improved conceptual understanding of the TMG aquifer system and ensure that present research work in context is being applied in terms of the methods proposed. The common approaches of recharge estimate, especially water balance methods are detailed. The literatures about previous work within the Montagu area and recharge estimation in nearly similar conditions (i.e. arid to semi-arid climate and similar geological conditions) in South Africa are surveyed in this chapter. The literatures studies contribute to the formulation of the recharge calculate outlined in relative chapter.

2.2 PREVIOUS WORKS

No previously detailed studies of recharge estimation of the TMG aquifer system have been done or published for the Montagu area. The Department of Water Affairs and Forestry (DWAF) did several hydrological surveys in the vicinity Hex River valley (Rosewarne, 1978, 1981a, 1984, 1997, 2002). Water resource investigations in the vicinity of Verlorenvalley (UMVOTO, 1999a, b, c, d; Weaver, 1995) and hydrogeological investigations in the Little Karoo provided information for the mountainous areas and Western Klein Karoo in the immediate vicinity of the study area.

The Council for geosciences (CGS) mapped the structures (dykes, major joints and fractures) in the study area. The CGS has interpreted the geology of the Worcester and Ladismith area, which form the parts of the study area on 1:250 000 (Gresse, 1992; Theron, 1991) scale geological maps. The Department of Water Affairs and Forestry (DWAF) mapped the hydrogeological character of parts of the study area on 1:500 000 scale General Hydrogeological Map (Meyer, 2001).

2.3 THE GENERAL APPROACHES OF RECHARGE ESTIMATION

Groundwater recharge as used in this study is defined as the downward flow of water reaching the water table (or piezometric surface), forming an addition to the groundwater reservoir (Lerner et al, 1990). There are as many methods available for quantifying groundwater recharge, as there are different sources and processes of recharge. They are (a) direct versus indirect, (b) water balance (c) Darcyan physical methods, (d) chemical, isotopic and gaseous tracer methods (Lerner et al., 1990; Kinzelbach et al., 2002). At present, a geographic information system approach is used to estimate groundwater recharge. Each of the methods has its own limitations in terms of applicability and reliability. Recharge is very difficult to estimate reliably, and more than one method should be used. Development of a conceptual model of recharge in the study area should also precede selection of the appropriate recharge estimation method in order to reduce both uncertainty as well as costs of quantifying recharge. Such a model should describe the location, timing and probable mechanisms of recharge and provide initial estimates of recharge rates based on climatic, topographic, land use and land cover, soil and vegetation types, geomorphologic and (hydro-) geologic data (including recharge sources, flow mechanisms, piezometry, groundwater exploitation, etc.). However, a user-friendly framework for recharge estimation does not yet exist (Xu et al. 2003). The choice of the recharge estimation methods will depend on the conceptualization of the recharge processes and the accuracy required in a given situation.

However, it is probably easier to assess recharge in humid areas rather than in arid or semi-arid areas. This is a consequence of the time variability of precipitation in arid and semi-arid climates, and spatial variability in soil characteristics, topography, vegetation and land use (Lerner et al., 1990). Recharge cannot be easily measured directly, especially, in hard rock regions. The methods used commonly to estimate recharge will be briefly described. The recharge estimation carried out in this study focuses mainly on water balance method and is therefore discussed in detail.

Direct techniques are very useful for measuring drainage through soils and hence can give an indication of groundwater recharge. However, the use of lysimeters is restricted due to they have relatively small surface areas and are shallow, so the drainage measured may not be fully representative of the flows that would reach a relatively deep water table below a thick subsoil. This method is expensive and seldom practicable for a point measurement.

The indirect techniques include soil moisture budget method, zero-flux plane method and estimation of water fluxes. All the methods require that measurements be made of soil moisture physical parameters, which are then used as input values for moisture flux estimation. The applications of the methods are therefore particularly limited in semi-arid area, problems are (i) low moisture fluxes, changes in parameter values may be small and as such often difficult to detect; (ii) it will be made for several years to obtain estimates of mean values because of high temporal variability of parameter values; (iii) large number of sampling points will be required to study the variability in recharge because of spatial variability of local topography.

The Water balance methods are commonly used to estimate recharge because of its relative simplicity. The advantages of water balance methods are that they can usually be estimated from readily available data (rainfall, runoff, water levels) and rapid to apply, and they account for all water entering and leaving the system (Lerner et al, 1990). The major disadvantage is that recharge is the residual that is a small difference between large numbers. The other disadvantages include the difficulty of estimating other fluxes. For example, evapotranspiration cannot be measured, yet it is often the largest outward flux.

The Darcian physical methods are combine methods of Darcy's law and equation of mass conservation. The principal advantage of these methods is that they attempt to represent the flow of water – the actual physical processes that would be interested in. The methods usually assume steady conditions, when only measurements of head and hydraulic conductivity are needed to apply the Darcy equation or its unsaturated equivalent. The methods work well for saturated flow, provided conductivities can be measured at the right scale. Unsaturated flow is much more difficult to estimate from field measurements because of the sensitivity of conductivity values to moisture content. This is a major problem in the semi-arid conditions.

The chemical, isotopic and gaseous tracer methods are widely used for recharge estimation in semi-arid areas using both environmental and applied tracers. Lerner et al. (1990) separate the methods into signature methods and throughput methods. Applied tracers (such as tritium, carbon-14) are normally only used in the signature methods (where a parcel of water containing the tracer is tracked and dated). Throughput methods involve a mass balance of tracer, comparing the concentration in precipitation with the concentration in soil water below the water table. Piston flow is generally assumed in most tracer studies. Chloride is probably the most widely used environmental tracer for the throughput method (Hendrickx and Walker, 1997; Wood, 1999). It is particularly effective in arid zones where there is significant concentration through evaporation. The method is based on the assumption that in an unconfined aquifer, the hydrochemically stable major ion, chloride, is derived solely from precipitation and is concentrated only by evapotranspiration before it reaches the water table. Because tracers do not measure water flow directly, a number of problems can arise, leading to over- or under-estimates of recharge. The principal problems are unknown tracer inputs, mixing and dual flow mechanisms.

A miscellany of other methods is in use for recharge estimation. These are mainly empirical and numerical models. In empirical model, recharge is correlated with other variables such as precipitation, elevation and canal flow. The relationship is then used (a) to extend the recharge record in time, or (b) is to transpose to other catchments of similar characteristics. It is clear that cachments should be closely matched in characteristics, and that the empirical relationship is only as good as the recharge estimates on which it was based. Changing ground water conditions (once resources are exploited) may change recharge, but empirical methods cannot estimate these changes, as they contain no model of recharge processes (Connelly et al, 1989). Numerical models including conventional model and direct model are used in recharge estimation. In numerical model, the hydrogeological model needs to well understand.

2.4 WATER BALANCE METHODS

The water balance methods are often used to estimate hydrologic fluxes on different scales. The water balance methods estimate recharge as the residual of all the other fluxes such as precipitation, runoff, evapotranspirtation, and change in storage. The principle is that other fluxes can be measured or estimated more easily than recharge. The hydrological water balance for an area can be defined as:

$$R(t) = P(t) - E(t) - Q(t) - \Delta q(t) \qquad (2.1)$$

where R(t) is recharge, P(t) is precipitation, E(t) is evapotranspiration, Q(t) is runoff, and ??(t) is change in volumetric water content. The term (t) designates that the terms are distributed through time.

The water balance methods include soil moisture balance, river channel water balance and water table rise method. The methods are commonly used because its simplicity, which requires minimal input data: precipitation, evapotranspiration, runoff and soil-water holding capacity.

The development of hydrologic models started in the 1960s with the Stanford watershed model (Crawford and Linsley, 1966). Until today the number of models and model systems as well as the number of different model concepts grew considerably.

Amongst the current generation of hydrological models, which can be classified as physically-based and conceptual models. Freeze (1978) introduced the first generation of distributed, physically-based models founded on rigorous numerical solution of partial differential equations governing flow through porous media, overland flow and channel flow. Lumped, conceptual models have been part of the hydrological literature, for an even longer period, and are presented as an alternative to physically-based models. They do not take into account the detailed geometry of catchments and the small-scale variabilities; rather they consider the catchment as an ensemble of interconnected conceptual storages. In particular, they do not explicitly include any laws of physics, which purportedly underlie physically-based models.

Most of the models have been developed for a specific scale and the simulation of a specific aspect of the hydrologic cycle. Physically based models like PRMS (Leavesly et al 1983), or SHE (Abbot et al., 1986), for instance, have been developed for the application in micro- to mesoscale watersheds. The application of these models in macro scale catchments areas is limited not only due to the lack of input data needed to run these models but also because of rationalization issues (e.g. Blöschl and Krikby, 1996). The problem to apply small-scale models in large catchment areas has lead to the development of models especially designed for macroscale applications. These models differ significantly to micro- and mesoscale models with respect to the representation of the relevant processes and the spatial and temporal resolution. The RHINEFLOW model (Kwadijk, 1993), for instance, calculates the water balance for the Rhine basin using a more integrated approach on a monthly basis. The HBV-model (Bergström S., 1995) is a more deterministic approach using daily resolution, applicable to larger areas. Klemes (1983) suggested two diametrically opposite approaches towards the development of theories and models of hydrological response at the catchment scale. He presents an example of the prediction of monthly runoff in a 39,000 km^2 catchment in Canada. Jothityangkoon et al (2001) presented a systematic approach to the development of a long-term water balance model for a large catchment in semi-arid Western Australia. The process controls on water balance are examined at the annual, monthly and daily scales. A systematic 'downward' approach for the formulation of models of appropriate complexity is presented based on an investigation of the climate, soil and vegetation controls on water balance.

For modeling the long-term availability of water resources in catchment area required. Empirical models like the proposed model turned out to be sufficient. These models allow a reasonable determination of the long-term water balance as a function of the interplay between the actual land cover and climatical, pedological, topographical and hydrogeological conditions. Empirical approaches are very cost effective because of the reduced temporal resolution (= 1 year) of the required climatic input data and a relatively small calibration effort. Therefore, empirical approaches are often used in practical water resources management issues in large regions or river basins. Using the empirical model of GROWA98, long-term an area differentiated water balance analysis in the river Elbe basin (German part) covering an area of about 100,000 km^2 is carried out by Ralf (2002).

Most of the well-known models available were developed for humid climates. However, despite the various limitations, it is important to briefly summaries some of the models which may be widely used in semi-arid conditions. These models are based on equation 2.1; water level, borehole abstractions and aquifer properties are considered as well.

The Saturated Volume Fluctuation (SVF) method is based on water balance over time based on averaged groundwater levels from monitoring boreholes. Recharge is calculated as (Bredenkamp et al., 1995):

$$S.\frac{\Delta V}{\Delta t} = I - O + RE - Q_a \tag{2.2}$$

where S is aquifer storativity, $?v$ is change in saturated volume of the aquifer, $?t$ is the time increment over which the water balance is calculated, I is inflow, O is outflow, RE is recharge, Q_a is abstraction during period.

The water level, borehole abstractions and aquifer properties including storativity and size of aquifer area are required in this method. The major advantage of SVF-type estimations is that they allow recharge estimations to be made from current data (Xu et al., 2003). The shortcoming of the method is that the measured water levels must be representative for the aquifer as a whole and inflow value is often assumed equal to outflow value.

The Cumulative Rainfall Departure (CRD) is based on the premise that water level fluctuations are caused by rainfall events. The method was applied extensively with success in South Africa. Recharge is calculated as (Xu and Van Tonder, 2001):

$$\begin{aligned} &R_T = rCRD_i = S_y\left(\Delta h_i + \left(Q_{pi} + Q_{outi}\right)/\left(AS_y\right)\right) \\ &with \\ &CRD_i = \sum_{i=1}^{N} P_i - \left(2 - \frac{1}{P_{av} i}\sum_{i=1}^{N} P_i\right) iP_t \end{aligned} \tag{2.3}$$

where r is that fraction of a CRD which contributes to recharge, S_y is specific yield, $?h_i$ is water level change during month i (L), Q_p is groundwater abstraction (L^3/T), Q_{out} is natural outflow, A is recharge area (L^2), P_i is rainfall for month i (L/T) and P_t is a threshold value representing aquifer boundary conditions. P_t may range from 0to P_{av}, with 0 representing a closed aquifer (no outflow), and P_{av} representing an open aquifer system (for instance controlled by spring flow).

The limitation deep (multi-layer) aquifer and sensitivity of specific yield do not be considered with this method. Another shortcoming of CRD methods is that the impact of abstraction from boreholes on water level is ignored The data requirements of this method include monthly rainfall records, water level, borehole abstractions and aquifer properties including storativity and size of recharge area. Water level fluctuations are caused by corresponding rainfall events should be known first.

The Groundwater Modelling (GM) is a method, which recharge inversely derived from numerical modelling general three-dimensional groundwater flow equation assuming uniform fluid density and viscosity is formulated as (Bear, 1972):

$$S_s \frac{\partial h}{\partial t} = \frac{\partial}{\partial x_i}\left(K_{i\,j}\right) + q_s \tag{2.4}$$

where ?*j* represent principal coordinate directions, *K* is hydraulic conductivity tensor (L/F), *h* is hydraulic head (L), S_s is specific storage (l/L), *x* is a space coordinate (L), *t* is time (T) and q_s represents fluid sources (such as recharge) or sinks (such as abstraction) per unit volume (l/T).

This method needs conceptual hydrogeological model, daily/monthly rainfall records, water levels, borehole abstractions, aquifer characteristics such as storativity, hydraulic conductivity, porosity etc. it is time cons uming, sensitive to boundary conditions and difficult to calibrate.

The EARTH model is another water balance type method. Lumped, distributed model simulates water level fluctuations by coupling climatic, soil moisture and groundwater level data for estimating recharge (Van der Lee and Gehrels, 1997; Gieske, 1992). The model needs a lot of parameters and storativity is difficult to know.

A shortcoming with most models is that uniform recharge over the model area is assumed, which is quite clearly not the case in most instances (Xu et al, 2003).

Otherwise, large number of hydrogeological data including recharge source, flow mechanisms, aquifer characteristics such as storativity, hydraulic conductivity, porosity, etc are required for most model.

The methods used, and the accuracy of their estimates of recharge, are highly controlled by the scale of investigation. The accuracy of water balance methods is largely dependent on the quantity and quality of data available for interpretation (e.g. spreading of the boreholes over the aquifer, frequency of water level and abstraction data, correctness of the conceptual model and boundary conditions) (Xu et al, 2003). The model normally requires a wealth of detailed climatic, physiographic data, which are not available for the entire study region, thus this results should have many uncertainties.

2.5 RECHARGE ESTIMATION IN THE TMG AND STUDY AREA

As early as 1970, Joubert (1970) suggested infiltration into the TMG could be as high as 60% of Mean Annual Precipitation (MAP), but this order of estimation was more an exception than the rule. The study of recharge in the TMG aquifer system is gaining increasing in recent 10 years. Workers attempted to estimate recharge using a variety of methods. However, comprehensive study of recharge of the TMG aquifer is absent although there are lots of the results from a series of recent case studies in which attempts were made to quantify recharge of the TMG aquifer. The commonly used methods in the TMG area are physical and tracer approaches as follows:

- Chloride Mass Balance (CMB) method
- Cumulative Rainfall Departure (CRD) method
- Saturated Volume Fluctuation (SVF) method
- Base flow method
- C-14 and ^{2}H method
- EARTH method
- GIS method
- Empirical method
- Spring Flow method
- Water Balance method

Recharge studies on the TMG aquifer were preformed in Vermaaks River wellfield due to the good data sets of monitoring of groundwater levels, abstraction, springflow and rainfall (Brdeenkamp, 1995; Kotze, 2000; Weaver et al, 2002). Otherwise, a number of hydrogeological studies of the TMG have been undertaken in Agter Witzenberg (Weaver, John and Talma 1999, 2000), CAGE (Hartnady and Hay, 2000; Er Hay and CJH Harnady, 2002), Hermanus (Rosewarne and Kotze, 1996) and Uitenhage Artesian Basin (Maclear, 1996; Kok, 1992; Parson, 2000; Xu and Maclear, 2003) with the different method.

Recharge estimates of the TMG aquifer in the study areas gained using spatial model (Fortuin, 2004) and soil water balance method (Jia and Xu, 2005). Both of the studies made use of GIS technology. Another recharge estimates were presented in Vegter (1995) and Visser (2005), but no method was documented. The average estimates are 25.4, 26.0, 39.9 and 92.2 mm/yr related to Visser (2005), Vegter (1995), Fortuin (2004) and Jia and Xu (2005), respectively. These estimates should be examples of comparison in this study.

2.6 SUMMARY

A wealth of recharge estimation methods for (semi-) arid areas is currently available with each method having its own limitations in different size and time scales. Clarity on the aim of the recharge study is crucial in choosing appropriate methods for recharge estimation, which must be according to the data input available. In comparison to the various techniques, including the range, space/time scales, and reliability of recharge estimates, the water balance method based on evapotranspiration and runoff models can be applied as data required for application of the method are readily available in most cases. Of course, accuracy of recharge estimation should be developed, and it poses an iterative process that includes refinement of estimates as additional data are gathered and other methods are used.

CHAPTER 3
LOCALITY DESCRIPTION

3.1 INTRODUCTION

The physical characteristics of environment play an important role in recharge of groundwater. Understanding the characteristics of the fractured-rock TMG aquifer, and climate and geomorphology in local area would improve the understanding of the recharge processes and the accuracy of groundwater recharge estimation. In this chapter the climate, geomorphology, geology and hydrogeology of the Montagu area, which may influence the recharge, are discussed.

3.2 LOCATION AND EXTENT OF THE STUDY AREA

The Montagu area is located in the mid-eastern part of the Western Cape Province, ranging from latitude 33°15'S to 34°10'S and longitude 19°30'E to 22°00'E (Figure 3.1). It is a west-east trending area including the Koo Valley, the Keisie Valley and part of the Hex River Valley. The study area forms part of the Western Klein Karoo region with an arid to semi-arid climate. A linear outcrop of the TMG is presented along the Langeberg chain, the Hex River Mountain and around the Warmwaterberg.

3.3 TOPOGRAPHIC CHARACTERISTICS

The study area is located between the Little Karoo high plain in the north and the Cape Ranges in the south and west. To the north of the study area the Little Karoo high plain has a flat topography at an elevation of about 400 m a.m.s.l. The Langeberg Mountain chain is in the south, which rises 300 m a.m.s.l in the east to 1,200 m a.m.s.l in the west with the highest point at 1,300 m a.m.s.l. To the west is the Hex River Mountain; it trends N-S with a maximum altitude of 2,200 m a.m.s.l. The Gourits River forms the outlet of the catchment in the east. The bottom of the valley is fairly hilly and the altitude varies from 300 m to 600 m a.m.s.l. It is a narrow unit with gently rising mountains to the south.

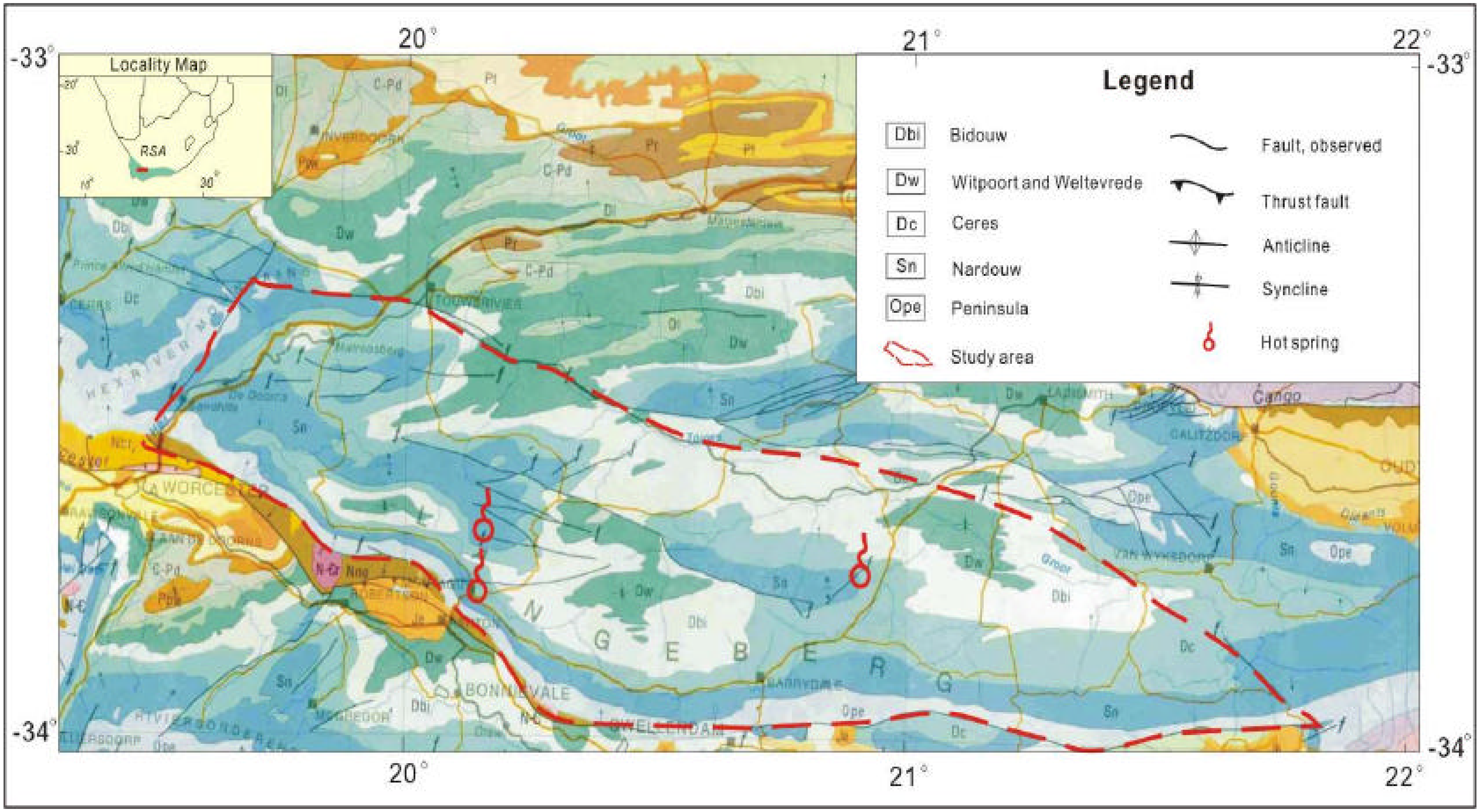


Figure 3.1 Geological map of the Montagu area (Adopted from Geological Map with 1:1million, CGS, 1997

Slopes are generally about 15°-30° and increases with altitude. There are steeper slopes on both sides of gorge. At the top of the mountains, the ridges are narrow and are about 100 to 500 meters wide.

The topography of the TMG outcrop is typical for the mountainous areas. The major topographic types include mountains, hills, valleys and gorges as shown in Figure 3. 2 - Figure 3. 6.

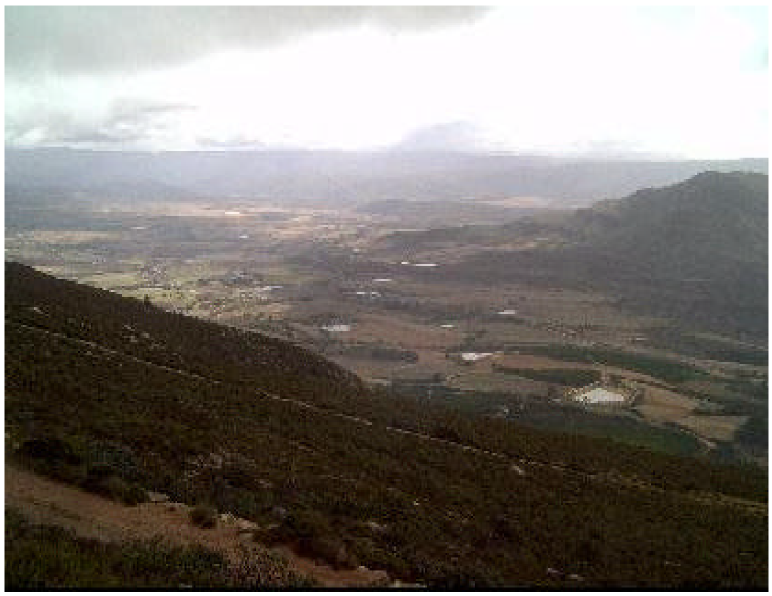


Figure 3.2 The Koo valley view from Protea Farm

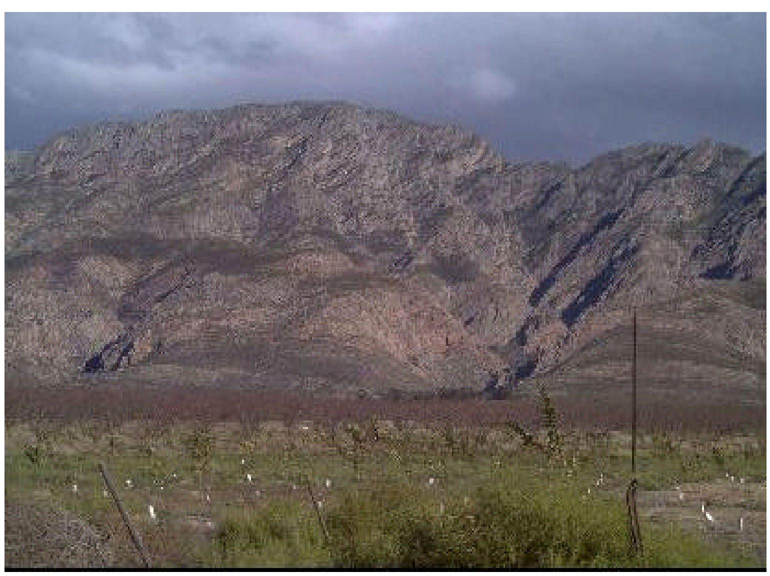


Figure 3.3 Monocline of the Longeberg in Montagu South

Figure 3.4 Monocline in the Hex River Pass

Figure 3.5 Folded mountain (anticline) in Montagu (Old English Fort)

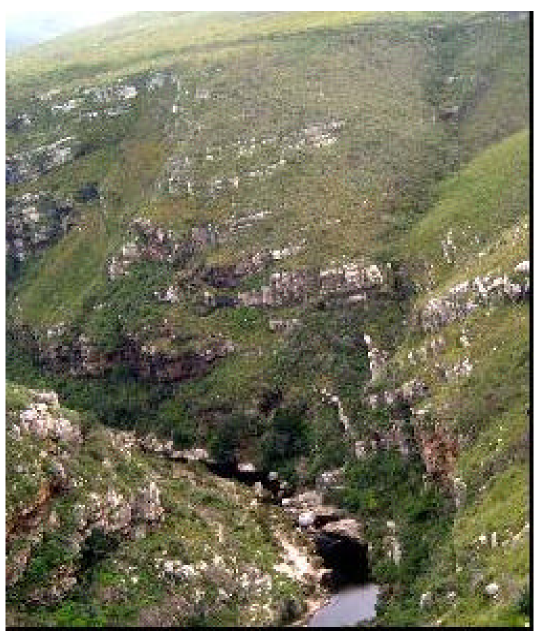


Figure 3.6 Gorge in the Tradouw Pass

3.4 CLIMATE

The climate of the study area is determined by altitude, topography and distance from the sea. There are three climatic types in the study area, namely a semi-arid climatic zone, Mediterranean climatic zone and temperate maritime climatic zone. The semi-arid climatic zone ranges from Worcester in the west via Montagu toward Attakwasberg along the Little Karoo border. The Mediterranean climatic zone is in the extreme west along the Hex River Mountain of the study area. It is a climate of mild, rainy winters and hot, dry summers. The temperate maritime climatic zone is in the east from Barrydale via Brandvlei to Herbertsdale. It has moderately warm summers and mild winters and the rain falls all year. There are no distinct boundaries that separate these three areas.

The general climatic information is available in low elevation areas of the outcrop of the TMG in the study area. The climatological data for these areas (i.e. precipitation, evapotranapiration and temperature) were collected from the South African Weather Bureau and WR90 (Midgley et al, 1994). The climatic characteristics in high elevation areas are not discussed in detail due to the lack of information.

3.4.1 Precipitation

The amount of precipitation and rainy season in the study area varies greatly in different climatic zones. The mountainous topography also affects local weather features, resulting in higher rainfall and even snow in winter at the higher altitudes. The variation of precipitation is dramatic because of the influence of topography. Drier periods are from May to July and December to January. Berg wind conditions prevail from May to July. The typical examples can be found as follow areas.

- Grootvadersbosch: the Langeberg lies in the transitional zone between winter and all-year rainfall regions. Grootvadersbosch has an average annual rainfall of about 1050 mm, but on the foot of the mountains, the rainfall changes into much lower.
- Majestic Swellendam mountains: the climate in the Swellendam area is typical of the Southern Cape, with hot summers and cold winters. The rainfall is fairly

evenly spread throughout the year with June and July the driest months and March, October and November the wettest.

3.4.2 Temperature

Temperature changes with variations in elevation, terrain, and ocean currents. There are very little differences in average temperatures from south to north in the study area but large monthly variations between maximum and minimum temperatures, as well as daily and seasonal temperatures for the region. The average temperature ranges from 10ºC to 20ºC as shown in Figure 3.7. The hot months occur in December, January and February. Temperature in June and July are the lowest. Temperature during the hot summer months (October to February) can be in excess of 40ºC, while the cold winter months (June to August), temperatures sometimes fall below 0ºC on the northern sides of mountains during the night. The frost and snow occur in July. Monthly average maximum temperatures vary from 40ºC near the Klein Karoo in January to 19ºC in the mountainous areas in July, whereas the monthly average minimum temperatures range from 7.5ºC in January to -4ºC in July.

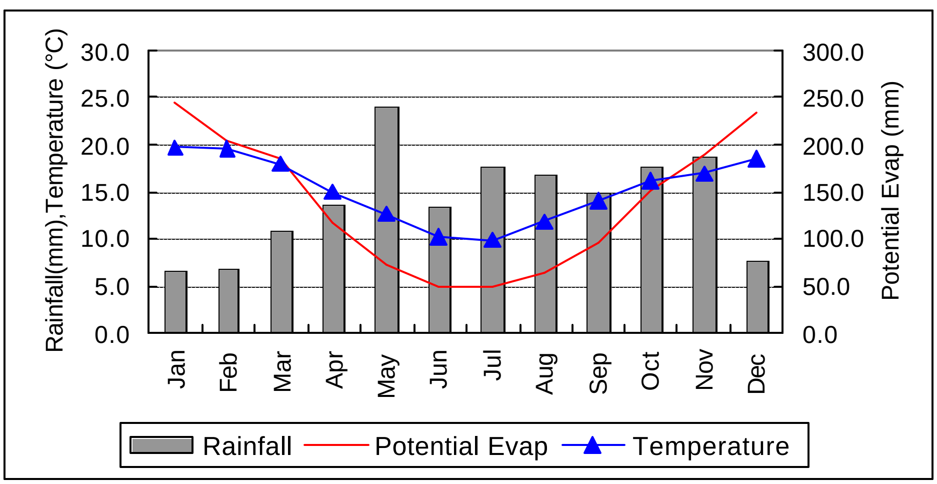


Figure 3.7 Average monthly temperatures, potential evapotranspiration and rainfall for Kiesiesvallei station (S-Pan values calculated from a percentage of MAE for evaporation zone in the study area after Midgley et al., 1994**)**

3.4.3 Potential Evapotranspiration

The average potential evapotranspiration (Symons pan) varies from 1300 mm in the south to 2000 mm in the north of the study area and exceeds the average annual rainfall generally. This factor is as much as 10-15 times the precipitation in some areas. The average monthly evapotranspiration values are shown in Figure 3.7 and indicate that the lowest values occur in June and July. The highest mean monthly values occur in December and January. The monthly average value in June is less than 50.0 mm. The monthly average value during the rainy season is approximately 123 mm, which corresponds to an average of 4 mm per day.

However, it is important to note that most rainfall in the Montagu area occurs as short storms. Therefore, on a daily basis, rainfall can exceed potential evapotranspirtation and thus potentially recharge the aquifer particularly through preferential flow paths by which rainwater can escape evapotranspiration.

3.5 DRAINAGE

There are two dendritic drainage patterns (Figure 3.8) in the study area that are controlled by geologic structures and lithology. They are the Touws Rriver drainage system in the north and the Breede River drainage system in the south. The Touws River finds its main source in the Hex River Mountain and flows from northwest to southeast. The Grootriver is a perennial tributary of it. The sources of the Groot River are the Swartberge Mountain and it runs from north to south. The two rivers merge in Donngkiaak and flow from west to east at last discharges into the Indian Ocean. Tributaries of the Breede River taking its source from the Langeberg Mountain Range run through the Langeberg Mountain Range and feed into the Breede River, which discharges in to the Indian Ocean. The main drainage direction in the study area is from northwest to southeast. The watershed of the two drainage systems coincides with the outcrop of the TMG. Most tributaries of the two drainages are ephemeral reaches and relatively are short. A few tributaries are ephemeral in the steep upper reaches, with more sustained flow in the lower reaches.

Runoff in the study area depends on the annual precipitation, the frequency of rainfall event, the topography, the nature of the soils and the geological characteristics. More runoff occurs in the Langeberg Mountain Range and the Hex River Mountain where higher annual precipitation occur. There are relevantly less drainage channels in flat areas than that in mountain areas. Midgley et al (1994) gave values of mean annual runoff for the area. It decreases from 200-500 mm in the west to 2.5-5 mm in the east of the study area. Based on the analysis of the 15 stream gauges, the modulus of runoff in the study area is $1.9\ell\cdot km^{-2}\cdot s^{-1}$ (refer to Appendix 15).

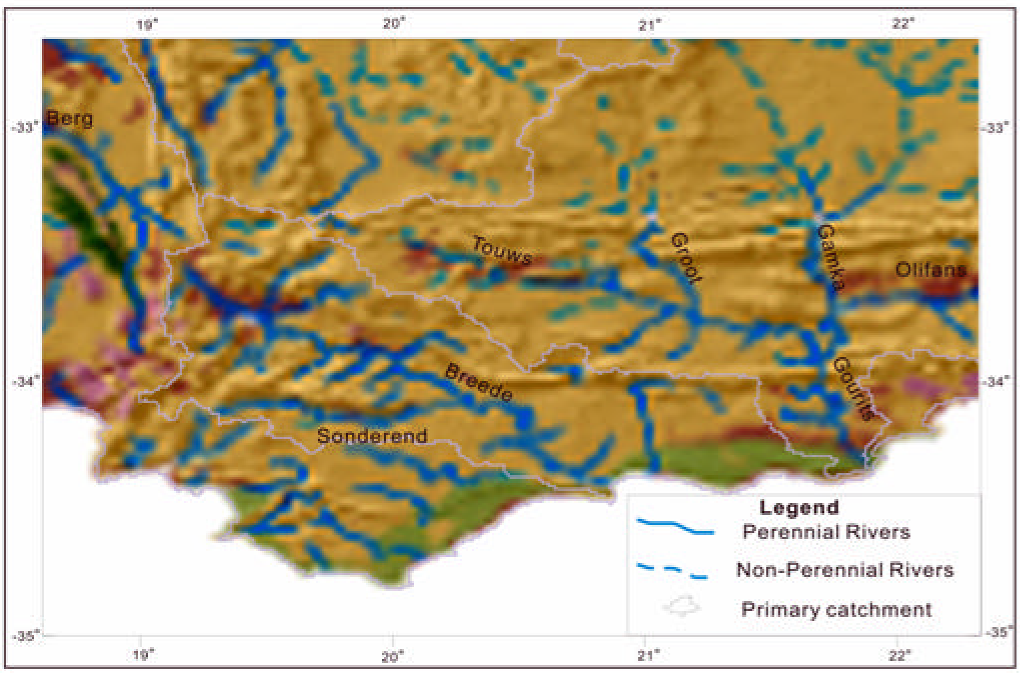


Figure 3.8 Drainage pattern for the study area

3.6 VEGETATION

Fybnos is a major indigenous vegetation type in the study area. It is a widespread vegetation type with the characteristic small-leaved and finely branched shrubs and reeds. Fybons species vary greatly with altitude and rainfall gradients and between similar habitats on different mountains and mountain ranges. Fybnos shrub with more Protea species distributes in the middle slops of the Langeberg as shown in Figure 3.9. A few annual grasslands with a few Protea species occur on the summit of the Langeberg range as can be seen in Figure 3.10 and Figure 3.11. The grasslands are spares in most outcrop of the TMG. Sparse vegetation distributes widely in the study area due to the steep slope and lithology, a typical example can be seen in Figure 3.12. It is important to note that indigenous vegetation has been replaced by increasing proportion of invasive alien plants (Pinus patula, Eucalyptus grandis and Black Wattle) on the foot of the mountains.

The density of the vegetation cover of the TMG is controlled by precipitation, soils and lithology. Areas of relatively dense vegetation cover are generally found in high precipitation areas. The area with vegetation cover above 80% (Figure 3.9) where the precipitation is above 1500mm per year. The vegetation cover density is below 15% in lower precipitation area (Figure 3.12) and even bare soil where total rainfall is < 200 mm per year. However, the vegetation cover density is influenced by soil too. Stony soil derived from the Peninsula sandstones has sparser vegetation than fertile soils derived from the Cedarberg Formation shale.

Figure 3.9 Fybnos shrub at Protea Farm (with more Protea species in the middle slop of the Langeberg)

Figure 3.10 Grass with a few Protea species in the Longeberg ridge

Figure 3.11 Grass land in the Langeberg ridge

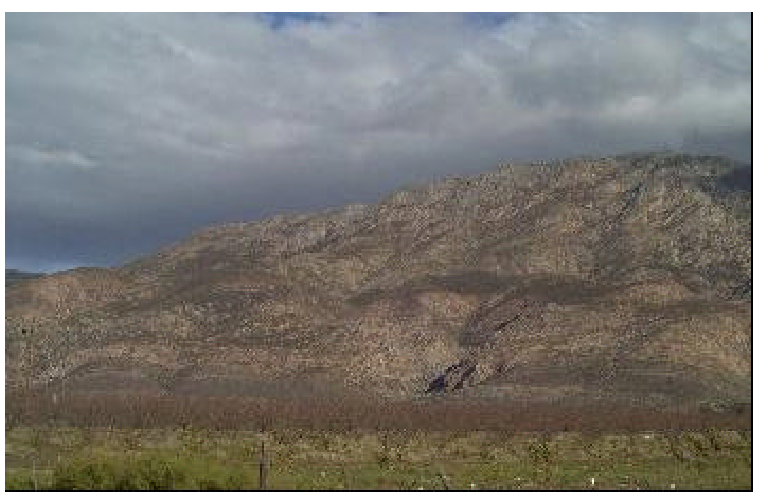


Figure 3.12 Sparse vegetation in the Montagu South

3.7 SOILS

The soils of the study area are highly influenced by the nature of the underlying host geology and the degree of weathering. Climate and geology play a large role in soil formation. In the semi-arid west and centre area, deep soils derived from the weathering or colluvial slopes of basement Bokkevveld Group shale, which are strongly structured and reddish colour with clay accumulation and are less acid and rich in nutrients. The soils in the outcrop of the TMG are derived from strongly mixed colluvial material and poor in nutrients. In the south and west mountain areas, rain mostly falls in the winter months. The soils form slowly and are generally thin and immature. Along the Langeberg Mountain in the extreme east, there are soils with a sandy texture-leached and with subsurface accumulation of organic matter, as well as iron and aluminium oxides. The soil depths in most of the area are shallower and limited by broken or solid rock outcrops. The soils derived from sandstone are often acidic, leached and mostly well drained. Generally the soils are friable and shallow; and their high gravel and rock content, low bulk density and high pore volume favour free water infiltration. The soil depths are thinner in both banks of Touws and Groot River.

3.8 GEOLOGY

The Klein Karoo basin fills took place from the early Ordovician to early Carboniferous times. The geology of the study area mainly consists of the sedimentary rocks of the Cape Supergroup, which lies unconformable on the Precambrian rocks (consisting of Malmesbury beds intruded by younger Cape granites). The Cape Supergroup is subdivided into the Table Mountain Group, Bokkevveld and Witteberg Group (Du Toit, 1954; Rust, 1967; Theron, 1972; Broqet, 1992). These successions of quartz arenites, shales and siltstones, with minor conglomerate and a thin diamictite were deposited in shallow marine environments under tidal, wave and storm influences, as well as in non-marine, braided-fluvial environments.

3.8.1 Lithostratigraphy

The geological divisions of the Montagu study area are presented in Figure 3.1 and Table 3.1 and discussed briefly below.

Table Mountain Group

A linear outcrop of the TMG occurs in the southwestern part of the study area along the Hex River Mountain, Langeberg Range and around the Warmwaterberg. The TMG consists of 95% quartz arenite with medium to coarse grain-size and variable amounts of feldspar and clay minerals. However, partings and thin beds of siltstone often separate these quartz arenite beds. These sandstones are of Ordovician to Silurian age (500 My) within in the all study area, only Peninsula Formation, Cedarberg Formation and Nardouw Subgroup of the TMG with outcrop area of approximately 3124 km^2 are represented.

Table 3.1 Geological sequence in the Montagu area

Super Group	Group	Sub-group	Formation	Lithology
Cape	Witteberg		Witpoort	Quartzitic sandstone, minor siltstone
		Weltevrede		Siltstone, mudstone thin-beded sandstones
	Bokkeveld	Bidouw		Siltstone,sandy shale
		Ceres		Siltstone, shale, sandstone
	Table Mountain	Nardouw	Rietvle	Feldspathic quartz renite/sandstone
			Skurweberg	Light-grey, thick-bedded felspathic quartze sandstone
			Goudini	Brown-weathering arenite, minor siltstone, shale
			Cedarberg	Thin black silty shale, siltstone and sandstone
			Peninsula	Largely thick-bedded, coarse-grained quartz arenite

Peninsula Formation comprises a succession of coarse-grained, white quartz arenite with scattered small pebbles and discrete thin beds of small-pebble, matrix-supported conglomerate. The pebbles are normally vein quartz, but sometimes (Rust, 1967) consist of black oolitic chert. The thicknesses of the Peninsula Formation in the study area vary from 600.0 m to 2000.0 m (Rust, 1973; Meyer, 2001). It is considerably thicker in the south than that in the north due to the severe folding, thrusting (Booth and Shone, 1992) within the formation.

Cedarberg Formation is an important marker unit, which interruptes the monotonous arenitic character of the TMG. The Cedarberg Formation is a thin (about 60.0 m in the study area), but remarkably continuous unit, consisting of black silty shale at the bottom, grading into brownish siltstone and fine sandstone at the top. According to Broquet (1992), the Cedarberg Formation was probably deposited when the basin was depressed glacio-isostatically, leading to a rise in sea-level and a decrease in sediment influx. Its confining character makes the Cedarberg Formation very important in a hydrogeological sense.

Nardouw sub-group, with its three subdivisions, the Goudini, Skurweberg and Rietvle Formations, is another thick unit of sandstone that varies between quartz arenite, silty and feldspathic arenites, accompanied by some very minor inter bedded conglomerate and shale. This lithological diversity, together with texture, grain size and bedding thickness differences, lead to pronounced differences in weathering, structural and hydrogeological characteristics. The basal unit, the Goudini Formation, is characterized by reddish weathering; thin sandstone beds with common shale intercalations, which are less resistant to weathering than the thick-bedded, arenitic Skurweberg Formation. The topmost unit, the Rietvlei Formation, is easily recognized by finer grain size, common high feldspar contents and has a more denser vegetation cover, which is visible on aerial photographs as darker tones of grey, compared to the lighter Skurweberg Formation. The contact with the overlying dark shales of the Bokkeveld Group is usually abrupt. The Nardouw Subgroup is approximately 600.0 m thick in the study area (Meyer, 2001).

Bokkevveld Group

Bokkevveld Group sediments mostly overlie the Table Mountain Group over vast flat areas of the Montagu area. The Bokkevveld Group is composed of two subgroups in the study area, namely the basal Ceres subgroup and the overlying Bidouw Subgroup.

The lower or Ceres subgroup consists of siltstone, shale, and sandstone. The Bidouw Subgroup includes fossiliferous shale and sandstone.

Witteberg Group

Witteberg Group including the Weltevrede Subgroup and Witpoort Formation is presented in the study area. The Witpoort Formation overlies these subgroups. The Witteberg Group mainly consists of quartzitic sandstone, minor siltstone.

Alluvial deposits occur along the floor of the steep-sided river and consist generally of an upper boulder bed overlying sand and gravel layers. The depth varies greatly from the valley floor to tributaries enter the valley.

3.8.2 Structure

The Cape Fold Belt (CFB) is a largely east-west striking feature located rough south of 33° S. It consists predominantly of sedimentary and metamorphic rocks, which were subjected to intense pressure, particularly form the south. The development of the CFB was modified by two major orogenic events, namely the Permo-Triassic Cape Orogeny and the fragmentation of southwestern Gondwana during the Mesozoic. Rocks of the Cape Supergroups were deformed by the Permo-Triassic Cape Orogeny (Söhnge and Hälbich, 1983). The outcrop pattern of the Cape Supergroup, namely parallel mega-anticlinal mountain ranges, separated by synclinal intermontane valleys, reflects the main structural features of the Cape Orogeny. The CFB consists of western branch and southern branch. Both branches merge with northeast-trending folds in the syntaxis of the southwestern Cape (De Villiers, 1944; Söhnge and Hälbich, 1983). The study area is located between the southern branch and the syntaxis of the CFB.

The southern branch, comprises north-verging, often overturned first-order folds, sliced by a few thrusts (Theron, 1962; Booth and Shone, 1992) and normal faults, with strong fracture cleavage in the quartz arenites and slaty cleavage in the Cedarberg Formation. The major faults trend easterly, but are accompanied by a transverse set of minor, approximately penecontemporaneous northeast-trending transfer faults, suggesting elements of tension. All of these faults display zones of brecciation a few tens of metres wide, cataclasm and numerous minor splays, as well as horse-tailing. The major large faults presented in the study area are summarized below:

- Touwsriver-Herbertsdale fault zone: a normal NWW-SSE striking system with an acute angle ranges from the Hex River Mountain in the west of study area, via Touws to Van Wyksdorp. It is arcuate in plan view and convex towards the Karoo Basin accompanied by a transverse set of minor, approximately penecontemporaneous northeast-trending transfer faults in the end of east. In the west end, it displays zones of cataclasm, and numerous minor splays, as well as horse-tailing. The exposed width and length is 10.0 km and 200.0 km, respectively. It forms the northern boundary.

- Worcester-Swellendam-Herbertsdale fault zone: it is a regional fault zone consisting of Swellendam-Herbertsdale normal fault and Worcester thrust faults. The Swellendam-Herbertsdale normal fault is a W-E striking system. It extends along the Langeberg Mountain from Robertson to Herbertsdale where it vanishes. This fault displays about ten metres wide and 180 km long. It joins the Worcester fault zone at the west end. The Worcester thrust fault zone changes trend from easterly to northwesterly and the area south of it is dominated by numerous large northeasterly-trending faults. The latter set of fractures must have formed contemporaneously with the Worcester fault or slightly later, because they end against this major fracture, which attains its maximum displacement of more than 5.0 km in the syntaxis. The faults in the study area ranges from the Hex River to Montagu and in 50.0 km in lengths and dips below 45º. These two faults form an arcuate plan view and convex to the Karoo Basin. It is the south boundary.

The area where the fold axial traces of the western and the southern branches merge is defined as the Syntaxis Domain. It is the most fractured part of the CFB, with components of the western and southern branch faulting both being present. This area consists of varied northeast striking faults (Gresse and Theron, 1992). Based on differing fold trends and shortening intensity, the syntaxis may be divided into two separate domains lying north and south of the Hex River anticline. The northern domain is characterized by north, northwest, northeast and minor east tending folds, while the southern domain contain only east and northeast trending folds. The curvature of both branches to form oroclinal arcs in the syntaxis, which merge with an intermediate trend, suggests their formation by roughly simultaneous northeast-southwest and northwest-southeast directed shortening (Newton, 1975; De Beer, 1990; De Beer, 1995). The Worcester Fault changes trend from easterly to northwesterly and the area south of it is dominated by numerous large northeasterly-trending faults. The major structure in the syntaxis of the study area is the Hex River Fault. The Hex River fault strikes NE-SW and ranges along the Hex River Mountain. It extends through the Touwsriver-Herbertsdale fault zone disappearing in the north and meets the Worcester-Swellendam-Herbertsdale fault zone in the southern edge. It is a minor normal fault with tens of metres wide and 20.0 km long and is near-vertical. It forms the west boundary of the unit.

3.9 HYDROGEOLOGY

Three aquifer systems occur in the Montagu area, namely sandstones of the TMG, shales and sandstones of the Bokkeveld Group and alluvium zone. The Bokkeveld Group shale and siltstone form one of the aquifer units within the Cape Subgroup. The rocks of the Bokkeveld Group are hydraulically interconnected with the TMG aquifer system and it is often the major aquifer in terms of direct exploitation in the Hex Valley area. It acts as an aquitard due to its predominant argillaceous characters in most study areas.

3.9.1 Aquifer

The alluvium zone consists of unconsolidated alluvial deposits and a 50-100m-thick, “fractured-and-weathered” or “regolith ” layer of the TMG strata (Hartnady and Hay, 2002). In flood plain areas the alluvium is finer grained. It is a type of reservoir for groundwater storage and feeds the groundwater into the underlying TMG aquifer.

The sandstones of the TMG form the main aquifer unit. The TMG aquifer is characterized mainly by sandstone members, which are alternating felspathic quartz sandstone and coarse-grained quartz arenite units with interbedded minor conglomerate and shale. Depth of groundwater level varies from 8.0 to 40.0 m below surface with an average depth of about 16.0 m (National Groundwater Data Base). The major aquifers in the study area are the Peninsula Formation and the Nardouw sub-group.

The Peninsula Aquifer is a pure quartz arenite with a very low primary porosity. However, it is a brittle quartzitic which would tend to fracture readily under stress and lead to blocking of fractures and thus secondary porosity. There are large variations of the hydraulic conductivity in this aquifer. Unfractured rock possesses low hydraulic conductivity and highly fractured rock or single, large fractures have very high hydraulic conductivity (Rosewarne, 2002).

The Nardouw Aquifer consists predominantly of quartz arenite containing silty / shaley interbeds and higher feldspar content. It is prone to ductile deformation, and generally has lower hydraulic conductivity than the Peninsula Formation. The secondary porosity in fractured strata is approximately double that of folded strata and more. Shale is impervious intercalation; it has a great impact on the aquifer properties. Groundwater movement is in the form of seepages along most shale intercalations. Otherwise, production from the chemical weathering of feldspar may clog secondary groundwater flow paths and reduce permeability further.

3.9.2 Regional flow pattern

Groundwater occurrence is related to distribution of the basement. The outcrops of the TMG are exposed in mountainous areas, which in turn influences precipitation distribution to a significant extent and also forms the recharge area with hydrodynamic head. Groundwater generally moves from levels of higher energy to levels of lower energy. An intricate network of fissures, joints, fractures and even cavities govern the infiltration, storage and transmission of groundwater in the largely competent and brittle-natured arenaceous units of the TMG. The aquifer is considered to be continuous on a large scale, despite the complicated structural control of the subsurface consisting of a series of horst and graben features. On a local scale, the shallow circulating springs seep from a network of joints, small, irregular fractures and from bedding planes within the TMG. Their yields are highly seasonal. Springs issue from contacts with interbeds of the Cedarberg Formation. Saturation zones of the Peninsula Formation commonly results in the formation of springs on the Peninsula Formation/Cedarberg Formation contact at suitable topographical levels, from where it overflows onto the Nardouw Subgroup. The Touwsriver and the Breede River are important discharge areas on a local scale.

The TMG, notably the often-fractured arenaceous components, is largely anisotropic, and thus does not display uniform aquifer characteristics. Table 3.2 lists information of boreholes in the eastern section of the Koo valley in the Langeberg (SRK phase B interim report, 2002). The variety of thickness, blow yield and EC imply that there are different groundwater flow paths and recharge source.

Table 3.2 Information of boreholes in the Koo valley (AfterSRK report)

Borehole No.	Water strike			
	Number	Depth (m)	Blow yield (l/s)	EC (mS/m)
Koo 02/01	1	12-14	0.7	20
	2	34-38	1.8	25
	3	45-47	2.8	26
	4	113-118	10.2	12
Koo 02/02	1	36-38	0.2	
	2	54-58	0.6	26
	3	101-104	0.6	16
	4	140-143	4	10
	5	146-149	4	10
	6	167-169	5.6	
	7	231-236	16.5	10
Koo 02/03	1	80-82	0.5	
	2	86-87	0.8	
	3	192-193	14.61	19
	4	209-210	21	19
Koo 02/04	1	61-65	5.59	
	2	104-107	10.17	
	3	144-146	6.94	
	4	170-172	10.2	
	5	204-207	21	
Koo 02/05	1	36-40	5.59	116.9
	2	66-70	4.42	116.9
	3	90-94	5	18
	4	150-154	8.46	15
	5	165-167	8.4	14

Three thermal springs occur in the Cape Supergroup rocks in the study area, namely Montagu (Avalon) spring, Baden spring and Warmwaterberg spring (Table 3.3). The

Montagu spring and Baden spring are situated in the syntaxis domain; the Warmwaterberg spring is situated in the southern branch of the CFB. The three hot springs are located at the intersection between faults in the Nardouw Subgroup (TMG) and the basal shale layer of the Bokkeveld Group. Probable depths of circulation are from 1500 m at Baden, 2000 m at Montagu to 2100 m at Warmwaterberg (Meyer, 2002). They are all strong yielding with total flow of 2.65Mm^3/yr and seasonal fluctuations are limited. The environmental isotope signatures of groundwater from springs indicate that recharge takes place at much higher altitudes than occur locally and deep groundwater flow along regional faults transports groundwater to its current discharge points (Kotze, 2000).

Table 3.3 Information of hot springs in the study area (After Meyer, 2002)

Name of spring	Co-ordinate		Temp. (°C)	Yield (l/s)	Cond. (mS/m)	Probable depth of circulation (m)
	South	East				
Montagu	33°45´57"	20°07´02"	43	38	11	2000
Baden	33°42´20"	20°07´33"	38	37	10	1500
Warmwater-berg	33°45´57"	20°54´08"	45	9	26	2100

CHAPTER 4
METHODOLOGY

Recharge forms a part of hydrologic cycle and is influenced by lots of factors. Water balance method estimates recharge based on the conservation of mass. A wide range of variables controls actual groundwater recharge and as a result site-specific estimates require a large amount of diverse data. However, the data, which are not available for the entire study region and this, restricts the use of more sophisticated methods. Modelling simplifications, such as using similar tools across a range of soils, are therefore necessary to reduce complex soil processes to a manageable level of sophistication for regional application. The proposed empirical model considers only the impact of soil and vegetation on recharge, which reduces the influences of limited data of other factors.

The methodology used to estimate recharge using the water balance method in vicinity of Montagu and then to analyse the experimentally obtained data and also data reported in the literature to conclude temporal and spatial variability of groundwater recharge. The study incorporated the following main steps in the research approach:

- Desk study
- Fieldwork
- Identification of appropriate estimation methods
- Calculation of recharge rates and volume
- Comparison of results with earlier studies

4.1 DESK STUDY

4.1.1 Review

The desk study involved review of all relevant available information, including:

- A literature review on recharge mechanisms and estimates in fractured rock media
- A literature review of previous studies of the TMG aquifer system (including hydrogeological maps, remote sensing, recharge and other data).

4.1.2 Data collection

The main data collected:

- Geological information
- Boreholes and wells and their hydrologic information drilled in the study area
- Monthly rainfall data, maximum and minimum temperature
- Vegetation factor and soil information

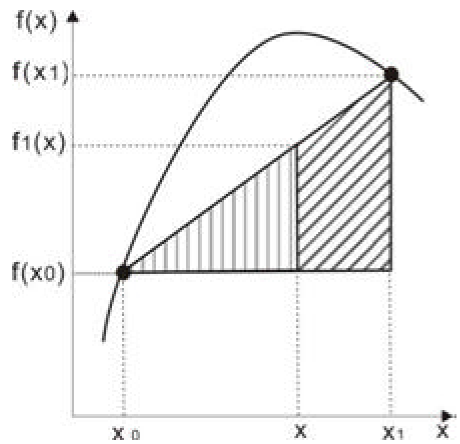
The main data sources were captured from the REGIS Database at Department of Water Affairs and Forestry (DWAF), South Africa; Geographical Information Systems of Cape Nature Conservation (CNC) and Water Research Commission (WRC) WR90 Flow data.

The information of geomorphology, vegetation and soils was obtained by field investigations from 2003 to 2004 in the study area.

4.1.3 Data collation

Some incomplete or incorrect data sets must be patched or estimated scientifically before they are used for modelling. The format of the data sets is different in the different databases, which must be harmonized. The missing data were interpolated as follows:

If a time series is incomplete in a period, the data have to be estimated using intermediate values. The simplest form of interpolation is to connect two data points with a straight line. This technique, called linear interpolation, is depicted graphically in Figure 4.1. Using triangles (Chapra and Canale, 1998), three data $f(x_0)$, $f(x_1)$ and $f_1(x)$ can be write as:



The shaded areas indicate the similar triangles used to derive the linear-interpolation formula [equation (4.1)] (After Chapra and Canale, 1998)

Figure 4.1 Graphical depiction of linear interpolation

$$\frac{f_1(x) - f(x_0)}{x - x_0} = \frac{f(x_1) - f(x_0)}{x - x_0}$$

which can be rearranged to yield

$$f_1(x) = f(x_0) + \frac{f(x_1) - f(x_0)}{x - x_0}(x - x_0) \qquad (4.1)$$

4.1.4 Data analysis

The statistical methods used in the study include the arithmetic, Cumulative deviation, standard deviation and coefficient of variation (refer to Neter et al., 1988).

Arithmetic mean

The equation for the Arithmetic mean is:

$$\overline{X} = \frac{1}{N}\sum_{i=1}^{n} X_i \qquad (4.2)$$

Cumulative deviations of the X_i values from their mean $\overline{X}$:

$$CD=\sum_{i=1}^{n}(X_i-\overline{X})=0$$

The cumulative rainfall departure (CRD) is defined as

$$CRD=\sum_{i=1}^{n}(X_i-\overline{X})=0 \tag{4.3}$$

Standard deviation

The most commonly used measure of variability in statistical analysis is called the variance. It is a measure that takes into account all the values in a set of items. The variance s^2 of a set of values $x_1, x_2, \ldots, x_n$ is defined:

$$s^2=\frac{\sum_{i=1}^{n}(x_i-\overline{X})^2}{n-1} \tag{4.4}$$

The standard deviation is a measure of absolute variability in a set of items. Frequently, the relative variability is a more significant measure. The most commonly used measure of relative variability is the coefficient of variation. The positive square root of the variance is called the standard deviation and is denoted as by s:

$$s=\sqrt{s^2} \tag{4.5}$$

Coefficient of variation (C_v) is the ratio of the standard deviation to the mean expressed as a percentage:

$$C_v=100\frac{s}{\overline{X}} \tag{4.6}$$

4.1.5 Error analysis

Error analysis was conducted according to the influence factors of recharge rate. The errors are discussed and compared with different results obtained in earlier studies. The recharge estimates with the water balance method are dependent on the temperature, latitude, soils, and classes of vegetation and cover percentage and precipitation. Therefore, the error analysis concentrated on the variation of the factors. The degrees of the sensitivity are related to the soil, vegetation cover and rainfall time factors.

4.2 FIELDWORK

The fieldwork performed in the study include:

- Geomorphological investigations using satellite images
- Hydrogeological survey

4.3 RECHARGE ESTIMATION

Despite numerous uncertainties associated with the simple soil-water budget model, it is still used in many studies from catchment scale to the global water balance and climatic change scenarios (Thornthwaite, 1948; Shiklomanov, 1983; Manabe, 1969; Mather, 1972; Alley, 1993; Willmott et al., 1985; Mintz and Walker, 1993; Mintz and Serafini, 1992). In this chapter the water balance methods used for estimating groundwater recharge in semi-arid environment were presented. The factors affecting recharge are discussed in light of their applicability to the Western Klein Kraoo, denominated as Montagu area.

4.3.1 Principle

Quantification of the hydrologic cycle defines a hydrologic budget equation, or water balance, that describes the hydrologic regime in a catchment, and it is based on the conservation of mass. The hydrological cycle is the process by which moisture is evaporated from the oceans, and ultimately, after complex processes, returns to the oceans. Figure 4. 2 provides a detailed pictorial representation of the process revolving around the cycle.

Inflow to the hydrologic system arrives as precipitation, in the form of rainfall or snowmelt, and outflow takes place as stream flow (or runoff) and as evopatranspiration, which is a combination of evaporation from open bodies of water, evaporation from soil surfaces, and transpiration from the soil by plants. Precipitation is delivered to streams both on the land surface, as overland flow to tributary channels; and by subsurface flow routes, as interflow and baseflow following infiltration into the soils. The body of water that will eventually reach the water table is groundwater recharge. Figure 4. 2 indicates that the process of rainfall recharge to ground water is determined by factors which occur in almost the entire hydrological cycle. These factors are inter-related, complex and natural phenomena, which are governed by the natural laws of conservation of energy, mass, and momentum. The determination of recharge therefore includes many scientific disciplines.

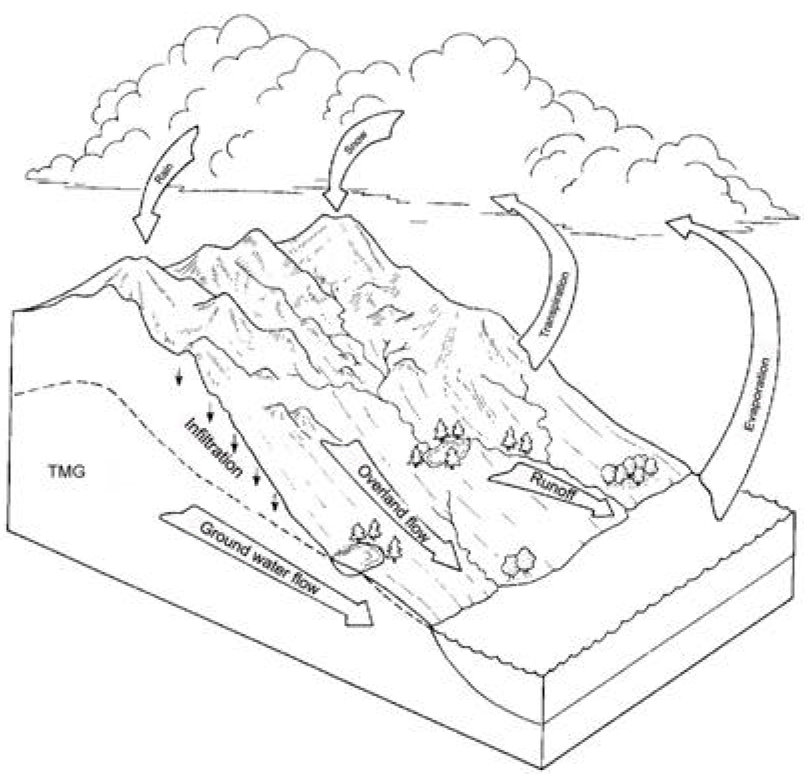


Figure 42 Systems representation of the hydrologic cycle

4.3.2 Model

The present study was motivated by the need to estimate recharge with the water balance approach for the semi-arid Montagu area, in the context of the controlling climatic and landscape characteristics. The study area, lacked information on water level information, spring yields and aquifer properties. The reliability of groundwater recharge estimation depends on the accuracy of landscape feature description inside the water balance model, without any possible calibration. The proposed empirical model yields natural groundwater recharge on a monthly basis, evaluated as the difference between the inflows (rainfall) and the outflows (evapotranspiration, surface runoff).

The different terms of the water balance at a catchment scale are defined as:

$$R_e(t) = P(t) - E_t(t) - Q(t) - \Delta q(t) \tag{4.7}$$

where $R_e(t)$ is groundwater recharge, $P(t)$ is precipitation, $E(t)$ is actual evapotranspiration, $Q(t)$ is runoff and ? ?*(t)* is change in storage of groundwater content. The term *(t)* designates that the terms are distributed through time. Monthly records of a site $P(t)$ were obtained from the WR90. $E(t)$ and $R(t)$ were initially estimated based on meteorological and site-specific data by simple methods described in a later section. However in catchments we are limited, if we average over many years of record, it can be assumed that ? ?*(t)* is negligible between subsequent years. This is a reasonable assumption for the climate of semi-arid area, since there is very little carry-over of soil moisture between years.

4.3.2.1 Estimate of evapotranspiration

A common approach for approximating E_t (t) involves first estimating evapotranspiration for reference vegetation E_{tr} (t), such as grass or Fynbos. E_t (t) is then calculated by multiplying E_{tr} (t) by a complex coefficient, K_c (t), an experimentally defined crop-specific variable whose value varies throughout the growing season, such that:

$E_t(t)=K_c(t)\,E_{tr}(t)$ (4.8)

This study estimates $E_{tr}(t)$ by applying the method of Hargreaves and Samani (1985), a technique based solely on temperature and latitude, and represented by:

$E_{tr}(t)=0.0023R_A T_D^{1/2}(T+17.8)$ (4.9)

Where

T_D=difference between average monthly maximum and minimum temperatures (°C)

T=average monthly temperature (°C)

R_A=extraterrestrial radiation ($MJm^{-2}d^{-1}$)

Duffie and Beckman (1980) represent R_A by:

$$R_A=G_{SC}d_r\frac{(w_s)\sin(f)\sin(d)+\cos(f)\sin(d)\sin(ws)}{p} \qquad (4.10)$$

Where

d=0.4093sin[2p(284+J)/365]

$?_s$=arcos [-tan (F) tan (d)]

F=latitude (radians; positive for north, negative for south)

Dr=1+0.033cos(2pJ/365)

G_{SC} is the solar constant with a value of $118.1MJm^{-2}d^{-1}$, and J represents the number of days since January 1 of the current year. Values from equation 4.10 were converted to units of md^{-1} by:

$$E_{tr}(t)[md^{-1}]=\frac{E_{tr}(t)[MJm^{-2}d^{-1}]}{l[MJkg^{-1}]r[kgm^{-3}]} \qquad (4.11)$$

Where ? is the latent heat of vaporization, given by (Harrison, 1963):

? $[MJkg^{-1}]=2.501-2.361\times10^{-3}T$[T in°C] (4.12) and ? is the density of water.

Complex coefficient Kc (t) is defined a variable whose value varies throughout the precipitation, vegetation cover and soil type and is expressed as:

$$Kc\,(t) = P\,(t) + C_s\,(t) + C_v\,(t) + C_c\,(t) \quad (4.13)$$

Where
P is the total precipitation per year (mm)
C_s is the soil factor
C_v is the vegetation factor
C_c is the coverage of vegetation

The C_v values for fynbos are estimated from the WR90 database presented by Midgley et al. (1994)

4.3.2.2 Estimate of runoff

A common method to estimate Q is to apply the SCS (Soil conservation Service) runoff equation (USDA-SCD, 1985), which is a simplified method for estimating rainfall excess that does not require computing infiltration and surface storage separately. Both processes are included as one of runoff watershed characteristic. The excess rain volume (runoff) depends on the amount of precipitation and the volume of total storage (retention). The runoff is predicted by the SCS equation:

$$Q = \frac{(P - 0.2S)^2}{P + 0.8S} \quad P \geq 0.2S \quad (4.14)$$

and

$$S = \frac{1000}{CN} - 10 \quad (4.15)$$

Where Q is the runoff volume, P is the total rainfall, and S is the retention factor and CN is the curve number.

4.3.3 Model calibration

The parameters related to the soil and vegetation are empirical values. These values should be calibrated using scientific method. The calibration in the model is conducted as follows:

- The factors of vegetation adopted from Midgley et al. (1994) remain constant in the model.
- The variation of temperatures including maximum and minimum temperature is controlled ±2°C because the long-term (1920-1989) change of the average temperature should not exceed range of 4°C. The range of the temperature is usually controlled within ±0.5°C.
- The coverage percentage of vegetation is estimated based on field investigation, the distribution of lithology and the feature of topography including slope gradient and land surface forms.
- The soil factors are changeable parameters in the model. The soils factor values are empirical values in initial calculation process; the values are given in referred to the type, thickness and distribution of soil defined by Midgley et al. (1994). The soil factors are given based on the trial calculation in the study area but the error range does not exceed 10% under similar condition of lithology and topography.

CHAPTER 5

ANALYSIS OF IMPACT FACTORS ON RECHARGE

5.1 INTRODUCTION

Recharge is influenced by a wide variety of factors including the climate, vegetation, topography, geology, and soils. Special aim focuses on the determination of spatial and temporal variability and tendency of rainfall in the Montagu area in order to facilitate the evaluation of the recharge in response to change of rainfall. The impacts of precipitation characteristics such as precipitation type, intensity, duration and distribution on actual evapotranspiration and direct runoff are discussed in detail.

5.2 IMPACT FACTORS ON RECHARGE

Recharge forms a part of hydrologic cycle and is influenced by lots of factors, which are classified as climatic factors, physiographic factors and geological factors.

The climatic factors include:

- Precipitation
- Temperature
- Solar radiation
- Humidity
- Wind

The physiographic factors include:

- Topography
- Vegetation
- Soil
- Land use
- Drainage area

The geological factors include:

- Lithology
- Structure
- Hydrologic characteristics

The physiographic features (including geological factors) influence the occurrence of groundwater recharge within a region and these, particularly the topography, play a significant role in influencing the precipitation and other climatic factors, such as temperature, humidity and wind. However, within a geographical location, it is primarily the rainfall (its intensity, duration and distribution) and the climatic factors associated with the physiographic factors affecting evapotranspiration and runoff, which are good index of the groundwater recharge in a region.

5.3 IMPACT OF PRECIPITATION FACTORS ON RECHARGE

The precipitation is the most important input to recharge estimation and is influenced mainly by climatic change, including global and regional climatic change, which affects precipitation characters such as pattern, amount, spatial distribution, periodicity, tendency, intensity and duration. The climate in the Montagu area is largely influenced by maritime air from the southern waters of the Indian Ocean and western waters of the Atlantic Ocean. The rainfall is mainly cyclonic and orographic with occasional thunderstorms (Kiker, 2000). As a result the rainfall amount and rainy season vary greatly. The temporal and spatial variation of rainfall results in similar variation in the recharge. Generally, the recharge increases with increasing precipitation under favorable condition.

5.3.1 Spatial distribution of precipitation

In this section rainfall data for 17 long-term records of rainfall stations lying within or around the outcrop of the TMG are presented. The spatial distribution of annual average rainfall using the Kriging's method in the study area is shown in Figure 5.1. It can be seen that the rainfall increases from north to south. The outcrop of the TMG receives less than 400 mm/yr of rainfall. In west of the Montagu, the mean annual precipitation increases from less than 200 mm along the inland foothills, fringing the Little Karoo to more than 400 mm at the Moedverloreberg and Longeberg Range. In east of Montagu, along the Langeberg Range, interior foothills may receive as little as 200 mm and upper slopes as much as 1,000 mm year of precipitation. The precipitation varies from less than 200 mm to 400 mm around the Warmwaterberg. The area around the Baden hot spring receives about 200 mm year rainfalls. It is note that the distribution areas of high rainfall in Figure 5.1 may not be accuracy because a few rainfall stations are located in the outcrop area of the TMG.

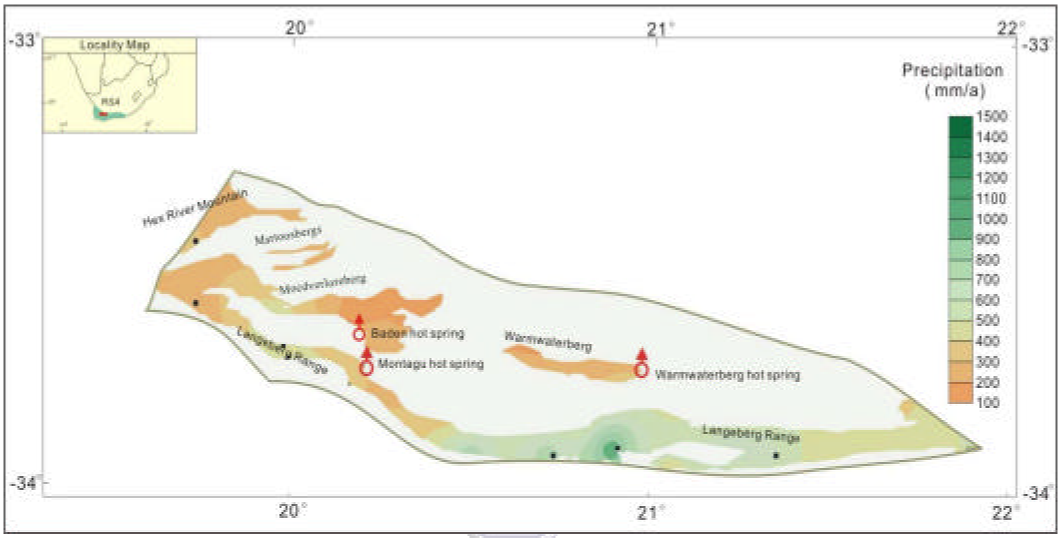


Figure 5.1 Average annual precipitations in the study area

A precipitation statistics for the study area are listed in Table 5.1. Complete rainfall data are presented in Appendix 1 to Appendix 4 for the 17 rainfall stations. All of the data are based on the long-term rainfall records attached in WR90. The longest statistical periods are from 1920 to 1989 at rainfall station 0025599. The raw records in most rainfall stations are not complete. The results show that the minimum monthly precipitation ranges from 0.0 mm to 37.8 mm in November at rainfall station 0025270. In most areas, minimum monthly precipitation is less than 5.0 mm. The maximum monthly precipitation ranges from 31.0 mm in December for rainfall station 0023070 to 306.6 mm in January for rainfall station 0023611. The maximum monthly precipitation exceeding 100.0 mm makes up approximate 50% of the study area. The average monthly precipitation ranges from 5.6 mm in December for rainfall station 0023070 to 108.6 mm in March for rainfall station 0025599. The standard deviation monthly precipitation ranges from 9.5 mm in July for rainfall station 0025162 to 74.0 mm in June for rainfall station 0043239. The MAP varies from 132.4 mm at rainfall station

0025162 to 999.0 mm at rainfall station 0025599. The annual maximum precipitations vary from 251.8 mm at rainfall station 0024101 to 1531.7 mm at rainfall station 0025599, and the annual minimum precipitation are from 40.5 mm at rainfall station 0025162 to 446.4 mm at rainfall station 0026240.

Table 5.1 Outline of precipitation statistic analyze in the study area

Station	Period	MAP (mm)	Maximum (mm)	Minimum (mm)	Stdev
0007311	1932-1974	367.1	607.5	204.7	77.3
0008136	1924-1974	387.4	613	144.7	99.8
0008782	1923-1989	682.8	947.9	399	126.6
0011451	1968-1989	418.4	828.6	197.3	136.2
0023070	1978-1989	235.0	418.2	131.7	79.9
0023218	1937-1950	261.9	382.6	160.9	63.7
0023602	1947-1989	274.5	489.8	144.1	82.5
0023611	1927-1987	494.2	968.7	276.2	134.9
0023706	1920-1946	392.1	520.1	197.5	81.5
0024101	1931-1965	169.0	251.8	101.9	47.5
0024684	1931-1952	293.3	438.4	188.8	71.4
0025162	1920-1974	132.4	265.3	40.5	55.7
0025270	1924-1938	764.4	944.6	561.1	114.5
0025414	1925-1989	280.9	563.7	50.1	100.1
0025599	1920-1989	999.0	1531.7	298.6	233.6
0026240	1928-1962	643.1	973.1	446.4	130.9
0026510	1936-1989	635.6	1078.1	383	144.9
0026824	1969-1989	232.0	512.8	95.9	99.0
0043239	1920-1939	286.7	606.6	180.9	95.1

5.3.2 Seasonal distribution of precipitation

The seasonal distribution of the precipitation is related to the climatic zones. The rainy season shifts from winter in the west towards all year in the east. A Mediterranean climatic zone dominates in the extreme west of the study area around the Hex River Mountain where rainfall arrives exclusively in the winter months. The records of four rainfall station show precipitation patterns in these regions (Figure 5.2) where the precipitation occurs exclusively in May, June, July and August. The amount of rainfall within the four months accounts for above 50% of amount of annua l rainfall. Average monthly maximum values range from 38.40 mm to 78.2 mm in August. December, January and February are the driest months and the average monthly rainfall is less than 20.0 mm; especially at station 0043239, the rainfall during the three months are less than 40.0 mm.

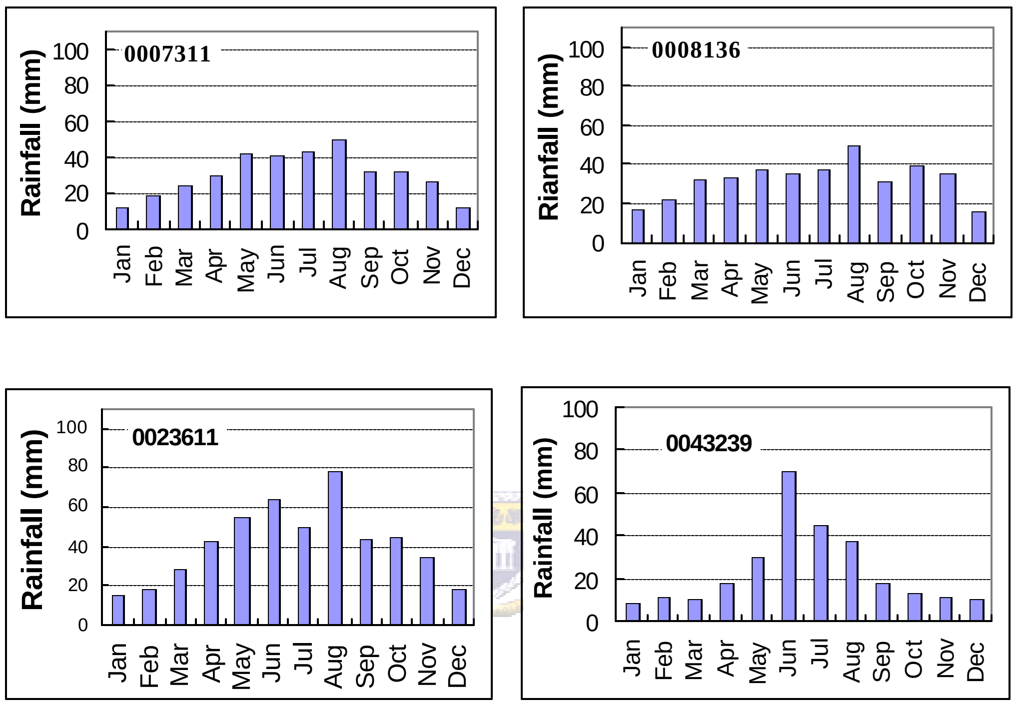


Figure 5.2 Rainfall pattern in winter rainfall area

In the most areas, the annual rainfall averages is less than 400 mm, although the northern foothills may receive as little as 200 mm. Generally, wetter period is from May to August and the monthly average rainfall is less than 40 mm except rainfall station 0011451. Calendar months of December, January and February are drier. The minimum of monthly average rainfall is less than 20 mm, especially rainfall station 0025162 with monthly rainfall less than 10 mm (Figure 5.3).

In the east of the study area, the mean annual rainfall is about 750 mm and as high as 1500 mm on the mountains. Rainfall occurs year-round and the distribution is bimodal. Wetter periods are from March to April and October to November. The maximum values vary from 60 mm to 109 mm per month. Drier months are from June to July and

December to January. The minimum precipitation is about 40 mm per month (Figure 5.4).

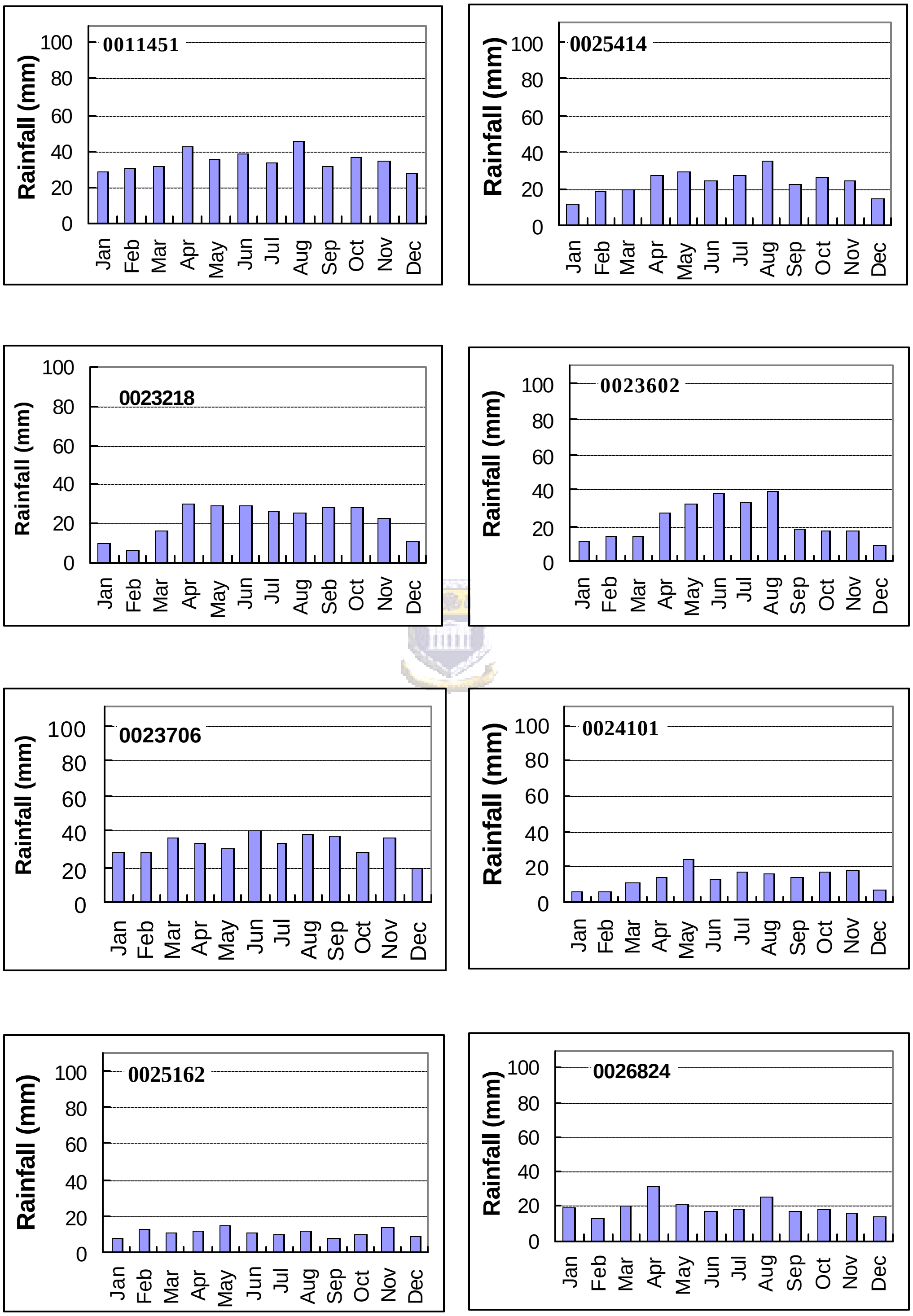


Figure 5.3 Rainfall pattern in semi-arid area

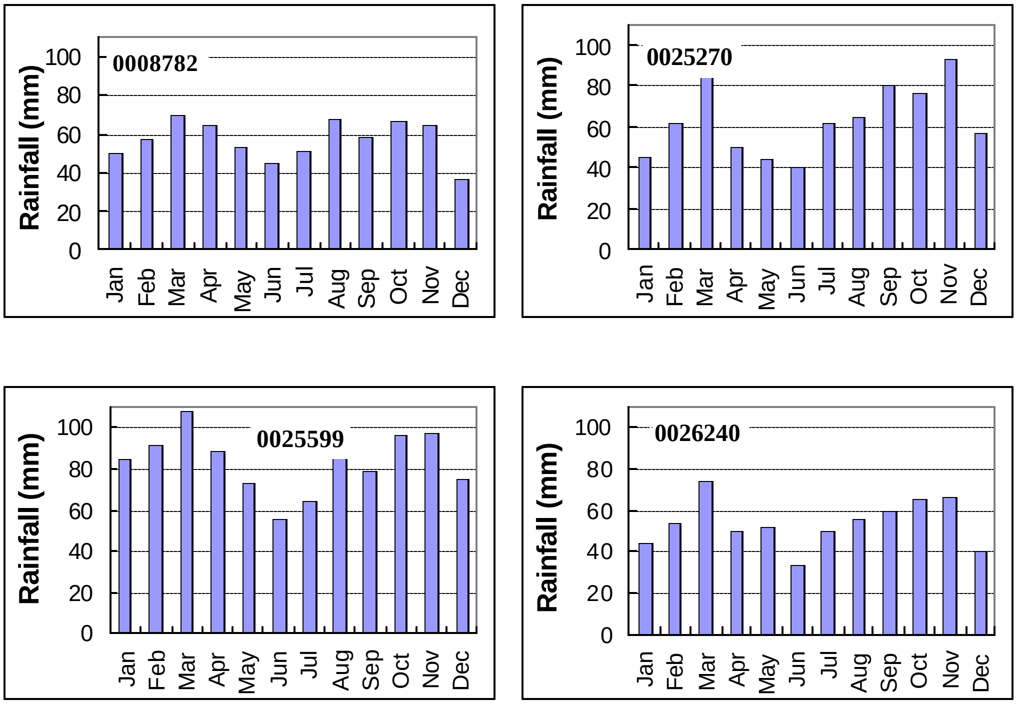


Figure 5.4 Rainfall pattern in all year rainfall area where rainfall is bimodal

5.3.3 Comparison of rainfall patterns

According to correlation analysis from 17 rainfall stations with long-term records (Table 5. 2), five groups of precipitation patterns were identified (Table 5.3). The results show that the rainfall patterns of the study area vary greatly although the study area is located in standard homogeneous rainfall districts denoted as No 7 and No. 8 defined by the South African Weather Service (SAWS).

The Group 1 and Group 2 are grouped because of the correlation coefficient of 0.76 between the rainfall stations 0011451 and 0026510. The Group 2 and Group 4 can also be grouped due to the correlation coefficient of 0.65 between the rainfall stations 0024101 and 0025162. The distribution of the groups can be seen in Figure 5.5. Similar rainfall patterns may possess same recharge characteristics such as the recharge period and recharge rate.

Table 5.2 Correlation analyses from 17 long-term records of rainfall stations

Rainfall Station	0007311	0008136	0008782	0011451	0023070	0023218	0023602	0023611	0023706	0024101	0024684	0025162	0025270	0025414	0025599	0026240	0026510	0026824
0007311	1.00	0.88	0.22	0.81	-	0.67	0.31	0.32	0.49	0.64	0.62	0.62	-0.07	0.61	-0.44	0.00	0.27	0.73
0008136		1.00	0.41	0.97	-	0.27	0.24	-0.12	0.14	0.44	0.37	0.35	0.57	0.56	-0.68	0.07	0.21	0.90
0008782			1.00	0.27	0.66	0.64	0.70	-0.66	0.81	-0.20	-0.04	-0.41	0.48	0.10	-0.19	-0.03	-0.45	0.39
0011451				1.00	0.20	-	0.78	0.56	-	-	-	-0.13	-	0.85	0.71	-	0.76	0.72
0023070					1.00	-	0.18	0.39	-	-	-	-	-	0.21	0.36	-	0.25	0.26
0023218						1.00	0.86	0.23	0.77	0.37	0.71	0.24	0.46	0.28	0.27	-0.02	-0.07	-
0023602							1.00	-0.22	-	0.69	0.52	0.25	-	0.58	-0.36	0.29	-0.04	0.55
0023611								1.00	-0.63	0.50	0.42	0.80	-0.08	0.38	0.35	0.47	0.90	0.06
0023706									1.00	-0.21	0.18	0.05	0.41	0.09	0.08	0.29	-0.72	-
0024101										1.00	0.74	0.65	0.22	0.64	-0.07	0.09	0.19	-
0024684											1.00	0.48	0.41	0.63	0.29	0.33	0.26	-
0025162												1.00	0.55	0.57	0.14	0.47	0.71	-0.39
0025270													1.00	0.70	0.05	0.65	-0.27	-
0025414														1.00	-0.20	0.43	0.63	0.58
0025599															1.00	0.45	0.23	0.29
0026240																1.00	0.84	-
0026510																	1.00	0.38
0026824																		1.00

Table 5.3 Groups of rainfall stations in the outcrop of TMG in the study area

Group	Rainfall station	Longitude	Latitude	Correl coef.	Period
Group 1	0007311	19.68	-34.18	0.73 to 0.88	1932-1974
	0008136	20.08	-34.03		1924-1974
	0011451	21.77	-34.02		1968-1989
	0025414	20.73	-33.90		1925-1989
	0026824	21.47	-33.73		1969-1989
Group 2	0023611	19.85	-33.68	0.80 to 0.90	1927-1987
	0026510	21.28	-34.00		1936-1989
	0025162	20.60	-33.70		1920-1974
	0026240	21.13	-34.00		1928-1962
Group 3	0023218	19.63	-33.63	0.70 to 0.90	1937-1950
	0023602	19.85	-33.53		1947-1989
	0008782	20.45	-34.03		1923-1989
	0023706	19.90	-33.77		1920-1946
	0025270	20.65	-34.00		1924-1938
Group 4	0024101	20.07	-33.68	0.74	1931-1965
	0024684	20.38	-33.90		1931-1952
Group 5	0025599	20.83	-33.98	-	1920-1989

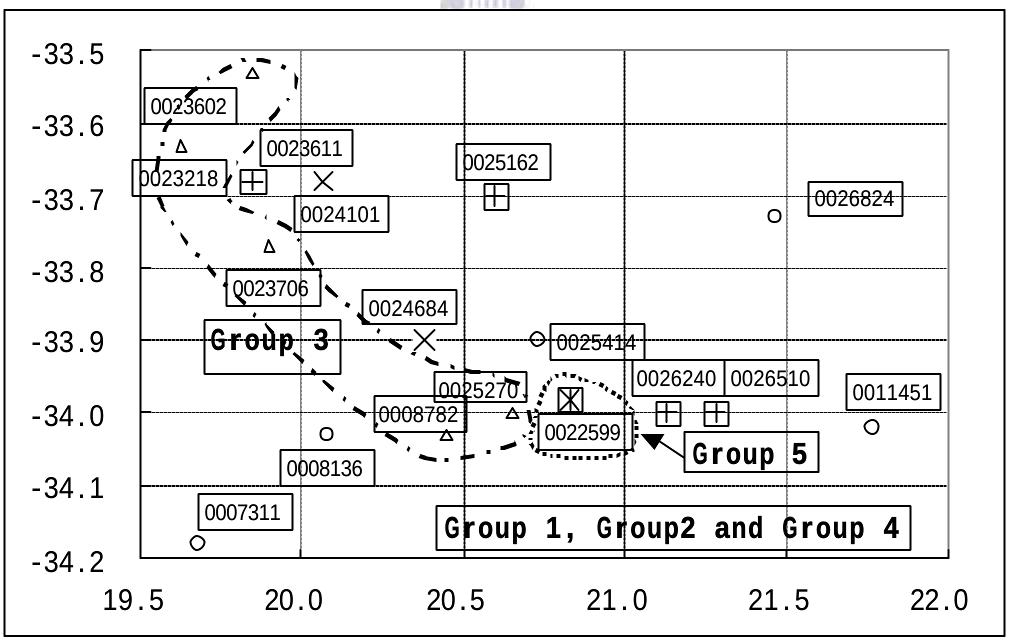


Figure 5.5 Spatial distributions of the groups of rainfall patterns

The Group 1 includes rainfall station of 0007311, 0008136, 0011451,0025414 and 0026824. Cumulate rainfall departure (CRD) analysis is used to evaluate the temporal changes of rainfall. The CRD patterns of the Group 1 are illustrated in Figure 5.6. The following points are observed:

1) The CRD patterns of the Group 1 appear similar pattern in the statistical period of 1925 to 1989.
2) The CRD pattern of the rainfall station 0007311 (orange curve) shows that the CRD increased from 1936 to 1944 and 1953 to 1965. The CRD decreased from 1932 to 1936, 1944 to 1953 and 1966 to 1974. The lowest value of CRD occurred in 1953 and the highest value occurred in 1966 during the statistical period.
3) The decreasing CRD curve occurred from 1924 to 1929, 1940 to 1949 and 1966 to 1979 (pink curve of rainfall station of the 0008136). The CRD increased from 1929 to 1939 and 1949 to 1965. A relevant high rainfall period lasted from 1929 to 1949. The minimum CRD value occurred in 1949 and the maximum occurred in 1965.
4) The CRD pattern of the rainfall station 0025414 (purple curve) shows there were two rainfall periods during statistical period. The first period was about ten years from 1928 to 1948. The CRD increased from 1928 to 1934 then decreased to 1948. The second period was from 1948 to 1973. The period of increasing CRD was from 1948 to 1965. The decreasing period was shorter than increasing period. The CRD increased from 1973 to 1989. The minimum CRD value occurred in 1948 and maximum was in 1989.
5) Similar CRD patterns occurred in the rainfall stations 0011451 (green curve) and the 0026824 (blue curve) from 1969 to 1989. The period of decreasing of CRD was from 1968 to 1979. The period of increasing of CRD was from 1979 to 1989. The lowest value of CRD of the rainfall stations occurred in December 1979, but the maximum occurred in 1986 and 1989.
6) Similar CRD patterns occurred in the rainfall stations, namely 0007311, 0008136 and 0025414 from 1932 to 1974. The CRD increased from 1956 to 1966 and then decreased until 1974. The CRD peaks of the three rainfall stations occurred in October, September and August 1966, respectively. Since 1974, the CRD patterns of the rainfall stations 0007311 and 0008136 were not shown due to the lack of the rainfall records. The CRD patterns of rainfall station 0011451, 0025414 and 0026824 resemble from 1968 to 1989.

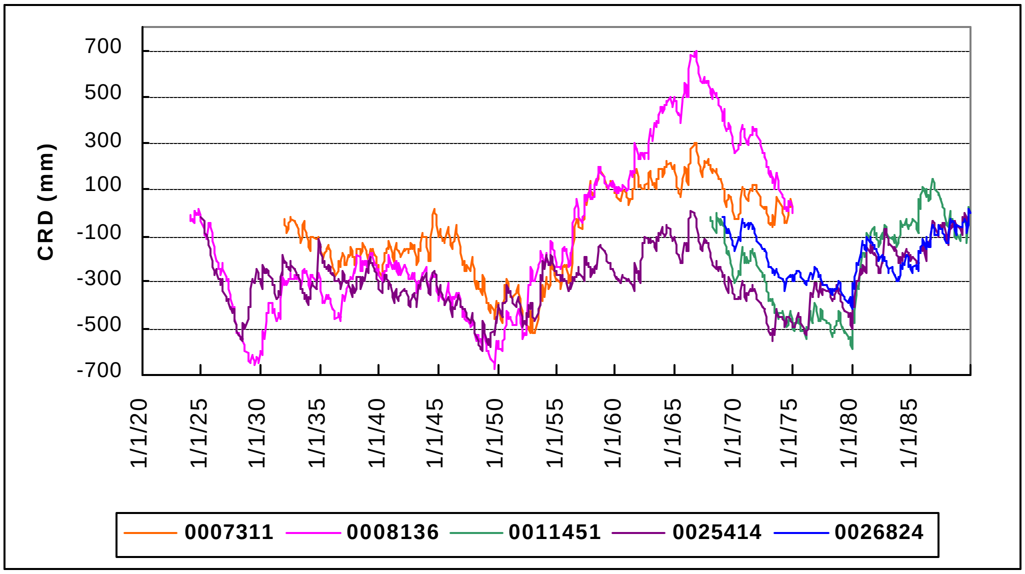


Figure 5.6 CRD patterns of the Group 1

The Group 2 with similar CRD patterns includes the rainfall stations 0023611, 0026510, 0025162 and 0026240 as can be seen in Figure 5.7. The following observations are summarised:

a) The CRD patterns of the rainfall station 0023611 (orange curve) show that the CRD decreased from 1927 to 1952. The CRD increased from 1952 to 1987. The minimum CRD value was in 1952 and the maximum CRD value was in 1987.

b) The CRD patterns of the rainfall station 0026510 (pink curve) show that from 1936 to 1949 the CRD decreased and the minimum CRD value occurred in 1949. The CRD increased from 1949 to 1989. The maximum CRD value was in 1985.

c) The CRD pattern of rainfall station 0026240 (purple curve) shows that there are two rainfall periods during statistic periods. The first period was about twenty years from 1928 to 1949. The CRD increased from 1928 to 1939 then decreased to 1949. The second period was from 1949 to 1961. The period of increasing CRD was from 1949 to 1954. The highest CRD value occurred in 1954 during this period.

d) The CRD patterns of the rainfall station 0025162 (green curve) show that the CRD decreased from 1920 to 1952 and then increased until 1974. The CRD increased from 1949 to 1985. The minimum and maximum of CRD value occurred in 1952 and 1921, respectively.

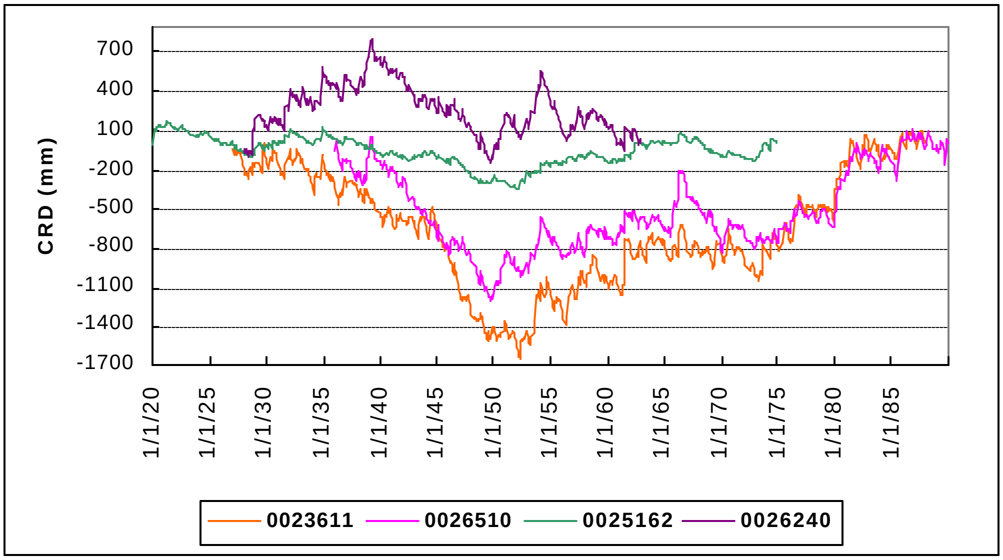


Figure 5.7 CRD patterns of the Group 2

The Group 3 includes 5 rainfall stations with station codes of 0023218, 0023602, 0008782, 0023706 and 0025270. The CRD patterns are shown in Figure 5.8. The rainfall records of rainfall station 0023706, 0023218 and 0025270 are from 1920 to 1946, 1937 to 1950 and 1924 to 1938 respectively. The CRD patterns of the three rainfall stations are not discussed here. The observations are as follows:

- The phases of the CRD value decreasing occurred from 1924 to 1928, 1966 to 1976 and 1981 to 1989. The CRD values increased from 1927 to 1945 and 1979 to 1981. Two rainfall periods occurred in the rainfall station 0008782 (blue curve). The first period was a long rainfall period from 1927 to 1975. The CRD increased from 1928 to 1943. The CRD values fluctuated from 1943 to 1965. The second period were from 1976 to 1989. The CRD peak occurred in 1966 and the lowest CRD occurred in 1968.
- The green curve shows rainfall tendency from 1947 to 1989 of the rainfall station 0023602. The period with increasing CRD was from 1949 to 1964 and 1973 to 1989. The CRD decreased from 1964 to 1973. The lowest and highest CRD value occurred in 1973 and in 1957, respectively.

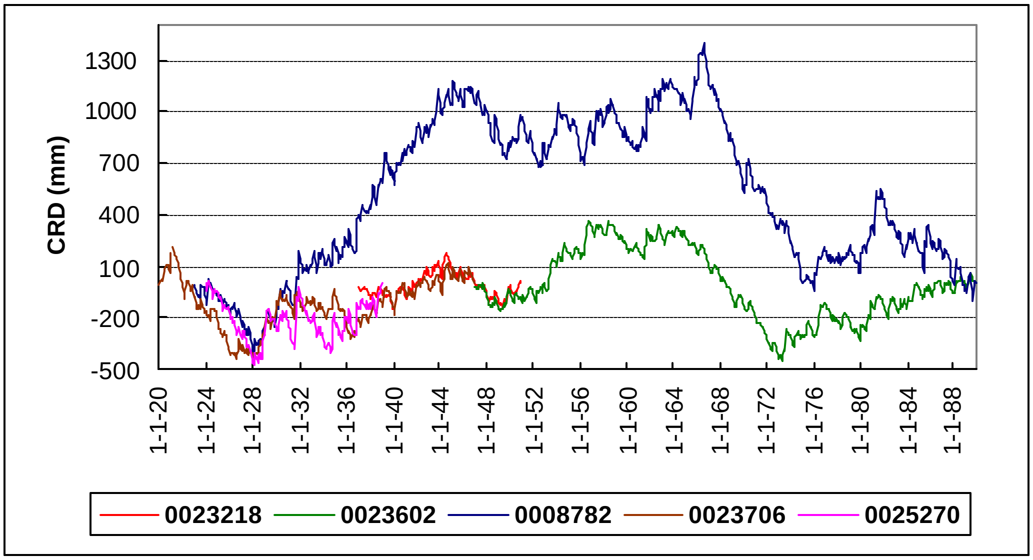


Figure 5.8 CRD patterns of the Group 3

The CRD patterns of the Group 4 show that there were no distinctive rainfall period from 1931 to 1964 and there were a slight variation between the maximum values and the minimum values of CRD at the two rainfall stations of 0024101 and 0024684 (Figure 5.9). It is implied that there was less change in rainfall within this period. The maximum and minimum of the CRD values of the rainfall station 0024684 (red curve) occurred in 1944 and 1948, respectively. From Figure 5.9 can see that the positive CRD values dominate at rainfall station 0024101, whereas, the negative CRD values at the rainfall station 0024684 are predominant. This means that the rainfall was more than the average value for most of the time at rainfall station 0024101; there were 6 extremely low rainfall events, which contributed to the low CRD value. There were five extremely high rainfall events, which form the high CRD values at rainfall station 0024684

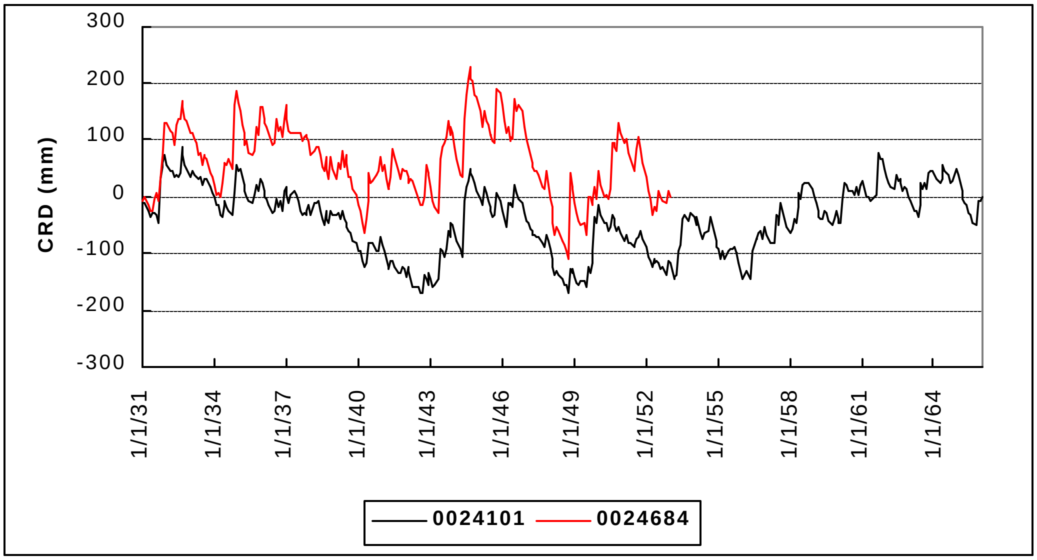


Figure 5.9 CRD pattern of the Group 4

Figure 5.10 illustrates CRD pattern of the Group 5, which includes one the rainfall station of 0025599. It implies that the rainfall period of this rainfall station was distinctly different from the others mentioned above. There was a long rainfall period with CRD increasing from 1920 to 1939 and decreasing from 1939 to 1955. A positive trend started from 1956 although there were some fluctuations. This may indicate the monthly rainfall is larger than the average value in the period.

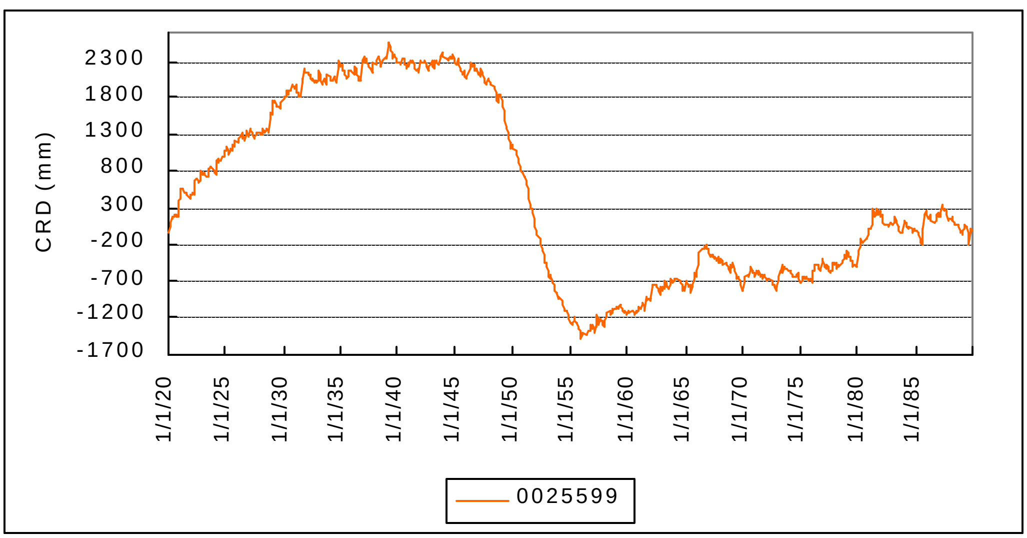


Figure 5.10 The CRD pattern of group 5

5.4 FACTORS RELATED TO EVAPOTRANSPIRATION

5.4.1 Precipitation

The precipitation is a predominant factor for evapotranspiration. Generally, evapotranspiration increases with the increasing precipitation. However, the impacts of precipitation characters on evapotranspiration are complex.

The potential evapotranspiration of three gauge stations distributed in outcrop of the TMG in the study area and rainfall amount near the area are listed in Table 5.4, which shows the potential evapotranspiration decreasing with the precipitation increasing. The result of Table 5.4 implies that the precipitation amount is not the only factor influencing the evapotranspiration. A number of other factors, such as rainfall seasonality, magnitude of rainfall event distribution, and rainfall intensity, as well as temperature have an important effect on the evapotranspiration.

Table 5.4 Relationship between potential evapotranspiration and precipitation
(After Midgley et al., 1994)

Group	Station	Number	Latitude	Longitude	Amount (mm/yr)
Group 1	Rainfall station	0025484	33.57	20.78	104.9
	Evap. station	J1E001	33.18	20.85	2121
Group 2	Rainfall station	0024101	33.41	20.07	169.0
	Evap. station	H3E001	33.70	20.00	1661
Group 3	Rainfall station	0023710	33.83	19.90	290.7
	Evap. station	H4E001	33.83	19.90	1410

5.4.2 Solar radiation

The heating effect of solar radiation increases with the amount of energy intercepted and absorbed. Assuming that moisture is available, thus the evapotranspiration is dependent primarily on the solar energy available to vaporize the water. The factors such as latitude, season of year, time of day and cloud cover that influence direct solar radiation also affect the rate of the evapotranspiration. Cloud cover affects the evapotranspiration by limiting the amount of solar radiation reaching the crop or soil. However, even on cloudless days daily fluctuations in evapotranspiration also occur. On clear days the rate of evapotranspiration increases rapidly in the morning and reaches a maximum usually in early afternoon or mid-afternoon. Latitude and season also influence direct solar radiation and evapotranspiration. So the areas that receive the maximum solar radiation have the greatest evapotranspiration and the areas that receive the minimum solar radiation has the least evapotrans piration.

The amount of solar radiation, which a unit area intercepts at any given time, is linked to the direction it faces (aspect), and to the steepness, or angle of slope (inclination). The study area lies in the southern hemisphere, northwest, facing slope receive more solar radiation than the southeast facing slopes. Therefore north to northwest orientated slopes are warmer than south to southeast slopes in the southern hemisphere (Bonnardot, ARC Infruitec-Nietvoorbij, unpublished; Burger & Deist, 1981) and have more evapotranspiration.

The study area receives more than 2,500 hours of sunshine per year and the annual 24 hour global solar radiation average is about 220 watts per square meter in this area (An energy overview of the Republic of South Africa). The lack of cloud cover in this area increases effective solar radiation.

5.4.3 Temperature

Temperature is usually the most important factor to the evapotranspiration. The evaporation will continue to increase at an increased rate as the temperature rises as long as there is water to evaporate. Although impacted by the greater elevation above sea level of the subcontinent, the temperatures tend to be lower than in other regions in similar latitudes, but the average annual temperature is 17 °C that causes higher evapotranspiration in the study area. However, large monthly variations between the maximum and the minimum temperatures, as well as daily and seasonal temperatures exist for the different regions in the study area resulting in the different actual evapotranspiration. Otherwise, the flat areas with small monthly or daily variations of the temperature have higher evapotranspiration than mountain areas where there are larger monthly and daily variations of temperature. The temperatures used in initial evapotranspiration model are monthly mean temperature, which are adopted from CNC website and listed in Table 5.5.

Table 5.5 Model of temperature in the TMG area (adopted from CNC website)

T ℃	Max	Min	Max	Min	Max	Min	Max	Min	Max	Min	Max	Min
Jan	39	7.5	37	4	36	5	33	5	33	3	32	2.5
Feb	40	6	35	4	36	4	34	5	33	5	31	2
Mar	38	6	35	3	33.5	3	32	4	30	2	28	1.5
Apr	32	2	30	1	30	1	28	2	27	1	26.5	0
May	29	2	26	0	26	-0.5	24	2	24	-1	21	-0.5
Jun	26	0	22	-1	22	0	21	0	20	-4	19	-2.5
Jul	25	-0.5	23	-0.5	22	-2	21	-1	19	-4	19	-2.5
Aug	30	1	27	-1.5	25	-1	25	1	22	-4	21	-3
Sep	32	2	30	0	29	-1	29	1	28	-4	25	-2
Oct	35	3	33	0	32	1	31	3	27	1	28	1
Nov	37	4	33	2	33	3	29	3	30	1	29	-0.5
Dec	38	5	36.5	4	35	4	31	5	31	2	31	0
MAP (mm)	<300		300-400		400-600		600-800		800-1000		>1000	

5.4.4 Wind

Wind speed plays a role in controlling the evapotranspiration rates by influencing the moisture gradient. Research shows that a 8-kilometer-per-hour wind will increase still-air evapotranspiration by 20 percent; a 25-kilometer-per-hour wind will increase still-air evapotranspiration by 50 percent (Chow, 1964). Otherwise influenced by seasonal wind, moisture-laden air is forced to rise over a mountain barrier, producing more rainfall on the windward side than on the leeward side. A wind potential study conducted at three sites by Diab in 1995 in South Africa, found mean annual wind speeds at a height of 10m to vary between 4.0 and 5.0 metres per second (m/s), with wind speeds of 5.5 to 7.0 m/s at a height of 50m, which increases the potential for evapotranspiration.

5.4.5 Relative humidity

Relative humidity and evapotranspiration are closely related because if the relative humidity is close to its holding capacity, the ability of plants to transpire may be inhibited. The higher the relative humidity, the slower the evaporation rate; the drier the air above the surface, the faster the evaporation. So the seasonal trend of evapotranspiration within a given climatic region follows the seasonal trend of solar radiation and air temperature. Minimum evapotranspiration rates generally occur during the coldest months of the year; maximum rates, which generally coincide with the summer season, when water may be in short supply.

5.4.6 Landform

The major landform features in the study area include mountains, valleys, ridges, gorges, etc. The landform impacts on evapotranspiration by their topographic attributes such as altitude, slope and catchment area.

Changes in elevation affect the amount and form of the precipitation, the intensity of storm events and the temperature. The altitude in the Montagu range from 300 m in the northeast above mean sea level to 2,200 m in the southwest above mean sea level. Precipitation generally increases with an increase in elevation. Under natural conditions, however, available moisture is less at the lower elevations because they receive less precipitation than the higher regions. Therefore, the actual amounts of evapotranspiration that occur at the lower elevations will usually decrease than at the higher elevations. At the higher elevations, water is often not limiting, and both evaporation and transpiration are largely determined by the supply of available heat energy (i.e. actual evapotranspiration will increase). Secondly, although solar radiation (direct radiation and diffuse radiation) increases with the elevation, there are definite variations in temperature with altitude and latitude (a decrease of approximately 0.5 °C per 100 m increase in height). Thus, areas at the higher elevations have cooler temperatures and more limited potential to vaporize available water supplies than areas at the lower elevations.

Generally, the slope gradient is classified into three grades in the study area, 10º-20º, 20º-30º and >30º. If the gradient is above 30º, bedrock is exposed and vegetation is sparse. All these factors cause the lower evapotranspiration in steep slopes (slope gradient above 30º). Slope roughness causes friction between the ground and air passing over it and causes atmospheric turbulence that increases evapotranspiration.

5.4.7 Vegetation

Vegetation affects the evapotranspiration in the follow ways:

- Vegetation intercepts water by trapping rainwater on leaves, branches and ground litter before it reaches the ground, evaporating directly back into the atmosphere.
- Vegetation extracts water from soil by the root system and transpired off through the leaves.

The two processes above will increase the evapotranspiration, but the amount of evapotranspiration is related to the percent vegetation cover, density, type, species, and spatial composition and growth stage. The rate of evapotranspiration increases as the vegetative cover increases. Major vegetation types in the study area and the relationship between vegetation and evapotranspiration is shown in Table 5.6. It can be seen from Table 5.6 that vegetation changes can have a significant impact on the amounts intercepted and ultimately on groundwater recharge. Interception are 0.5-5.1%, 0.9-2.6% and 10-20% related to indigenous vegetation (Protea fynbos), grassveld and alien vegetation. Fybons with small-leaved and finely branches have less evapotranspiration than alien vegetation with tall, open canopy. Numerous Herbaceous annuals have shallow roots (<0.3 m) while herbaceous perennials usually have relatively shallow root systems (<1.5 m) which absorb less water from soil to transpire than shrub such as Cape Protea species that have the long taproots (Higgins et al., 1987).

Based on the study of WR90, the factors of vegetation in the outcrop of the TMG area listed in Table 5.7.

Table 5.6 Interception losses for different vegetation types in the study area

(After David et al., 2003)

Vegetation type	Type of estimate	Loss (units)	Source
Protea shrubland fynbos	Measured and modelled	5.1%, rainfall 1500mm	Versfeld (1988)
Indigenous forest Bushveld Fynbos Karoo Grassveld	modelled	3.1-3.5 mm/rainday 1.0-4.4 0.5-2.0 0.2-0.8 0.9-2.6	Schulze (1981)
Pinus radiata plantation	Measured and modelled	Rainfall 1300-1700mm 10.3%-8 years old 12.2%-11 years old 20.0%-29 years old	Versfeld (1988), Pienaar (1964) West Cape
Plantation: Pinus patula Eucalyptus grandis	Measured and modelled	Rainfall 1700mm 10% 5%	Dye (1996a), Mpumalanga
Wattle	Measured	15-20%	Beard (1956)

Table 5.7 Vegetation factors in the outcrop of the TMG (After Midgley et al.1994)

Month	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.
Fynbos factors	0.60	0.55	0.55	0.55	0.45	0.40	0.20	0.35	0.50	0.60	0.60	0.60

5.4.8 Soil

Soil stores water in its pores before water recharges the aquifer system. Water stored in upper layer evaporates directly; it stored in deeper layer is absorbed by vegetation root then transpires to leaves and to be evaporated. The amount of evapotranspiration from soil is controlled by the soil attribute such as soil texture, soil structure and soil moisture content because ability of soil stores water and transports water is different for every soil.

On the one hand, debris and sandy soil have larger pore spaces and favorable structure to move water through the surface into the soil matrix; on the other hand, they have less capillarity and the rate of diffusion of capillary water (only a few cm capillary rise). So sand and debris are cruel to evapotranspiration. Deeper soil has a larger soil moisture reserve than thinner soil, which can supply more water to evaporate.

Soil factors are adopted according to types, texture and depth of soil by referring to the WR90 database. The empirical parameters of soil factors and coverage percentages of vegetation in the TMG area used in the model were according to the filed investigation as listed in Table 5.8.

Table 5.8 Soil factor and coverage percentages of vegetation

Rainfall station	Soil factor	Coverage percentage
0025599	0.15	68-71
0026510	0.2	58-62
0025414	0.26	28-32
0025162	0.43	8-12
0024101	0.29	13-17
0023602	0.25	28-32
0011451	0.2	28-32
0008782	0.1	53-55

5.4.9 Land use

Land use includes nature, civicism, industry and agriculture. Agriculture includes irrigated agriculture and nonirrigated agriculture. The average annual evapotranspiration for irrigated lands varies greatly and, apart from the climatic controls, is dependent on the grass or crop type, quantity of water applied, and length of the growing season. During a drought, natural vegetation may experience moisture stress and wilting, whereas irrigated grasses and crops continue to grow and transpire at a normal rate (if water supplies are available for irrigation). Water is not always readily available for evapotranspiration. Over many rural surfaces there are times, notably on a summer afternoon, when the soil and plants are incapable of moving water to the surface ready for evaporation or transpiration as fast as the atmosphere can do the evaporation.

Bare soil surfaces begin to dry out and plants begin to wilt. Their rate of evapotranspiration will be less than the potential rate.

5.5 FACTORS RELATED TO RUNOFF

Runoff is another important component of water budget. Impact of factors on the runoff is similor to the evaportranspiration in a certain degr ee, but the formation mechanism of the runoff is totally different from that of the evaportranspiration. Runoff amounts vary depending on the features of precipitation and land surface. Runoff occurs when the precipitation rate exceeds the soil's infiltration capacity and increases with precipitation amount increasing.

The rainfall patterns controls formation and distribution of runoff. In the extremely west study area, relative more rainfall occurs exclusively in May, June, July and August resulting in larger proportion of runoff in a year. In the east study area, rainfall distributes all year round but concentrates in March, August, October and November, therefore runoff may be mainly related to these months. Notice that very little rainfall occurs in most study areas but mountainous areas and runoff may greatly contribute by mountainous catchments and occasional high rainfall events.

Runoff is reduced on a concave slope, but increased on a convex slope due to the gradient of the steepest portion. As the gradient increases, the kinetic energy of rainfall remains constant, but transport accelerates toward the foot as the kinetic energy of the runoff increases and outweighs the kinetic energy of the rainfall when the slope exceeds 15% (Roose, 1992). The roughness increases surface storage and promotes greater infiltration, at the same time, rough slope develops friction between water and slope surface and extends time of concentration. The common convex slopes help the runoff, but lots of crannies and debris on the outcrop of the TMG decreases the runoff.

Vegetation plays a role of buffer of water to the ground surface by interception and the stream flow withholding from the initial period of runoff. Vegetation has a significant impact on infiltration both by providing canopy and litter cover to protect the soil surface from raindrop impacts and by producing organic matter, which binds soil particles and increases its porosity. Higher porosity increases infiltration and percolation rates and the water-holding capacity of the soil (Valentini et al., 1991; Dawson, 1993) and decreases runoff. Vegetation, including its ground litter, forms numerous barriers along the path of the water flowing over the surface of the land. This increases surface roughness and causes water to flow more slowly particularly on gentle slopes, giving the water more time to infiltrate and to evaporate.

Infiltration capacity is dependent on the porosity of a soil, which determines the water storage capacity and affects the resistance of water to flow into deeper layers. Porosity differs from one soil type to the other. A sandy soil with a high porosity will “accept” water more readily enabling a more rapid rate of infiltration. The rate of infiltration decreases as the degree of saturation increases until it reaches a steady rate at saturation. The sandy soil formed from weathering derive in most outcrop of the TMG in the study area has larger infiltration capacities. Some areas of the catchment tend to contribute more towards runoff than others due to factors such as variations in soil water retention properties, surface drainage and the accumulation of moisture in low lying areas.

Kinetic energy of raindrops in a high intensity storm causes a breakdown of the soil aggregate as well as soil dispersion with the consequence of driving fine soil particles into the upper soil pores. This results in clogging of the pores, formation of a thin but dense and compacted layer at the surface, which highly reduces the infiltration capacity. This effect, often referred to as capping, crusting or sealing, explains why in arid and semi-arid areas where rainstorms with high intensities are frequent, considerable quantities of surface runoff are observed even when the rainfall duration is short and the rainfall depth is comparatively small. On coarse, sandy soils the capping effect is comparatively small. Therefore, soils with high infiltration rates have low runoff potential than with low infiltration.

A study in sandy soil of Zimbabwe showed that relative infiltration rate varied form 55% to 84% to 100%, increasing with open grassland to closed canopy to open canopy (Kennard and Walker, 1973). Infiltration rates are positively related to litter and grass basal cover, being up to 9 times faster with 100% litter cover than for bare soil (O’Connor, 1985). One the study found that replacement of deep-rooted eucalypt forest with shallow-rooted grassland reduced infiltration rates, decreased saturated hydraulic conductivity 10-fold and sorbtivity 3-fold (Sharma et al., 1987b).

To a certain catchment, the drainage shape and channels are more important than the drainage area. Long and narrow catchments have longer times of concentration resulting in lower runoff-rates than more square watersheds of similar size, which have a number of tributaries discharging into the main channel.

Based on the above discussion, the runoff is a function of Curve Number (CN) depending on the soil water content (moisture condition), feature of land and different land use description. It is gained from the U.S. Department of Agriculture and Natural Resources Conservation Service (NRCS), formerly known as the Soil Conservation Service (SCS). The CN number varies from site conditions, soil type and management conditions and ranges from 36 to 99 (USDA-SCS, 1985). Considering the features of the soil and nature of land in the outcrop of the TMG, the ranges of CN are adopted from 88 to 98.

5.6 SUMMARY

The amount of precipitation and rainy season varies greatly in the different zones. Especially, in the outcrop of the TMG, the variation of precipitation is dramatic because of the influencing of climate and topography. The average annual precipitation increases from 200.0 mm to 300.0 mm in the western to 450.0 mm in the eastern of the study area. The mean annual precipitation is more than 500.0 mm along the Matroosberg in the west but less than 200.0 mm/yr has been recorded along the inland foothills, fringing the Little Karoo. In the south along the Langeberg Mountain, interior foothills receive as little as 200.0 mm and upper slopes as much as 1,000.0 mm/yr. There are more rainfall in mountainous area where is the outcrop of the TMG than that in the flat area or Little Karoo.

According to correlation analysis from the 17 rainfall stations with long-term records, the five groups were identified The different rainfall patterns of the five groups show that the rainfall patterns of the study area vary greatly although the study area lies in the same homogeneous rainfall districts. The similar patterns of CRD are observed in the different groups of the rainfall stations. The effective recharge may mainly occur from June to August in winter rainfall areas or March to May and August to November duo to bimodal rainfall pattern in all year round rainfall district, but the highly single rainfall event may contribute to effective recharge in semi-arid area.

The precipitation is a predominant factor because it is the source of the evapotranspiration and the runoff. The factors impacting on the precipitation will impact on the evapotranspiration and the runoff. Generally, the evapotranspiration and runoff increases with the precipitation amount increasing. The precipitation characteristics such as precipitation type (rainfall, snow, sleet, etc.), intensity, duration and distribution influences the evapotranspiration and the runoff as well; this would results in highly potential evapotranspiration but lowly actual evapotranspiration. In addition soil, vegetation, altitude, slope and catchment area affect the evapotranspiration and the runoff. The amount of evapotranspiration and runoff varies dramatically when influenced by precipitation characters associated with other factors. Thus all the factors influence the precipitation, the evapotranspiration and the runoff directly, and then influence recharge indirectly. The recharge increases with precipitation increasing; it decreases with increasing evapotranspiration and runoff.

CHAPTER 6

ANALYSIS OF RESULTS

6.1 OUTLINE OF OUTCOMES

Recharge is considered as the residual between precipitation and direct runoff and actual evapotranspirtation at year-round scale in this study. Actual evapotranspiration estimates are based on the temperature and the characteristics of soils and vegetation. The data are selected before they are used in the calculation by comparing the background conditions, such as the homogeneous rainfall districts, the catchment size and the potential evaporation. Soil factors are adopted according to types, texture and depth of the soil by referring to the WR90 database. The factors of vegetation in the outcrop of the TMG area are listed in Table 5.5. The empirical parameters of soil factors and coverage percentages of vegetation in the TMG area used in the model are according to the filed investigation as listed in Table 5.7. The temperatures used in the initial evapotranspiration model are monthly mean temperature, which are adopted from the CNC website and listed in Table 5.8. For runoff estimation, considering the features of the soil and nature of land in the outcrop of the TMG, the ranges of the CN number are adopted from 88 to 98.

Outline of estimates of precipitation, runoff, actual evapotranspiration and recharge of the rainfall stations are listed in Table 6.1 (for details see Appendix 5 to Appendix 12). The runoff levels vary from 0.0 to 219.2 mm/yr, and average values of runoff ranges from 0.7 to 137.4 mm/yr. The actual evapotranspiration calculated ranges from 40.2 mm/yr to 1261.0 mm/yr. Recharge rates are different from station to station. The minimum recharge rates range from 0.1 mm/yr related to 40.5 mm/yr of precipitation at rainfall station 0024101 to 3.3 mm/yr related to 399.0 mm/yr of precipitation at rainfall station 0008782. The maximum recharge rates are from 2.9 mm/yr related to 266.3 mm/yr rainfall at rainfall station 0025162 to 38.0 mm/yr related to 682.8 mm/yr of rainfall at rainfall station 0008782. The average recharge rates range from 1.0 mm/yr related to 132.4mm/yr of rainfall at rainfall station 0025162 to 18.9 mm/yr related to 682.8 mm/yr of precipitation at rainfall station 0008782. Stations with higher precipitation values display a large range of precipitation and vice versa as can be seen in Figure 6.1. So do the recharge rates (Figure 6.2).

Table 6.1 Outline of ranges of precipitation, runoff, evapotranspiration and recharge in the study area

Rainfall station	Item	Precipitation	Runoff	Evap	Re	RE %	Period used
0008782	Min	399.0	11.05	383.23	3.26	0.67	1947-1989
	Max	947.9	112.04	819.84	37.98	5.38	
	Median	652.9	43.88	590.96	18.67	2.75	
	Mean	660.4	49.99	591.23	19.19	2.89	
	Stdev	133.7	29.78	105.78	7.40	1.04	
0011451	Min	197.3	0.00	195.63	0.99	0.50	1968-1989
	Max	828.6	99.50	707.03	24.01	4.65	
	Median	400.5	9.03	384.71	9.23	2.30	
	Mean	422.6	19.30	392.35	10.93	2.41	
	Stdev	134.4	24.62	107.23	6.73	1.02	
0023602	Min	144.1	0.00	142.87	0.70	0.27	1947-1989
	Max	489.8	24.78	458.23	6.79	1.69	
	Median	262.2	2.96	257.91	2.35	0.85	
	Mean	274.2	5.42	266.23	2.60	0.91	
	Stdev	81.5	5.55	75.95	1.45	0.33	
0024101	Min	101.9	0.00	101.03	0.32	0.24	1931-1965
	Max	251.8	9.52	246.07	3.50	1.53	
	Median	150.9	0.09	148.40	1.22	0.85	
	Mean	169.0	1.48	166.10	1.40	0.81	
	Stdev	47.5	2.52	45.45	0.64	0.20	
0025162	Min	40.5	0.00	40.17	0.09	0.11	1936-1974
	Max	265.3	9.19	259.76	2.90	1.20	
	Median	128.8	0.00	125.38	0.98	0.81	
	Mean	131.5	0.75	129.75	0.99	0.74	
	Stdev	50.0	1.72	48.32	0.55	0.18	
0025414	Min	117.0	0.00	116.27	0.69	0.44	1968-1989
	Max	563.7	24.86	533.25	5.59	1.04	
	Median	282.8	4.18	279.21	1.94	0.66	
	Mean	288.8	7.22	279.49	2.11	0.69	
	Stdev	109.3	7.53	102.17	1.23	0.18	
0026510	Min	383.0	5.02	363.77	1.76	0.46	1936-1974
	Max	965.3	113.13	836.02	22.05	3.40	
	Median	593.5	35.38	543.12	15.76	2.46	
	Mean	618.6	40.63	563.41	14.59	2.39	
	Stdev	131.7	24.73	107.74	4.09	0.64	
0025599	Min	298.6	5.48	291.72	1.40	0.45	1920-1989
	Max	1531.7	255.93	1261.05	28.22	2.42	
	Median	1021.8	145.96	870.05	15.74	1.59	
	Mean	1001.3	140.56	844.67	16.03	1.60	
	Stdev	232.5	61.42	170.79	5.39	0.40	

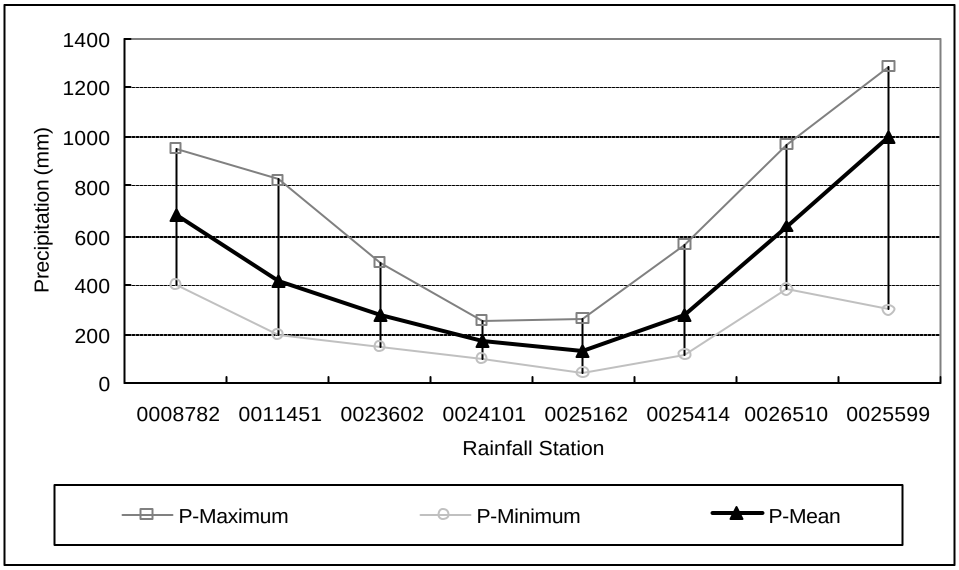


Figure 6.1 Outline of rainfall of the rainfall stations

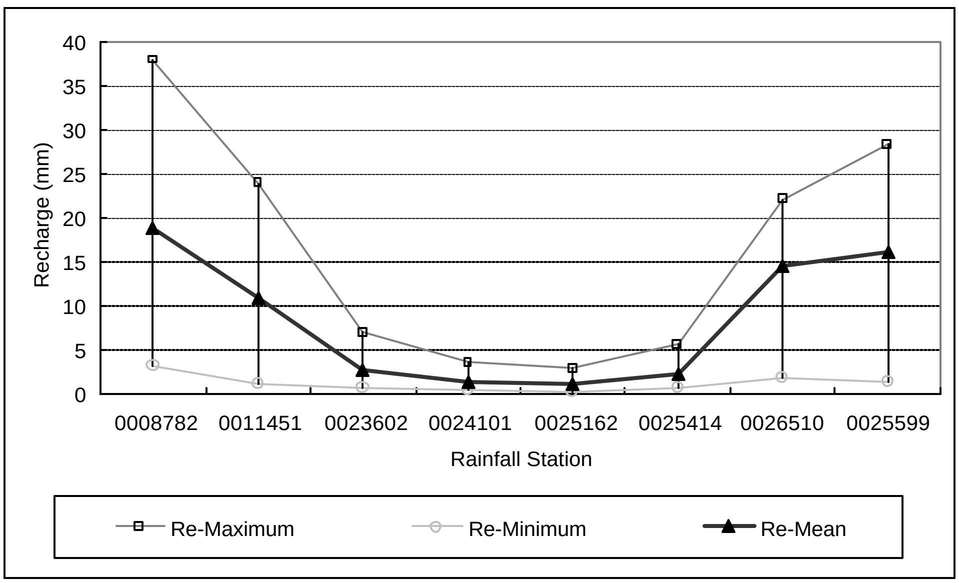


Figure 62 Outline of the recharge estimates of the rainfall stations

6.2 RUNOFF

The estimated runoff based on eight rainfall stations shows that the runoff order varies from 0.0 to 219.2 mm/yr, and average values of runoff ranges from 0.7 to 140.6 mm/yr as can be seen in Table 6.1. Values of runoff more than 100.0 mm/yr occur in the period with rainfall more than 759 mm/yr. Runoff more than 200.0 mm/yr occurs in the precipitation more than 1,000 mm/yr. The average runoff is 46.2 mm/yr in the study area. The variability of the runoff can be primarily attributed to the pattern of precipitation, especially rainfall event. According to the flow of the 15 stream gauges located in the study area, the average annual flux is 441.6×10^6 m^3, which equals to 59.1mm/yr within drainage area of 7473 km^2. This implies the direct runoff makes up approximate 78.2% of the stream flow. It should be noted that the runoff in the outcrops of the TMG is not accurate obviously because that the factors of drainage area and effective radius of rainfall station are not taken into account. The detailed calculated results of the runoff of each rainfall station and stream gauge are presented in Appendix 5 to Appendix 12 and Appendix 15.

The patterns of runoff related to precipitation are shown in Figure 6.3. It is discussed by rainfall ranging from 0-400 mm/yr, 400-1000 mm/yr and higher than 1000 mm/yr. The following points are addressed:

- Runoff increases with increasing precipitation.
- There is different runoff in the different rainfall stations. The runoff changes with characteristics of soils.
- Runoff may not occur if precipitation is less than 350.0 mm/yr. Values of less than 100.0 mm/yr occur in wide areas and can be primarily attributed to the precipitation patterns in these areas where runoff is related to high rainfall events instead of total precipitation in this case. However, the contributions of a few relative high rainfall events to runoff are finite.
- Runoff increases with the total precipitation increasing if precipitations exceed 400.0 mm/yr. Runoff ranging from 50.0 to 150.0 mm/yr are common in the study area. There is more runoff if the precipitation exceeds 1000 mm/yr.
- Percentage of runoff (the ratio of runoff to precipitation) increases with increasing precipitation (Figure 6.4). From statistical analysis point of view, there are linear relationship with formula as $y = 0.015x-2.262$ with correlation coefficient 0.91 between the runoff and the precipitation.
- Percentage of runoff less than 4 coincides with precipitation below 400 mm/yr. The periods with the runoff percentage ranging from 3-14% are related to the precipitation ranging from 400-1,000 mm/yr. The percentage of runoff more than 14% is related to the precipitation higher than 1,000 mm/yr.

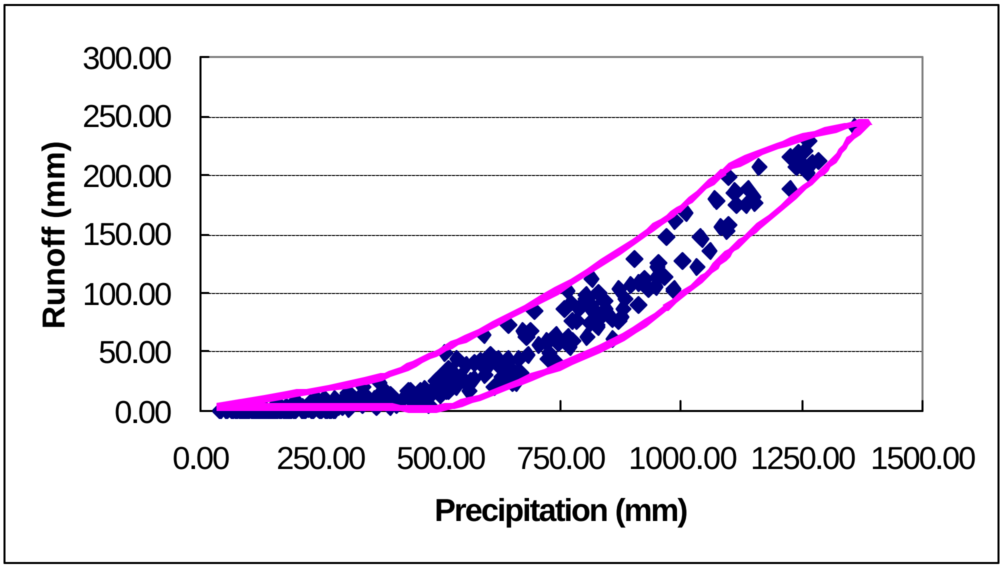


Figure 6.3 Runoff pattern in the study area

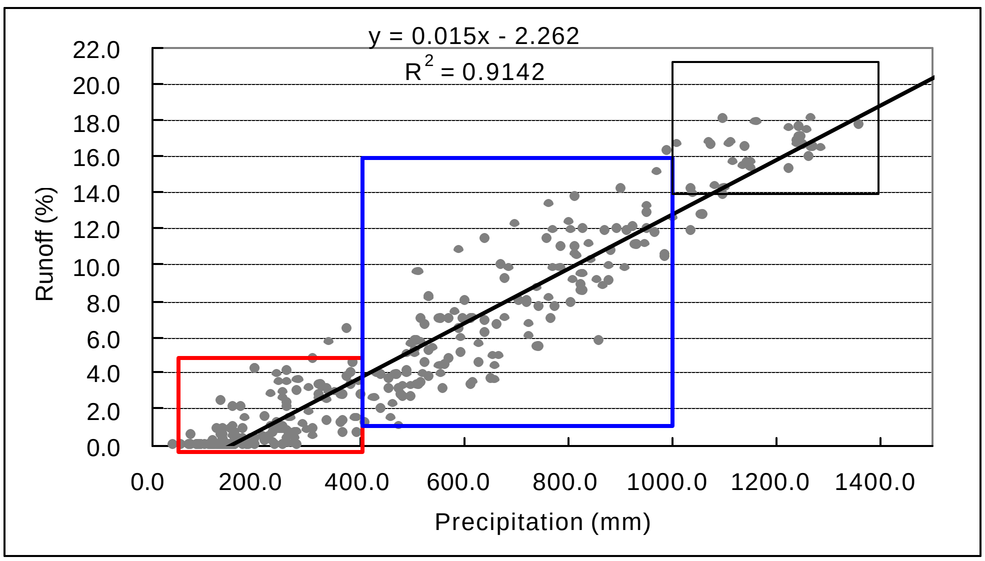


Figure 6.4 Runoff (%) pattern in the study area

6.3 ACTUAL EVAPOTRANSPIRATION

The statistical results of calculated actual annual evapotranspiration are presented in Table 6.1 (For details refer to Appendix 5 to Appendix 12). The actual evapotranspiration calculated ranges from 40.2 mm/yr to 1261.0 mm/yr. The average annual actual evapotranspiration is from 129.8 to 458.9 mm. The values of actual evapotranspiration increase with increasing precipitation as can be seen in Figure 6.5. A reverse correlation between evaportranspiration percentage and the precipitation occurs (Figure 6.6). The percentages of evaportranspiration are 94-99%, 84-96% and less than 84% related to rainfall within 400.0 mm/yr, 400.0-900.0 mm/yr and higher than 900.0 mm/yr, respectively (Table 6. 2). In other words, small rainfall events mainly contribute to evaportranspiration, but large precipitation events produce more runoff and evaportranspiration.

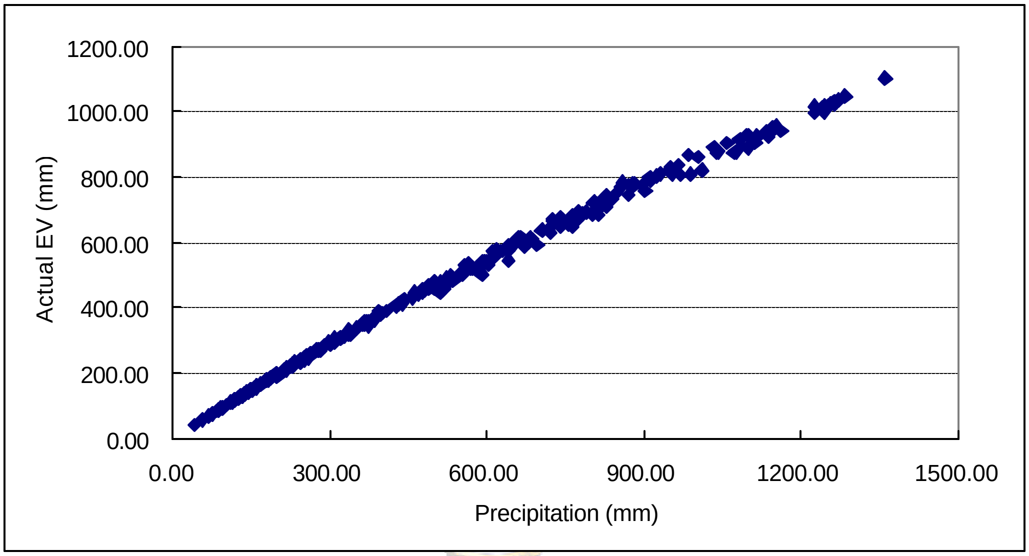


Figure 6.5 Pattern of total evaportranspiration versus precipitation

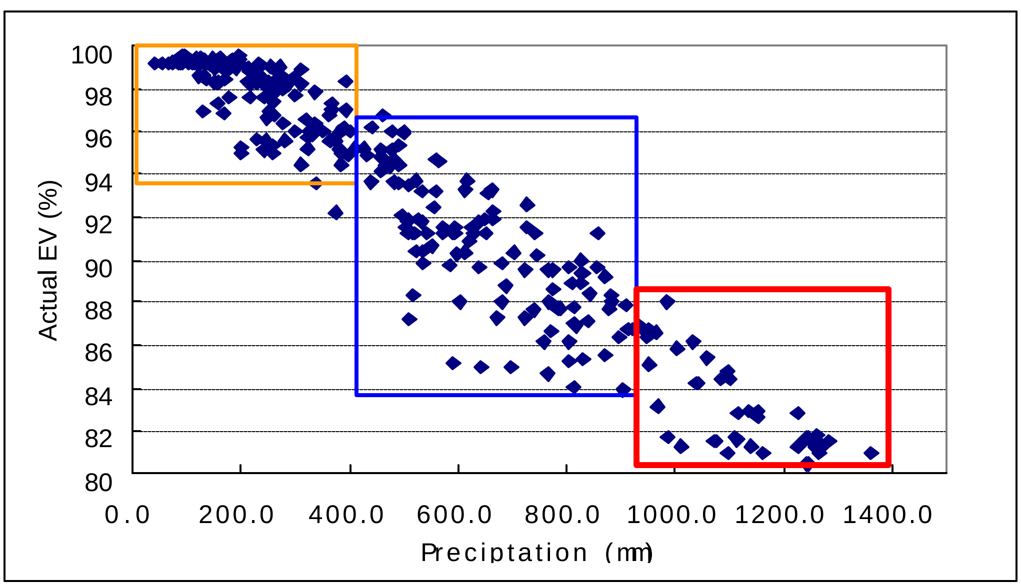


Figure 6.6 Scatterpatterns of evaportranspiration ratio versus precipitation

Table 6.2 Relationship between percentage of actual evapotranspiration and precipitation

EV (%)	Precipitation (mm/yr)
94-99	<400
96-84	400-900
<84	>900

6.4 RECHARGE

6.4.1 Spatial distribution of recharge rate in the study area

Recharge rates in the study area, in terms of rainfall stations, are presented in Appendix 5 to 12. The spatial distribution of average annual recharge rates using the Kriging's method in the study area is shown in Figure 6.7. It is important to note that the ranges of recharge have been exaggerated due to the finite rainfall stations in the outcrop of the TMG in the study area.

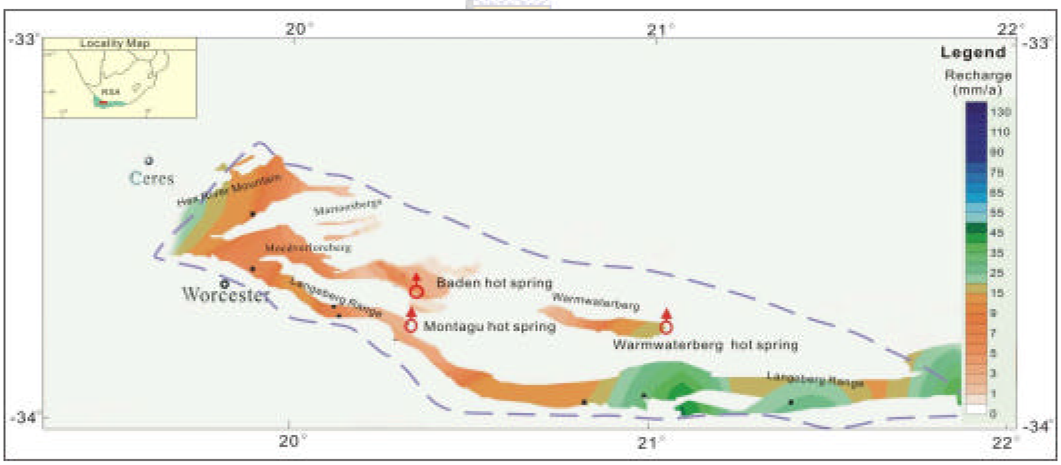


Figure 6.7 Annual average recharge rates in the outcrop of the TMG in the Montagu area

The following observations are obtained from Figure 6.7:

1) The recharge rates vary greatly in the study area. The recharge rates range from 0.5 mm/yr to 40.0 mm/yr. Most of the areas recharge rates are less than 15.0 mm/yr. Average recharge rate is 12.5 mm/yr. The highest recharge rate is 38.0 mm/year at rainfall station 0008782; the lowest recharge rate is 0.1 mm/yr.

2) The recharge in the Langeberg Ranges seems to have two type trends: the western section is from Montagu to the margin of the Hex River Mountain; the eastern section is from Montagu to the Garcia Nature Reserve. The recharge increases from 3.0 mm/yr to 25.0 mm/yr in the western section. In the east section, the recharge rates are from 3.0 mm/yr to 40.0 mm/yr. Two high recharge areas are distributed in the Grootvadersbosch Nature reserve and Garcia Nature Reserve in the eastern section
3) The recharge rates from west to east of the Warmwaterberg are from 1.0 mm/yr to 15.0 mm/yr.
4) The recharge rates in the Moedverloreberg located in the northern Koo Valley and Kiser Valley are from 1.0 mm/yr to 10.0 mm/yr. From east to west of Kwadousberg, the recharge rates are from 4.0 mm/yr to 9.0 mm/yr. It is worth mentioning that the recharge in the north mountainous areas of the Baden hot spring is from 1.0-7.0 mm/yr.
5) The recharge rates in the Martoosbergs are from 1.0 mm/yr to 3.0 mm/yr, which are related to the Great Karoo climate.
6) The results are generally less than that from Vegter 1995. The recharge rates range from 12.0 to 40.0 mm/yr as can be seen in Figure 6.8 (Vegter 1995), in which recharge rates more than 20.0 mm/yr are predominant in the study area. Figure 6.8 shows there are similar recharge rates in most areas, where the rainfall patterns, topography and climate may greatly change.

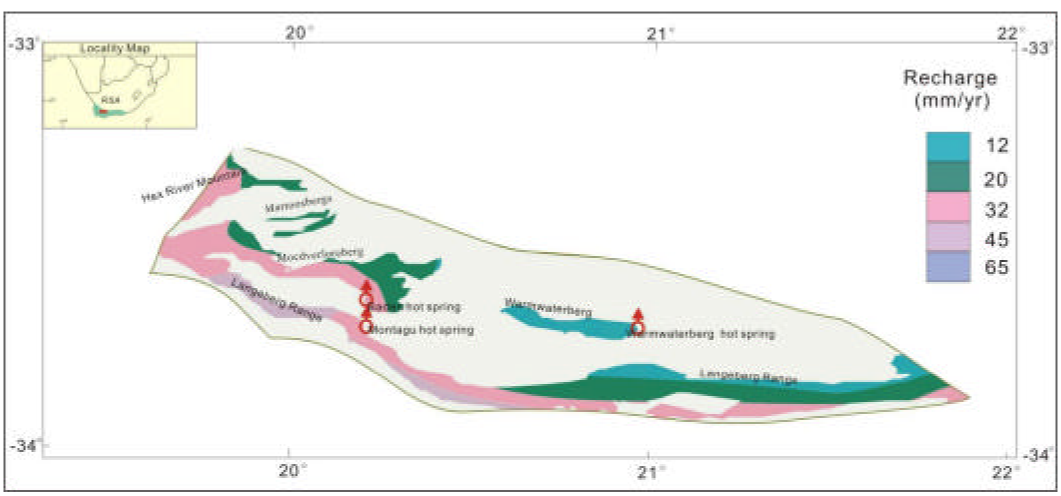


Figure 6.8 Annual average recharge rates in the TMG area (After Vegter, 1995)

6.4.2 Relationship between the recharge and the precipitation

A correlation diagram was used to discuss the relationship between the recharge and the precipitation. Figure 6.9 shows the distribution of the recharge rate versus the precipitation. The following points are observed:

1) The recharge values are limited in a narrow strip (pink box). The recharge rates range from 0.1 mm/yr to 38.0 mm/yr in the study area and values less than 20.0 mm/yr is predominant.
2) Relatively low recharge rates coincide with low precipitation in most areas. The recharge rate is less than 5.0 mm/yr if the precipitation is below 400 mm/yr. The ranges of 10.0-20.0 mm/yr of recharge rates occur between 600-1,200 mm/yr of precipitation. There are relatively low recharge values if the precipitation exceeds 800 mm/yr, which can be primarily attributed to more runoff and evapotranspiration.
3) The recharge values higher than 20.0 mm/yr are more related to the precipitation higher than 800 mm/yr.
4) The recharge rates are grouped clearly, which is group A (green Box) and group B (blue box) as shown in Figure 6.9. In group A, the recharge rates are below 5.0 mm/yr. Most of the recharge rates are less than 2.0 mm/yr when the precipitation is less than 200 mm/yr (Figure 6.10). A large recharge range occurs when the precipitation exceeds 400 mm/yr.
5) The low recharge rates below 2.0 mm/yr may be related to extreme high rainfall events in semi-arid area. This means that separate high rainfall events contribute to groundwater recharge (periodic recharge).
6) Table 6.3 lists outline of recharge rates in the study area. The recharge values less than 20.0 mm/yr is 87.25% in the most years. Hereinto, recharge rates less than 10.0 mm/yr are 52.7% of all statistical months within the rainfall stations.

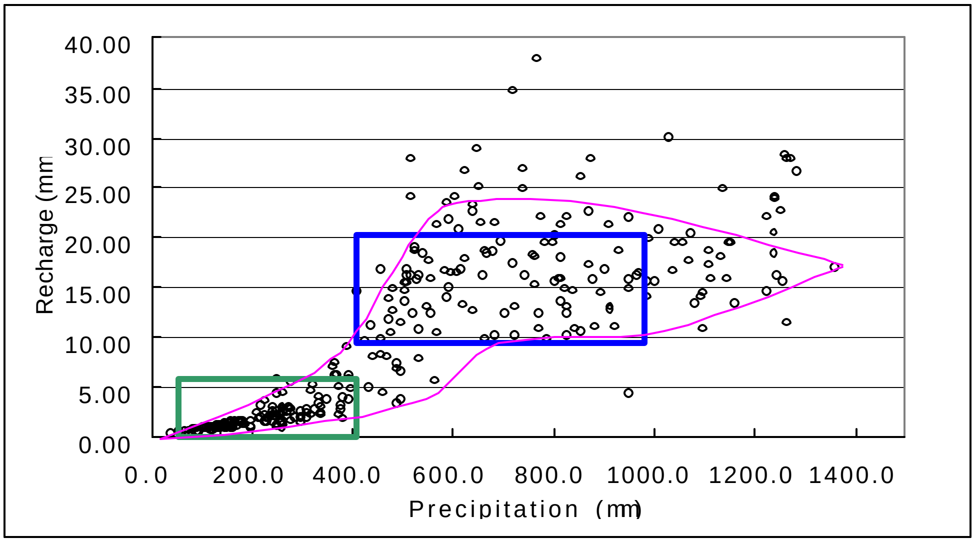


Figure 6.9 Scatter diagram of recharge rate versus precipitation

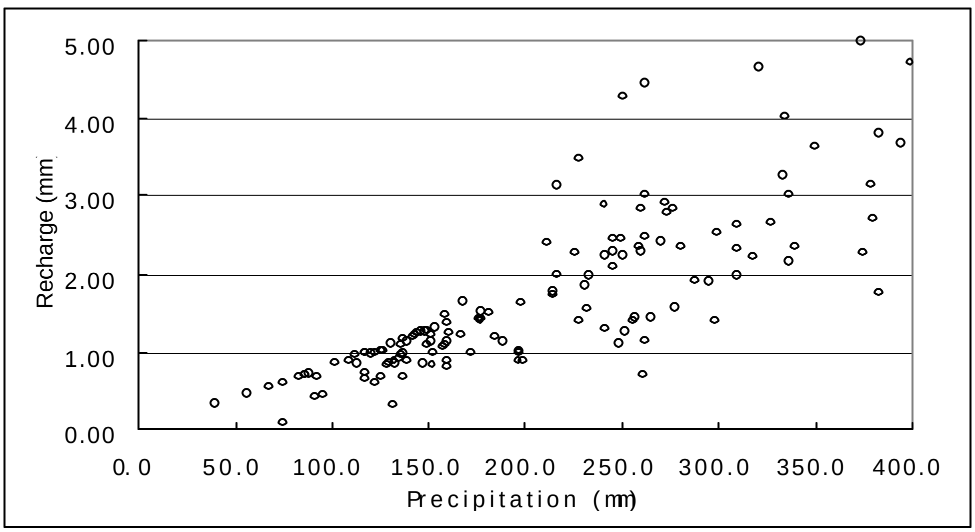


Figure 6.10 Relationship between recharge and precipitation within 400 mm/yr

Table 6.3 Outline of ranges of recharge in the study area

Recharge ratio (mm)	<1.0	1.0-2.0	2.0-3.0	3.0-6.0	6.0-10.0	10.0-20.0	>20.0
Size	34	60	30	19	22	108	40
Frequency (%)	10.9	19.2	9.6	6.1	7.0	34.5	12.8
Range of P (mm)	40.5-261.7	113.0-383.0	211.7-380.3	216.9-498.9	251.5-788.4	408.7-1531.7	516.0-1284.5
Range of Runoff (mm)	0-8.52	0-17.47	0-19.4	0-24.22	0.03-86.75	6.02-256.93	17.57-219.24
Range of Evap (mm)	40.2-258.8	112.0-363.8	208.2-364.8	213.0-478.8	246.5-691.8	388.9-1261.1	456.8-1046.4

For further discussion of the relationship between the recharge and the precipitation, a diagram of the recharge percentage versus the precipitation is shown in Figure 6.11. The following points are obtained:

1) The recharge percentages vary with the precipitation. Three groupings are observed. There is lower recharge percentage less than 2% in group 1 (red box); the recharge rates are from 1-5% in group 2 (blue box); in group 3 (orange box) the recharge rates are from 1-3%.
2) The recharge percentages ranges from 1% to 2% in group 1 coincide with low precipitation (< 400 mm/yr) in these areas. However, the recharge percentages are less than 1% if rainfall is less than 200 mm/yr. Recharge rates increases with increasing rainfall if the rainfall exceed a threshold value of 200 mm/yr (Figure 6.12).
3) The recharge percentage ranges of 1-5% are related to the precipitation from 400 to 800 mm/yr. Most higher recharge percentages are related to ranges of 450-800 mm/yr of rainfall.
4) Relatively low recharge percentages (< 3%) occur if precipitation exceeds 800 mm/yr.
5) The relationship between recharge percentage and the precipitation is not linear.
6) The outcrop of the TMG with increased precipitation along the Langeberg Mountain Range and the Hex River Mountain form recharge areas in the study area.

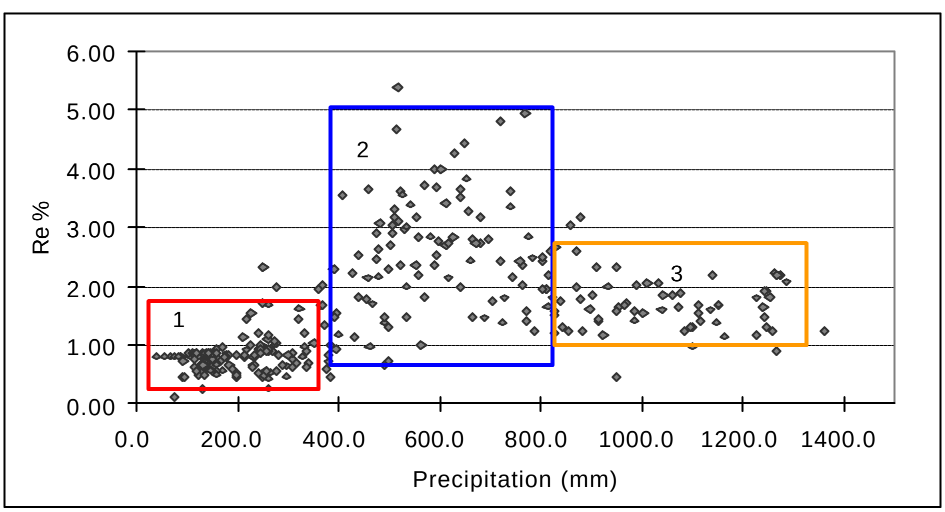


Figure 6.11 Recharge rates (%) versus precipitation

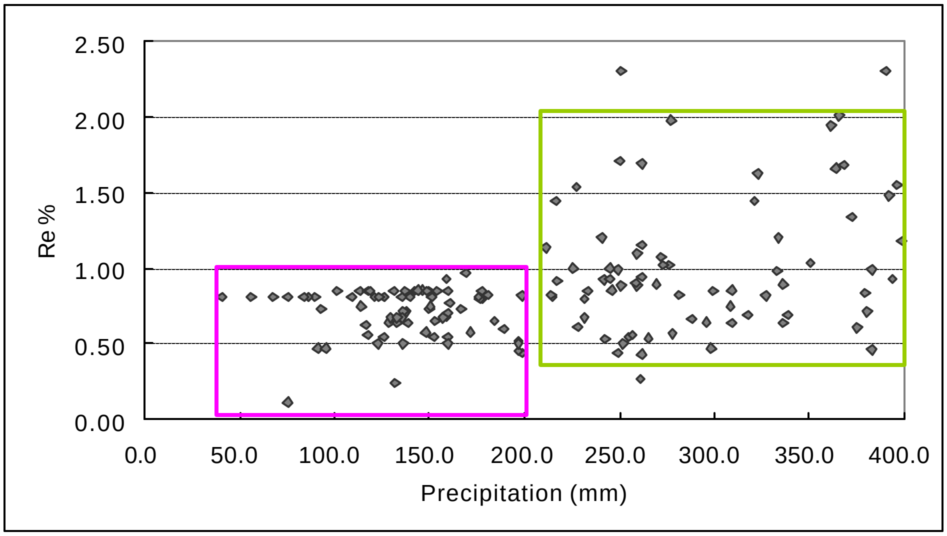


Figure 6.12 Recharge rates (%) versus precipitation within 400 mm/yr

The relationship between annual precipitation, runoff, actual evapotranspiration and recharge of typical rainfall stations are shown in Figure 6.13 to Figure 6.20. The results imply there are good agreement between the precipitation and the recharge, and the runoff and the actual evapotranspiration in the study area.

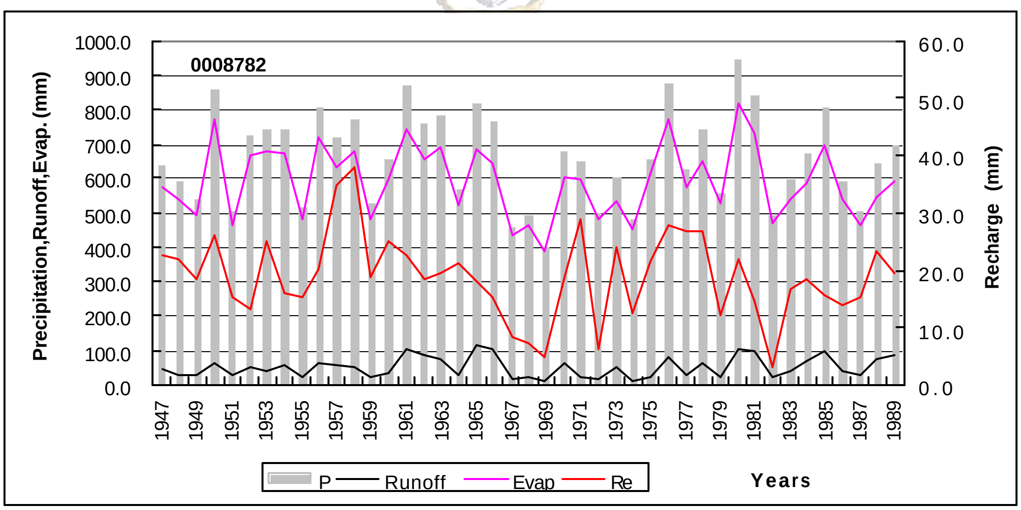


Figure 6.13 Relationship between precipitation, runoff, actual evapotranspiration and recharge for the rainfall station 0008782

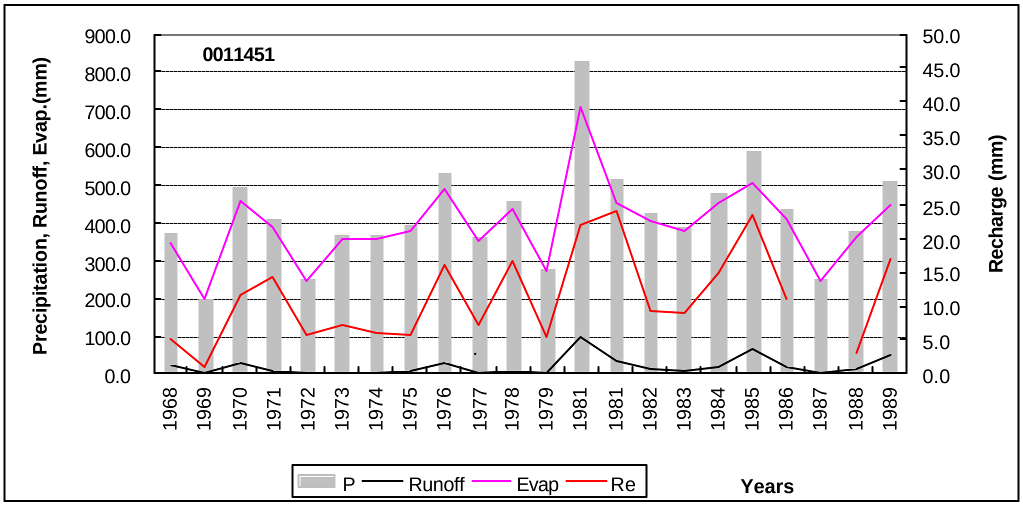


Figure 6.14 Relationship between precipitation, runoff, actual evapotranspiration and recharge for the rainfall station 0011451

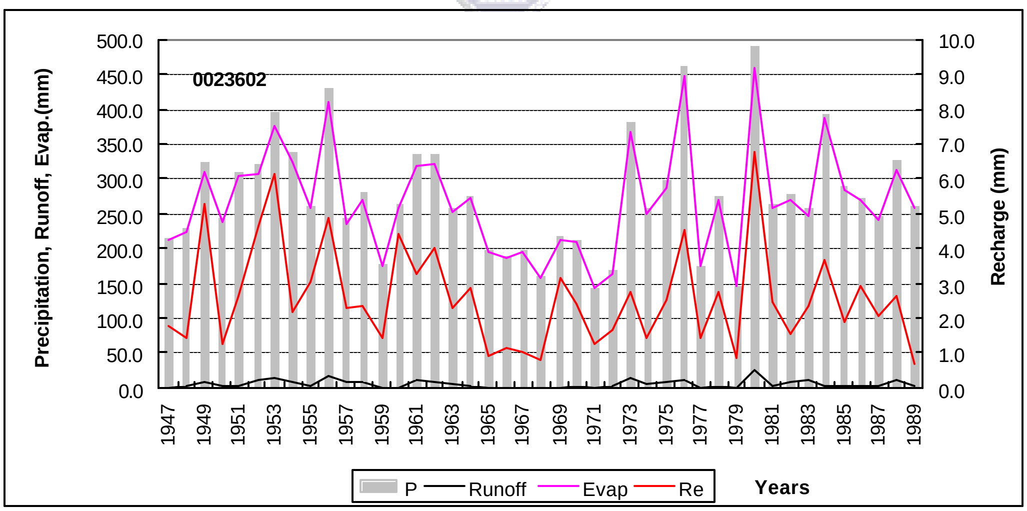


Figure 6.15 Relationship between precipitation, runoff, actual evapotranspiration and recharge for the rainfall station 0023602

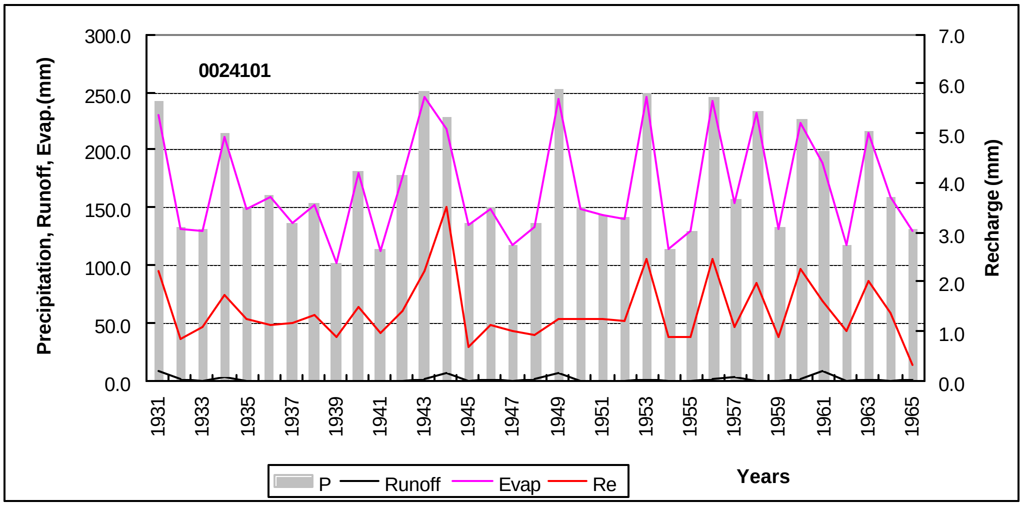


Figure 6.16 Relationship between precipitation, runoff, actual evapotranspiration and recharge for the rainfall station 0024101

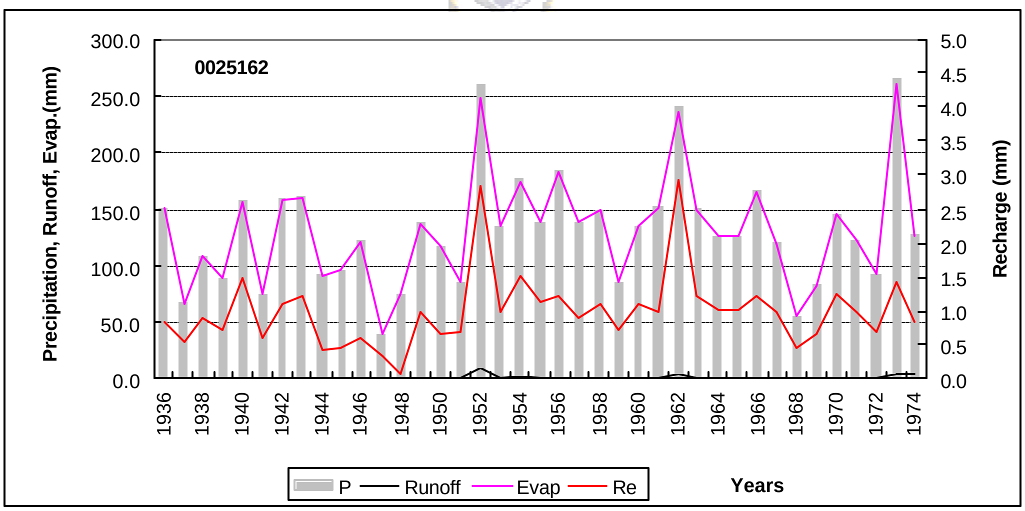


Figure 6.17 Relationship between precipitation, runoff, actual evapotranspiration and recharge for the rainfall station 0025162

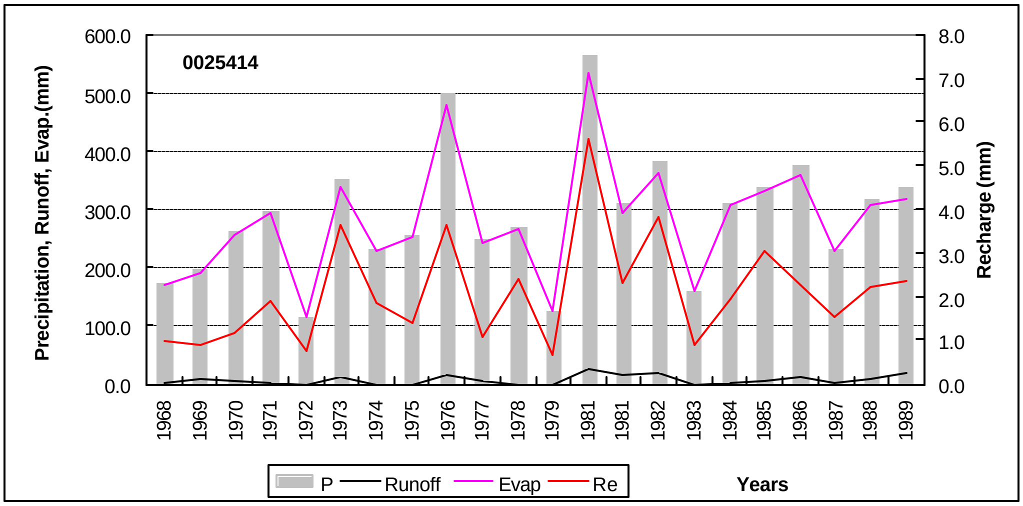


Figure 6.18 Relationship between precipitation, runoff, actual evapotranspiration and recharge for the rainfall station 0025414

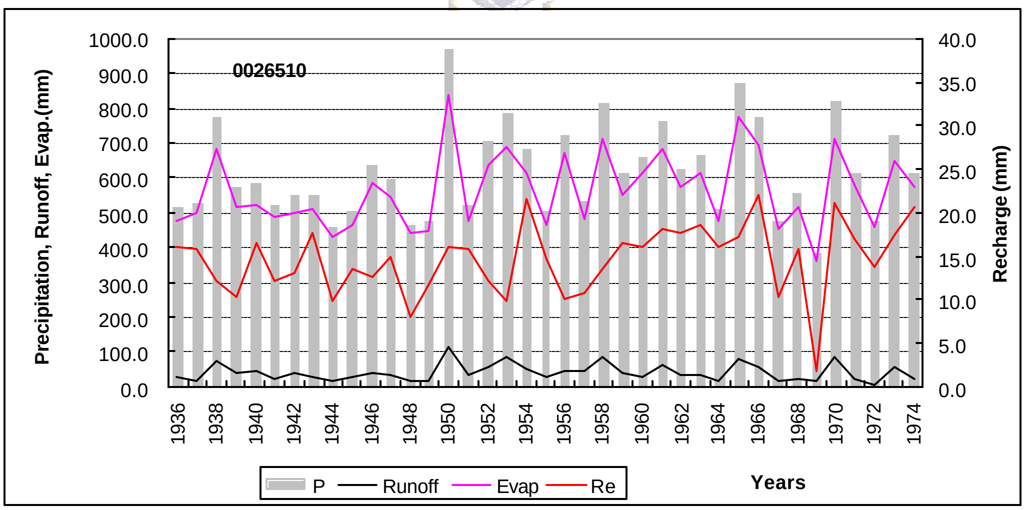


Figure 6.19 Relationship between precipitation, runoff, actual evapotranspiration and recharge for the rainfall station 0026510

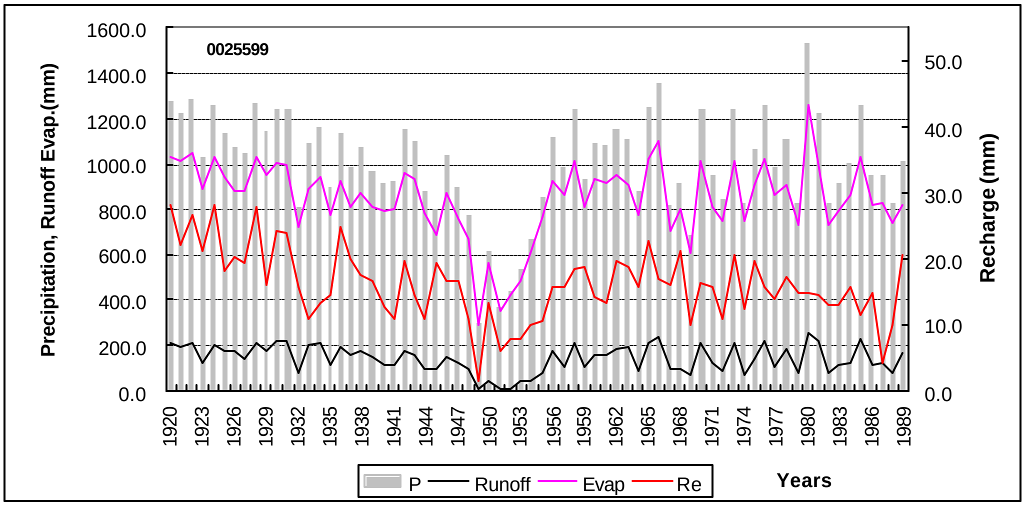


Figure 6.20 Relationship be tween precipitation, runoff, actual evapotranspiration and recharge for the rainfall station 0025599

6.5 RECHARGE VOLUME IN THE MONTAGU AREA

An attempt is made to work out annual groundwater recharge volumes in the quaternary catchments in the outcrop of the TMG in the Montagu area. On the basis of the water balance approach, the recharge volumes are calculated as

$$Q_a = 1000RE \times S_c \qquad (6.1)$$

where Q_a is annual recharge volume (m^3), RE is annual recharge rate (mm/yr), S_c is the catchment area (km^2).

Based on equation (6.1), the preliminary recharge volumes of the quaternary catchments in the outcrop of the TMG in the study area are listed in Table 6.4. The total annual recharge volume ranges from 16.91 to 87.20 million m^3 and the average annual recharge volume is 54.15 million m^3.

Table 6.4 Recharge volumes of the quaternary catchment in the outcrop of the TMG in the study area

Quaternary catchment	Outcrop Areas (Km^2)	Estimate of recharge ($10^6 m^3/yr$)			Quaternary catchment	Outcrop Area (Km^2)	Estimate of recharge ($10^6 m^3/yr$)		
		Mean	Max	Min			Mean	Max	Min
H20A	24.20	0.73	1.20	0.12	H90B	116.68	4.74	6.39	1.23
H20B	78.17	0.53	0.85	0.25	J12A	94.21	0.75	1.08	0.30
H20F	91.81	2.42	2.65	0.48	J12B	36.59	0.29	0.72	0.12
H20G	56.86	1.31	1.97	0.26	J12D	36.60	0.34	0.73	0.13
H30A	66.52	0.73	1.03	0.26	J12F	121.42	0.76	1.26	0.10
H30B	70.59	1.20	1.77	0.14	J12J	76.52	0.44	0.89	0.23
H30C	188.28	1.83	2.31	0.11	J12K	44.20	0.16	0.7	0.21
H30D	52.13	0.80	1.54	0.23	J12L	110.59	0.60	1.7	0.10
H40A	87.92	0.38	0.52	0.13	J12M	100.70	0.65	4.0	0.11
H40B	141.05	0.60	1.41	0.20	J13A	96.29	0.65	1.4	0.12
H70C	66.67	1.96	2.76	0.71	J13B	96.71	0.62	1.9	0.31
H70D	99.29	3.10	3.51	1.02	J13C	78.97	0.88	1.6	0.25
H70E	76.94	2.40	2.97	0.32	J40A	237.58	3.63	4.8	1.61
H80A	148.92	4.91	6.05	1.12	J40B	212.11	2.37	8.5	1.21
H80B	78.37	4.13	5.41	1.81	J40C	211.16	3.59	8.4	1.18
H90A	126.09	6.65	7.18	2.54	Total recharge		54.15	87.20	16.91

In order to compare recharge with stream flow, it is assumed that the storage of the TMG aquifer remains stable in one-year hydrological circle, and the recharge totally discharges into the stream, including cool and hot spring, the Contribution of Recharge to Stream (CRS) is calculated as follows:

$$\text{CRS (\%)} = \frac{\text{recharge volume/recharge area}}{\text{stream flow volume/drainage area}} \times 100\% \qquad (6.2)$$

The naturalised monthly stream flow is adopted from Midgley (1994). The average recharge water is approximately 29.3% of the stream flow as illustrated in Table 6.5. This implies that recharge may contribute about 29.3% of the stream flow in terms of baseflow. In other words, direct runoff mainly contributes to the stream flow. The detailed statistical results can be referred to Appendix 15.

Table 6.5 Comparison of steam flow and recharge in the study area

Gauge Code	Period used	MAR (Mm^3)	Drainage area km^2
H2H001	1927-1987	91.7	697
H2H004	1966-1989	35.1	175
H2R001	1967-1986	20.9	139
H3R001	1906-1986	5.9	94
H4R002	1954-1986	9.2	377
H7H004	1950-1989	4.7	28
H7R001	1968-1989	107.2	601
H8R001	1964-1987	27.9	148
H9H002	1962-1989	20.6	89
H9H004	1968-1989	15.1	50
H9H005	1963-1989	50.0	228
H9R001	1967-1989	11.9	37
J1R002	1920-1971	2.3	558
J1R003	1956-1984	30.0	4001
J1R004	1976-1984	9.1	251
Total	-	441.6	7473.0
Recharge from the outcrop area of the TMG		54.2	3124.1
Recharge/stream flow (%)		29.3	

6.6 DISCUSSION

6.6.1 Relationship between rainfall events and recharge

As Chapter 5 described, the impact of high rainfall events on the recharge is significant, especially in semi-arid areas. The comparison between annual groundwater recharge, annual rainfall and annual actual evapotranspiration shows that the values of recharge increase with increasing rainfall and decrease with increasing actual evapotranspiration (Figure 6.21) and that, annual groundwater recharge and annual rainfall are highly correlated at rainfall station 0024101. A plot of mean monthly actual evapotranspiration and rainfall for the one-year period (Figure 6.22) shows that there are few months when rainfall exceeds actual evapotranspiration. These months are mainly, May, July, September, October, November and December and it is expected that during these months, effective recharge take place.

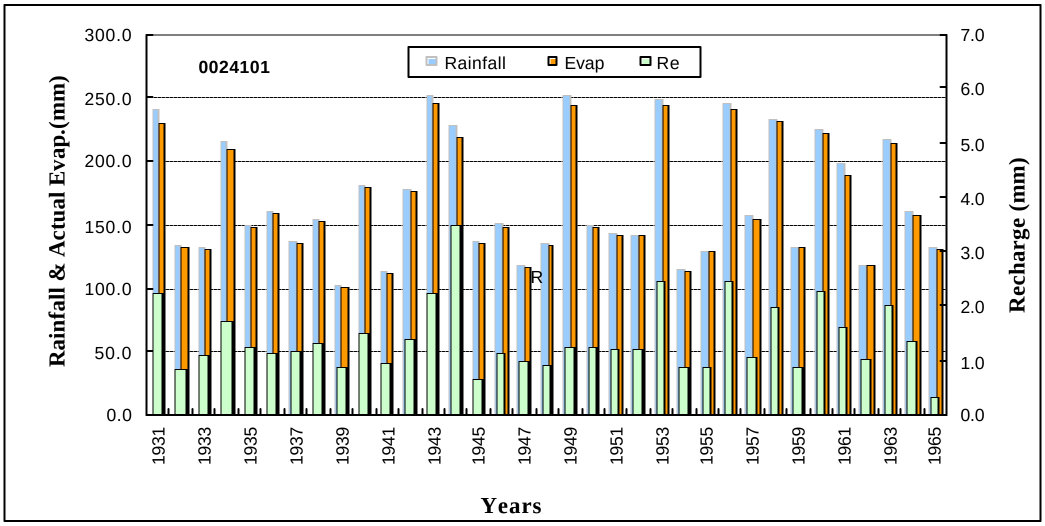


Figure 6.21 A comparison between annual groundwater recharge, rainfall and actual evapotranspiration at rainfall station 0024101

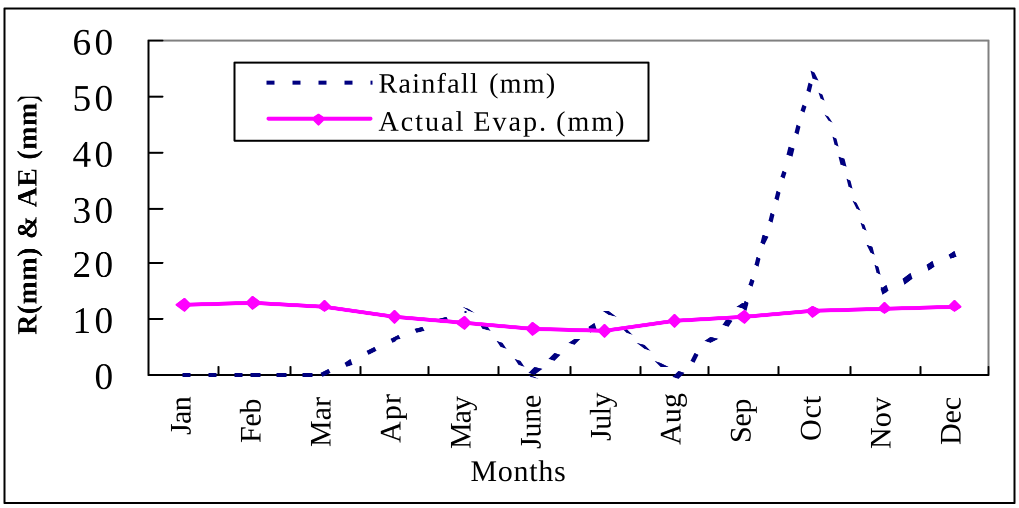


Figure 6.22 Mean monthly variation of rainfall with actual evapotranspiration for the year 1965 of rainfall station 0024101

A clear relationship between high rainfall events and recharge can be observed at rainfall station 0024101 as shown in Figure 6.23. A critical value is assumed as 0.5 times of average monthly precipitation of 14.1mm. A high rainfall event was identified if the difference between critical value and monthly actual precipitation is positive. The relationship between recharge and high rainfall events can be referred to Appendix 13. The result implies that only several high rainfall events can contribute recharge in the rainfall station. A linear positive regression relationship with y = 0.0146x+0.6955 between recharge and total precipitation of high rainfall events are figured out in the rainfall station.

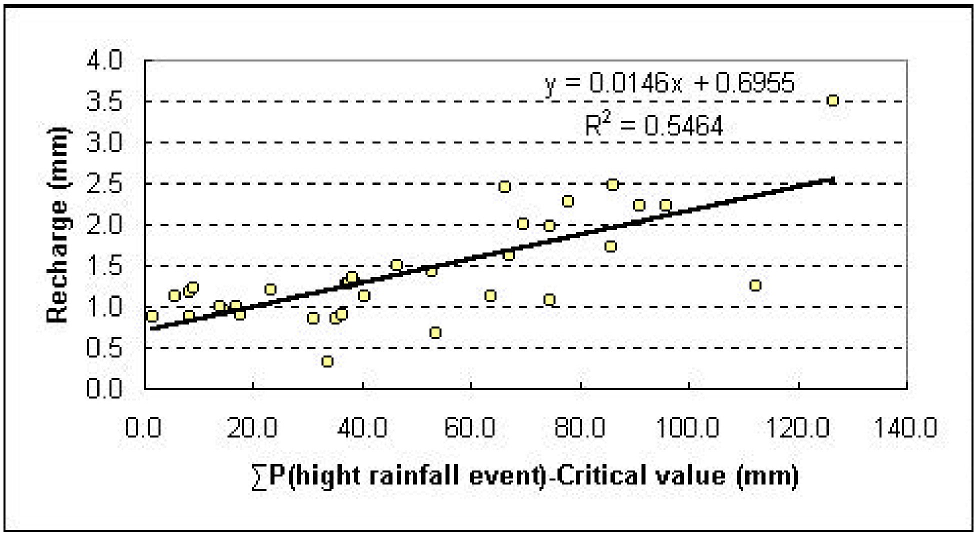


Figure 6.23 Relationship between recharge and total precipitation of high rainfall events in rainfall station 0024101

6.6.2 Recharge results with reference to other studies in the Western Klein Karoo area

The different estimates of the TMG aquifer in the study area were listed in Table 6. 6 (detailed see to Appendix 14). The lowest average estimates occur in this study; the highest one is from Jia and Xu (2005). The average estimates are 26.0 and 39.9 mm/yr related to Vegter (1995) and Fortuin (2004), respectively. Based on this study, Jia and Xu (2005) and Fortuin (2004), the Cv (coefficients of variance for annual mean precipitation) are 63, 23.1 and 36.4, respectively; but the coefficients of variance for recharge rates are 58.6, 56.8 and 76.7, respectively. If the average recharge is converted to the contribution of recharge to the stream flow, the CRS values are 156.0%, 67.5%, 44.0%, 43.0% and 29.3% related to Jia and Xu (2005), Fortuin (2004), Vegter (1995) Visser (2005) and this study, respectively. It is important to note that there was no information about rainfall stations and calculation method in the other studies mentioned above.

Table 6.6 Comparison with other results in the western Klein Karoo area (unit: mm)

Source	Item	Minimum	Maximum	Mean	Stdev.	Cv	Average CRS (%)	Method
Jia and Xu, 2005[1)]	MAP	300.0	748.0	480.2	111.0	23.1	-	Soil water balance GIS based, no algorithm
	Recharge	6.6	194.9	92.2	51.5	56.8	156.0	
Fortuin, 2004[2)]	MAP	250.2	953.9	558.3	203.2	36.4	-	GIS based
	Recharge	6.4	110.0	39.9	30.6	76.7	67.5	
Vegter, 1995[1)]	Recharge	12.0	40.0	26.0	-	-	44.0	No documented
Visser, 2005[3)]	Recharge	-	-	25.4	-	-	43.0	No documented
This study[1)]	MAP	129.5	999.0	346.6	218.2	63.0	-	Water balance
	Recharge	2.0	38.0	12.4	7.3	58.6	29.3	

1) Outcrop area
2) Quaternary catchment area
3) Quaternary catchment H40B

In the soil water balance model GIS based (Jia and Xu, 2005), 18 soil types and the soil water holding capacity ranges from 80mm to 120mm (time scale is not documented) in the outcrop of the TMG are considered. Actually, in most of the areas, the soil derived from new deposits or young formations overlying the TMG vary from place to place, even though in the same catchment; there is no more cover in the outcrop of TMG in the study area. Perhaps this is why the errors occur.

Recharge based on GIS spatial-modelling is calculated as follows (Fortuin, 2004):

$Re = MAP \times R_f \times S_f \times L_f$, with

R_f (%) =[MAP (mm)/10000]

S_f= 100-[% slope/100]

where R_f is a variable recharge rate, S_f is slope factor, and (L_f) is lithological factor.

This method is obviously empirical model. There are 17 catchments with annual mean precipitation higher than 500 mm/yr of the entire 32 catchments. The MAP is as high as 558.3 mm/yr, which exceeds the range of semi-arid area with average annual rainfall of

200-500 mm/yr (Lloyd, 1986), although the study area is located in Klein Karoo. In a certain degree, these estimates seem to be overestimated. In the mean time, the lithological recharge factors of fracture rock aquifer are larger than those of fluvial and various coastal deposits as presented in Table 6.7 (Fortuin, 2004). Obviously these values are not confirmed.

Table 6.7 Lithological recharge factors (Adopted from Fortuin, 2004)

Lithology	Lithological recharge factor
Peninsula Formation (arenite)	1.30
Cedarberg Formation (shale)	0.70
Nardouw Formation [1)] (sandstone)	1.10
Fluvial deposits	0.85
Various coastal deposits	1.00

1) Nardouw Formation should be Nardouw Subgroup.

6.6.3 Comparison of recharge estimates of studies in neighbouring area

To get an overview of the variability and magnitude of recharge figures in the Western Klein Karoo, results obtained from the Kammanassie area (Bredenkamp, 1995; Kotze, 2000; Weaver et al., 2002) are compared with this study. The Kammanassie area is located in the eastern Klein Karoo where there are similar climatic and geologic characteristics with the Western Klein Karoo. Using a range of methods and estimates of storage based on C-14 data, recharge for the entire area was set at 5% MAP (Kotze, 2000). Recharge values obtained from other methods, such as CMB, SVF (equal volume), SVF (fit), CRD, Baseflow and EARTH model are generally larger than this value. These estimates can be referred only in the similar characteristics of the outcrops and rainfall patterns.

According to Visser (2005), the flow of three cool springs in the Koo Valley with catchment code of H40B-20, H40B-32 and H40B-43 within catchment area 20.35km^2 is about 465160m^3/yr, if the volume is converted to recharge rate, which is equal 22.9mm/yr (range of 2.3~66.7mm/yr). Notice that the recharge area of the cool springs is not clarity. The three hot springs namely Avalon (Montagu), Baden and

Warmwaterberg with 2.65 million m^3/yr, if the recharge comes from whole outcrop area of the study area, the recharge rate should be approximately 0.85 mm/yr.

6.7 SUMMARY

Water balance modelling was performed using long-term average climatic and physical data from the different rainfall stations. As results, the actual evapotranspiration, runoff and recharge have been quantified. The relationships between precipitation and recharge, runoff and evapotranspiration were analysed. A comparison of recharge rates with the estimates in earlier studies was discussed. The long-term average recharges were estimated using the water balance method within the Table Mountain Group Aquifer in the Western Klein Karoo area. The following conclusions are made:

a) Based on modelling, runoff levels vary from 0.0 to 219.2 mm/yr, and average values of runoff ranges from 0.7 to 137.4 mm/yr and actual evapotranspiration calc ulated range s from 40.2 mm/yr to 1261. 0 mm/yr.

b) The recharge rates range from 0.1 mm/yr to 38.0 mm/yr in the study area and values less than 20.0 mm/yr is predominant. Relative low recharge rates coincide with low precipitation in most regions. The recharge values higher 20.0 mm/yr are more related to precipitation exceeding 800.0 mm. The low recharge rates below 2.0 mm/yr are related to high rainfall events in semi-arid areas. Effective recharges related to high rainfall events in semi-arid area are commonly. The sources of hot springs are part of the recharge.

c) Recharge percentage does not follow a linear relationship with precipitation, but rainfall events are associated. Low recharge percentage (<1%) coincide with low precipitation (< 400 mm/yr) in some periods of the stations. The ranges of 1-5% are related to precipitation from 400 to 900 mm/yr. There are relative low recharge rate (< 3%) if precipitation exceeds 800 mm/yr.

d) The recharge volume in the outcrop of the TMG in the study area is approximately 54.2×10^6 m^3/yr. The average recharge water is approximately 29.3% of the stream flow.

e) The amount of recharge is area dependent. From the water budget, approximate 78.2% and 29.3% of the stream flow contributes by direct runoff and recharge, respectively. The sum of the two values of 107.5% was produced because the drainage area and the effective radius of the rainfall stations are not takeninto account. In other words, the contributions of both direct runoff and recharge to stream flow by the rainfall in the outcrop areas should be more than that in the valleyand plain areas due to the more rainfall.

f) The differences in recharge values found for different rainfall station can only be a reflection of a wide range of environmental conditions, such as rainfall, vegetation, geology and geomorphology within the study area.

CHAPTER 7

CONCLUSIONS AND RECOMMENDATIONS

A water balance approach based on actual evapotranspiration and direct runoff models was used for recharge estimation in the Western Klein Karoo (Montagu) area of Western Cape Province, South Africa. The Table Mountain Group (TMG) aquifer in Montagu area comprises approximate 4,000m thick sequence of sandstone with an outcrop area of 3,124 km^2. The recharge area, which is the outcrop area of the TMG, is characterized by mountainous topography with sparse to dense vegetation, shallow and intermittent diverse soils and mean annual rainfall of 350-450 mm/yr. In this study, theoretical aspects of the water balance method in groundwater recharge estimation are reviewed with particular emphasis on its applicability to semi-arid areas. Geology, hydrogeology, climate, geomorphology, vegetation and soil conditions in the study area are described and analyzed. Recharge rates and volumes in the outcrop of the TMG are figured out based on the water balance model.

7.1 HYDROGEOLOGICAL SETTING

The TMG aquifer is mainly characterized by sandstone members, which are alternating felspathic quartze sandstone and coarse-grained quartz arenite units with interbedded minor conglomerate and shale. The groundwater occurrence is related to the distribution of the TMG. The outcrop of the TMG area forms the recharge areas of the TMG aquifers. The groundwater discharges as cool and hot springs as well as seepage zones all of which compose the base flow contribution to streams. The modulus of runoff in the study area is 1.9 $\ell \cdot km^{-2} \cdot s^{-1}$.

Spatial and temporal variability of rainfall in the Montagu area is addressed. According to correlation analysis from the long-term records of the 17 rainfall stations lying within or around the outcrop of TMG, five groups are identified. The different rainfall patterns of the five groups show that the rainfall patterns of the study area vary greatly, although the study area lies in the same homogeneous rainfall districts. However, similar patterns of CRD are observed in the different groups of the rainfall stations.

7.2 FACTORS INFLUENCING RECHARGE

Recharge processes are influenced by a wide variety of factors including climatic, physiographic and geological factors. Within a geographical location, it is primarily the rainfall and the climatic factors associated with the physiographic framework affecting evapotranspiration and runoff that determine the regional rainfall pattern and distribution, which ultimately influences groundwater recharge. The factors including precipitation, solar radiation, temperature, topography, vegetation and soils related to the recharge in the study area are analyzed.

7.3 RUNOFF

For the rainfall stations in the study area, estimated direct runoff rates range from 0.2 to 250.0 mm/yr with a mean value of 46.2 mm/yr in the outcrop area of the TMG. Runoff between 0.5 mm/yr and 100.0 mm/yr dominates the records for the rainfall stations. Values of more than 100.0 mm/yr occur in the higher upland areas. Large figures of runoff with more than 200.0 mm/yr are estimated in particular for the mountainous areas with altitude above 1,000 m a.m.s.l Runoff is very low if precipitation is less than 350.0 mm/yr, but runoff takes place if high rainfall events occur. Percentage of runoff ranging from 2% to 15% of the rainfall is related to the precipitation of 400-1,000 mm/yr. Percentage runoff ranging between 14.0% and 18.1% of the rainfall is related to precipitation exceeding 1,000 mm/yr.

7.4 ACTUAL EVAPOTRANSPIRATION

The actual evapotranspiration ranges from 40.2 mm/yr to 1261.0 mm/yr. The values of evaptranspiration increase with increases in precipitation. A reverse correlation exists between actual evaportranspiration percentage and precipitation. The percentages of evaportranspiration are 94-99%, 84-96% and less than 84% related to rainfall within 400 mm/yr, 400-900 mm/yr and higher than 900 mm/yr, respectively. In other words, low rainfall mainly contributes to evaportranspiration in the study area. Large precipitation or rainfall events contribute not only to runoff, but also to evaportranspiration.

7.5 RECHARGE

Recharge rates range from 0.1 mm/yr to 38.0 mm/yr and values less than 20.0 mm/yr is predominant for the rainfall stations. Considered in the study area, relatively low recharge rates coincide with low precipitation in most regions. Recharge is less than 5mm/yr if precipitation is less than 400 mm/yr. Low recharge rates coincide with low precipitation and high percentage of evapotranspiration. The ranges of 10.0-20.0 mm/yr of recharge occur in the areas of precipitation ranging from 600 mm/yr to 1,200 mm/yr. The recharge rates exceeding 20.0 mm/yr are more related to the precipitation more than 800 mm/yr. The low recharge rates below 2.0 mm/yr are mainly related to high rainfall events in the study area. This implies that a single highly rainfall event may contribute groundwater recharge.

Recharge values less than 20.0 mm/yr make up 87.3% of values in the study area as reflected in the data. Hereinto, recharge rates less than 10.0 mm/yr make up 52.7% of all statistical months within the rainfall stations. Recharge percentage is non-linear relationship with precipitation. The area with recharge less than 1% of the precipitation coincides with the areas of low precipitation (<400 mm/yr). The ranges of 1-5% are related to the precipitation range from 400 mm/yr to 900 mm/yr.

7.6 RECHARGE VOLUME

The amount of rrecharge is area dependent. Base on the water budget, the recharge volumes are calculated for quaternary catchments in referring to the area of the outcrop of the TMG in the study area. The average annual recharge volumes of the quaternary catchments range from $16.9\times10^6 m^3$/yr to $87.2\times10^6 m^3$/yr. The totally average recharge volume is approximately $54.2\times10^6 m^3$/yr, which accounts for 29.3% of the stream flow.

7.7 SUGGESTIONS AND RECOMMENDATIONS

Management strategies for groundwater utilization focus on the available water in the aquifers. The spatial variation of climate, topography, vegetation and soil associated with hydrogeological settings make it difficult to estimate recharge, and further more, the insufficient and inaccurate information result in errors in recharge estimation.

From the water budget, it is approximate that 78.2% and 29.3% of the stream flow is contributed by direct runoff and recharge, respectively. The sum of the two values, 107.5%, was produced because the factors of drainage area and effective radius of the rainfall stations were not taken into account. In other words, the contribution of both direct runoff and recharge to stream flow by the rainfall in the outcrop areas (high elevation) should be more than that of the valley and plain areas due to greater rainfall.

The method used yields a point estimate and then extends to the whole study area. The spatial distribution of recharge may be exaggerated or underestimated due to the finite number of rainfall stations in the outcrop of the TMG of the study area. The differences in recharge values estimated in different rainfall stations may only be a reflection of different recharge conditions, such as rainfall, vegetation, geology and geomorphology within the study area. A more detailed and special investigation of related information should be undertaken in order to improve the recharge rate accuracy. The recharge rate based on this method should also be crosschecked with other recharge estimation methods if additional data is available.

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Appendix 1 Minimum monthly precipitations in statistic period in the study area (mm)

Station code	Period used	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
0007311	1932-1974	0.0	0.0	3.1	1.1	3.8	6.8	2.7	2.8	2.6	0.0	0.0	0.0
0008136	1924-1974	0.0	0.0	0.0	0.0	0.0	5.8	0.0	0.0	2.9	0.0	0.0	0.0
0008782	1923-1989	5.1	8.2	11.4	4.5	2.0	3.9	0.0	0.0	0.0	0.0	0.0	0.0
0011451	1968-1989	3.0	0.0	4.5	3.5	2.5	10.3	0.0	1.4	3.0	7.0	0.0	0.0
0023070	1978-1989	0.0	0.0	0.0	0.0	0.0	0.0	0.0	6.0	0.0	0.0	0.0	0.0
0023218	1937-1950	0.0	0.0	0.0	7.6	6.8	6.6	3.5	2.0	3.5	1.8	0.5	0.0
0023602	1947-1989	0.0	0.0	0.0	0.0	1.2	4.1	0.0	1.3	2.0	0.0	0.0	0.0
0023611	1927-1987	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	3.0	0.0	0.0	0.0
0023706	1920-1946	0.0	0.0	0.0	0.0	0.0	4.1	0.0	0.0	8.1	0.0	0.0	0.0
0024101	1931-1965	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
0024684	1931-1952	0.0	0.0	0.0	0.0	1.3	1.8	6.3	0.0	2.5	0.0	0.0	0.0
0025162	1920-1974	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
0025270	1924-1938	21.9	28.9	34.9	1.0	9.9	10.5	23.8	25.6	15.1	12.7	37.8	17.8
0025414	1925-1989	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
0025599	1920-1989	10.2	16.2	11.1	0.0	11.8	7.8	5.5	9.1	0.0	18.6	1.5	10.3
0026240	1928-1962	7.0	13.0	8.1	6.4	5.8	5.8	0.0	0.0	0.0	16.2	0.0	0.0
0026510	1936-1989	3.8	10.5	12.7	0.0	4.6	2.3	4.4	2.4	0.0	11.8	1.7	4.6
0026824	1969-1989	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
0043239	1920-1939	0.0	0.0	0.0	0.0	0.0	3.0	4.9	3.0	1.5	0.0	0.0	0.0

Appendix 2 Maximum monthly precipitations in statistic period in the study area (mm)

Station code	Period used	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
0007311	1932-1974	69.9	115.1	61.4	111.0	133.7	100.2	84.9	124.5	87.0	115.5	116.6	50.4
0008136	1924-1974	78.3	86.6	113.2	157.5	122.2	118.4	111.2	223.3	112.5	141.3	154.8	81.9
0008782	1923-1989	164.2	200.6	195.7	241.7	204.0	110.0	132.8	307.0	234.4	201.9	239.6	213.3
0011451	1968-1989	145.3	96.3	79.7	143.5	111.0	119.0	78.0	174.8	82.3	94.0	93.5	58.4
0023070	1978-1989	136.0	55.0	50.0	145.0	70.0	57.0	90.5	59.0	47.5	37.0	62.0	31.0
0023218	1937-1950	49.8	37.9	49.2	78.8	91.8	61.8	64.0	72.3	65.5	72.6	111.7	58.9
0023602	1947-1989	121.1	101.7	47.7	92.5	91.5	95.4	82.5	154.0	63.3	65.1	129.3	53.7
0023611	1927-1987	306.0	179.5	105.5	249.0	174.1	190.5	145.5	401.0	150.4	232.1	180.4	114.3
0023706	1920-1946	151.1	130.6	103.9	92.0	104.6	108.0	75.9	119.9	74.8	104.6	114.5	106.3
0024101	1931-1965	41.7	32.5	57.0	42.0	110.0	52.5	58.5	86.7	89.9	57.4	78.2	37.6
0024684	1931-1952	74.4	54.6	49.9	53.2	127.8	75.4	105.4	57.4	67.8	176.0	77.5	88.7
0025162	1920-1974	54.0	116.9	56.4	78.7	67.6	38.8	45.3	60.0	70.7	69.1	95.3	72.7
0025270	1924-1938	80.3	161.6	138.9	119.8	94.7	70.1	112.3	186.5	276.2	222.0	249.7	139.3
0025414	1925-1989	116.0	107.2	124.5	160.5	164.4	88.3	91.0	154.0	137.0	201.4	105.4	106.4
0025599	1920-1989	250.8	268.6	279.4	321.0	222.6	147.0	159.9	316.0	298.4	260.1	325.4	256.0
0026240	1928-1962	142.2	171.4	178.8	134.7	167.7	93.0	108.4	241.5	239.6	223.5	248.3	212.9
0026510	1936-1989	171.6	184.8	157.8	235.0	170.5	82.5	130.8	232.0	147.6	170.5	219.7	125.1
0026824	1969-1989	132.0	64.8	71.5	84.5	55.0	45.8	53.5	97.8	77.0	56.3	88.5	56.5
0043239	1920-1939	65.8	51.1	33.1	59.7	73.4	293.8	141.2	82.0	61.6	61.2	31.1	63.8

Appendix 3 Average monthly precipitations in statistic period in the study area (mm)

Station code	Period used	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
0007311	1932-1974	12.0	18.4	24.1	31.0	43.1	40.7	43.4	50.3	32.4	32.9	26.9	11.9
0008136	1924-1974	17.0	22.7	33.0	33.5	37.0	35.3	36.9	49.0	32.0	39.5	35.2	16.3
0008782	1923-1989	50.7	57.2	68.8	64.6	53.6	44.7	51.5	67.3	57.6	66.4	64.1	36.5
0011451	1968-1989	29.0	30.4	32.0	43.1	35.8	38.0	33.6	45.5	31.4	36.4	35.1	28.0
0023070	1978-1989	16.9	10.5	13.9	36.9	22.6	31.0	25.7	29.4	20.5	14.0	7.9	5.6
0023218	1937-1950	9.6	6.7	16.4	29.6	28.7	29.3	26.8	25.8	28.1	28.1	22.2	10.6
0023602	1947-1989	11.4	13.9	14.5	27.9	31.8	38.4	32.9	39.0	19.1	17.8	17.7	10.2
0023611	1927-1987	16.1	18.6	28.8	42.8	54.5	64.0	49.7	78.2	43.5	43.9	35.6	18.4
0023706	1920-1946	28.3	28.8	36.8	33.1	30.2	41.4	33.9	38.4	37.4	28.3	36.5	19.0
0024101	1931-1965	6.8	6.9	10.9	13.5	24.1	13.4	17.7	16.8	14.7	17.8	18.7	7.7
0024684	1931-1952	11.9	12.0	14.8	21.9	34.8	28.1	36.9	22.3	26.6	40.3	29.2	14.7
0025162	1920-1974	7.5	13.5	11.3	12.0	15.0	11.0	10.3	12.1	7.6	10.1	13.6	8.4
0025270	1924-1938	45.7	61.9	87.7	50.8	44.8	40.2	61.6	64.8	80.5	76.7	93.2	56.5
0025414	1925-1989	12.0	18.0	20.1	27.7	29.7	24.3	26.7	34.3	22.8	26.3	24.1	14.9
0025599	1920-1989	84.0	91.2	108.6	89.1	72.9	55.1	64.3	87.2	78.5	96.0	97.1	74.8
0026240	1928-1962	44.3	53.4	73.9	49.6	51.9	33.4	49.9	55.5	59.4	65.5	65.9	40.4
0026510	1936-1989	52.1	59.4	64.3	58.6	47.3	37.1	43.6	56.9	52.1	61.3	61.6	41.2
0026824	1969-1989	18.6	12.9	19.9	31.9	21.6	17.6	18.1	25.9	17.1	17.8	16.6	14.0
0043239	1920-1939	8.2	12.2	10.3	18.0	29.9	70.5	45.4	37.6	18.3	14.1	12.3	10.0

Appendix 4 Standard deviation monthly precipitations in statistic period in the study area

Station code	Period used	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
0007311	1932-1974	15.8	22.1	18.1	23.7	29.7	24.0	21.8	31.4	19.4	27.5	28.2	12.9
0008136	1924-1974	17.8	22.5	25.9	29.0	26.9	20.9	25.8	39.6	20.8	31.7	35.1	20.2
0008782	1923-1989	35.5	39.9	43.2	45.7	37.0	22.3	26.5	51.2	37.5	41.9	49.0	33.6
0011451	1968-1989	29.7	25.1	21.0	37.1	28.7	25.9	22.3	42.1	23.0	26.6	23.9	18.5
0023070	1978-1989	40.5	19.1	14.9	46.1	19.3	16.3	25.2	17.9	13.8	13.5	18.8	9.6
0023218	1937-1950	16.3	12.2	17.7	18.1	24.6	18.8	17.8	20.3	17.2	23.1	31.2	16.1
0023602	1947-1989	23.9	20.3	13.2	23.6	27.7	24.5	21.5	30.5	14.2	16.0	25.3	12.1
0023611	1927-1987	42.6	28.9	26.5	45.7	43.8	45.3	33.8	72.5	29.6	43.9	38.0	23.9
0023706	1920-1946	37.3	33.8	29.1	29.1	29.3	29.2	21.9	29.7	18.7	27.8	34.3	25.7
0024101	1931-1965	10.9	10.0	15.0	10.6	24.4	12.0	15.3	19.3	18.1	17.6	20.0	10.8
0024684	1931-1952	16.9	14.7	13.4	15.3	32.6	22.4	28.8	17.9	18.4	48.3	24.4	20.2
0025162	1920-1974	13.4	22.7	12.5	14.8	17.4	10.0	9.5	13.5	12.0	14.2	18.3	15.0
0025270	1924-1938	19.0	35.7	33.0	31.0	28.1	17.9	30.6	38.8	62.4	53.5	56.3	33.3
0025414	1925-1989	20.6	22.8	22.9	32.8	31.1	16.4	23.3	32.9	23.1	36.2	25.4	20.8
0025599	1920-1989	53.2	52.8	56.8	59.5	46.4	29.5	32.3	56.7	46.0	53.9	61.4	50.8
0026240	1928-1962	31.4	41.1	39.5	33.4	38.0	18.8	28.2	50.8	46.5	46.7	58.9	40.5
0026510	1936-1989	39.8	41.3	37.1	45.8	36.6	18.8	30.3	46.4	34.4	38.8	44.3	28.2
0026824	1969-1989	29.2	16.3	22.0	27.2	19.1	9.8	15.0	25.0	17.3	18.8	21.7	16.8
0043239	1920-1939	14.9	16.9	10.9	17.4	23.5	74.1	37.7	23.8	16.7	18.3	10.0	15.7

Appendix 5 Results of runoff, actual evapotranspiration and recharge rates of the rainfall station 0025599

Year	MAP (mm)	Runoff (mm)	Evap(mm)	Re (mm)	RE%
1920	1271.70	210.28	1033.50	27.92	2.20
1921	1225.40	187.78	1015.62	22.00	1.80
1922	1284.50	211.50	1046.44	26.56	2.07
1923	1033.30	121.99	890.06	30.00	2.06
1924	1262.00	201.64	1032.14	28.22	2.24
1925	1134.80	175.46	941.27	18.07	1.59
1926	1074.90	178.52	876.12	20.26	1.88
1927	1041.80	145.14	877.30	19.36	1.86
1928	1267.20	209.48	1029.84	27.88	2.20
1929	1146.90	180.45	950.56	15.89	1.39
1930	1244.30	219.24	1001.06	24.00	1.93
1931	1243.10	219.16	1000.10	23.85	1.92
1932	809.60	73.98	719.84	15.78	1.95
1933	1098.10	197.98	889.37	10.75	0.98
1934	1162.20	207.56	941.29	13.36	1.15
1935	896.00	107.60	774.02	14.38	1.60
1936	1137.90	188.28	924.76	24.86	2.18
1937	988.00	160.60	807.49	19.90	2.01
1938	1071.00	180.15	873.21	17.64	1.65
1939	969.10	146.79	805.80	16.52	1.70
1940	913.40	108.66	792.03	12.71	1.39
1941	923.60	111.83	800.87	10.90	1.18
1942	1152.70	177.12	956.12	19.46	1.69
1943	1098.70	157.16	927.16	14.38	1.31
1944	883.00	94.87	777.23	10.90	1.23
1945	801.80	99.10	683.32	19.38	2.42
1946	1038.40	147.40	874.44	16.56	1.59
1947	901.90	128.27	756.89	16.73	1.86
1948	772.10	91.84	669.39	10.87	1.41
1949	298.60	5.48	291.72	1.40	0.47
1950	618.40	43.42	561.74	13.24	2.14
1951	364.60	9.96	348.57	6.07	1.66
1952	440.70	8.81	423.92	7.97	1.81
1953	535.00	43.47	483.67	7.86	1.47
1954	663.60	44.12	609.69	9.79	1.48
1955	856.00	78.13	767.36	10.51	1.23
1956	1115.50	175.16	924.54	15.80	1.42
1957	984.60	102.23	866.80	15.56	1.58
1958	1240.40	209.40	1012.65	18.35	1.48

Appendix 5 Results of runoff, actual evapotranspiration and recharge rates of the rainfall station 0025599 (continued)

Year	MAP (mm)	Runoff (mm)	Evap(mm)	Re (mm)	RE%
1959	932.00	103.17	810.17	18.66	2.00
1960	1095.60	152.10	929.37	14.13	1.29
1961	1082.50	155.61	913.48	13.41	1.24
1962	1150.50	180.75	950.29	19.46	1.69
1963	1112.20	186.61	906.96	18.63	1.68
1964	879.50	87.09	776.71	15.70	1.78
1965	1251.80	208.39	1020.80	22.61	1.81
1966	1360.90	241.53	1102.42	16.95	1.25
1967	812.80	89.35	707.55	15.90	1.96
1968	910.60	89.12	800.36	21.11	2.32
1969	686.60	67.25	609.31	10.04	1.46
1970	1246.30	212.68	1017.46	16.16	1.30
1971	952.90	126.27	810.94	15.69	1.65
1972	842.50	86.65	744.97	10.88	1.29
1973	1240.10	207.65	1011.99	20.45	1.65
1974	827.10	70.88	743.86	12.36	1.49
1975	1059.00	135.22	904.30	19.48	1.84
1976	1258.00	220.15	1022.37	15.49	1.23
1977	984.70	103.81	866.89	14.00	1.42
1978	1110.60	185.54	907.86	17.21	1.55
1979	824.20	73.05	736.34	14.81	1.80
1980	1531.70	255.93	1261.05	14.72	0.96
1981	1226.30	215.23	996.60	14.47	1.18
1982	827.30	78.74	735.58	12.98	1.57
1983	913.40	108.29	792.03	13.08	1.43
1984	1002.60	126.60	860.52	15.48	1.54
1985	1264.30	228.85	1023.98	11.47	0.91
1986	949.60	114.29	820.46	14.85	1.56
1987	950.10	121.96	823.85	4.29	0.45
1988	827.90	78.23	739.64	10.03	1.21
1989	1010.30	168.48	821.06	20.76	2.05

Appendix 6 Results of runoff, actual evapotranspiration and recharge rates of the rainfall station 0026510

Year	MAP (mm)	Runoff (mm)	Evap(mm)	Re (mm)	RE %
1936	518.50	29.55	472.87	16.08	3.10
1937	531.10	20.25	495.09	15.76	2.97
1938	772.70	75.85	684.61	12.24	1.58
1939	570.20	39.83	520.02	10.35	1.81
1940	583.00	43.01	523.39	16.60	2.85
1941	521.90	20.55	489.11	12.25	2.35
1942	552.10	38.49	500.6	13.01	2.36
1943	553.90	24.36	511.93	17.6	3.18
1944	457.10	16.94	430.38	9.78	2.14
1945	503.80	29.28	460.97	13.55	2.69
1946	639.60	39.90	587.05	12.65	1.98
1947	593.50	35.43	543.12	14.95	2.52
1948	467.70	18.16	441.60	7.95	1.70
1949	475.00	14.81	448.49	11.70	2.46
1950	965.30	113.13	836.02	16.15	1.67
1951	523.30	34.57	472.95	18.90	3.61
1952	704.70	56.05	636.42	12.24	1.74
1953	788.40	86.75	691.80	9.84	1.25
1954	681.30	48.01	611.77	21.52	3.16
1955	505.00	26.70	463.63	14.67	2.90
1956	724.40	43.88	670.48	10.04	1.39
1957	535.10	43.97	480.42	10.71	2.00
1958	813.30	85.96	713.86	13.49	1.66
1959	611.50	42.65	552.40	16.45	2.69
1960	660.70	28.60	615.99	16.11	2.44
1961	764.70	62.30	684.38	18.02	2.36
1962	626.80	35.38	573.59	17.82	2.84
1963	665.60	32.93	614.02	18.65	2.80
1964	511.00	17.15	477.69	16.16	3.16
1965	869.40	76.67	775.60	17.13	1.97
1966	774.10	59.25	692.79	22.05	2.85
1967	477.70	13.06	454.29	10.35	2.17
1968	556.60	21.94	518.86	15.79	2.84
1969	383.00	17.47	363.77	1.76	0.46
1970	817.70	85.69	710.80	21.21	2.59
1971	617.50	21.91	578.78	16.81	2.72
1972	475.90	5.02	457.13	13.75	2.89
1973	721.60	58.36	645.85	17.39	2.41
1974	611.80	20.59	570.40	20.81	3.40

Appendix 7 Results of run off, actual evapotranspiration and recharge rates of the rainfall station 0025414

Year	MAP (mm)	Runoff (mm)	Evap(mm)	Re (mm)	RE %
1968	171.5	1.6	168.9	1.0	0.6
1969	199.1	8.5	189.7	0.9	0.4
1970	262.4	4.0	257.3	1.1	0.4
1971	295.7	2.6	291.2	1.9	0.6
1972	117.0	0.0	116.3	0.7	0.6
1973	350.7	10.2	336.9	3.6	1.0
1974	232.0	0.4	229.8	1.8	0.8
1975	255.5	1.1	253.0	1.4	0.5
1976	498.9	16.5	478.8	3.6	0.7
1977	249.1	7.2	240.8	1.1	0.4
1978	269.8	0.2	267.2	2.4	0.9
1979	125.9	0.0	125.2	0.7	0.5
1981	563.7	24.9	533.3	5.6	1.0
1981	309.0	14.7	292.0	2.3	0.8
1982	383.2	17.3	362.1	3.8	1.0
1983	160.2	0.0	159.3	0.9	0.5
1984	309.1	1.5	305.6	2.0	0.6
1985	336.6	4.4	329.2	3.0	0.9
1986	375.1	14.2	358.6	2.3	0.6
1987	232.1	1.5	229.0	1.6	0.7
1988	318.3	8.7	307.3	2.2	0.7
1989	339.0	19.4	317.2	2.3	0.7

Appendix 8 Results of runoff, actual evapotranspiration and recharge rates of the rainfall station 0025162

Year	MAP (mm)	Runoff (mm)	Evap(mm)	Re (mm)	RE %
1936	152.0	0.8	150.4	0.8	0.5
1937	67.7	0.0	67.2	0.5	0.8
1938	109.0	0.0	108.1	0.9	0.8
1939	89.3	0.0	88.6	0.7	0.8
1940	158.7	1.3	156.0	1.5	0.9
1941	75.1	0.0	74.5	0.6	0.8
1942	159.4	0.0	158.3	1.1	0.7
1943	160.9	0.1	159.6	1.2	0.8
1944	91.7	0.0	91.3	0.4	0.5
1945	95.5	0.0	95.1	0.4	0.5
1946	122.4	1.1	120.7	0.6	0.5
1947	40.5	0.0	40.2	0.3	0.8
1948	75.1	0.5	74.5	0.1	0.1
1949	138.0	0.1	136.9	1.0	0.7
1950	117.8	0.3	116.8	0.7	0.6
1951	86.1	0.0	85.4	0.7	0.8
1952	259.6	9.2	247.6	2.8	1.1
1953	136.1	0.1	135.0	1.0	0.7
1954	178.0	2.8	173.7	1.5	0.8
1955	139.6	0.0	138.5	1.1	0.8
1956	184.5	0.0	183.3	1.2	0.7
1957	139.0	0.2	137.9	0.9	0.6
1958	149.5	0.0	148.4	1.1	0.7
1959	86.5	0.0	85.8	0.7	0.8
1960	135.5	0.0	134.4	1.1	0.8
1961	153.2	1.6	150.6	1.0	0.7
1962	241.0	2.9	235.2	2.9	1.2
1963	151.0	0.0	149.8	1.2	0.8
1964	126.4	0.0	125.4	1.0	0.8
1965	125.8	0.0	124.8	1.0	0.8
1966	166.4	0.7	164.5	1.2	0.7
1967	120.8	0.0	119.8	1.0	0.8
1968	56.2	0.0	55.7	0.5	0.8
1969	83.5	0.0	82.8	0.7	0.8
1970	146.3	0.1	145.0	1.3	0.9
1971	123.1	0.0	122.1	1.0	0.8
1972	92.8	0.0	92.1	0.7	0.7
1973	265.3	4.1	259.8	1.4	0.5
1974	128.8	3.1	124.9	0.8	0.6

Appendix 9 Results of runoff, actual evapotranspiration and recharge rates of the rainfall station 0024101

Year	MAP (mm)	Runoff (mm)	Evap(mm)	Re (mm)	RE %
1931	241.5	9.5	229.7	2.2	0.9
1932	133.1	1.1	131.2	0.8	0.6
1933	131.1	0.0	130.0	1.1	0.8
1934	214.8	3.4	209.7	1.7	0.8
1935	149.3	0.0	148.0	1.3	0.8
1936	160.2	0.2	158.8	1.1	0.7
1937	136.8	0.0	135.6	1.2	0.8
1938	154.1	0.0	152.8	1.3	0.8
1939	101.9	0.0	101.0	0.9	0.8
1940	181.2	0.0	179.7	1.5	0.8
1941	113.0	0.0	112.0	1.0	0.8
1942	177.9	0.1	176.4	1.4	0.8
1943	251.1	2.8	246.1	2.2	0.9
1944	227.9	6.5	217.9	3.5	1.5
1945	136.5	0.5	135.3	0.7	0.5
1946	150.9	1.4	148.4	1.1	0.8
1947	117.4	0.0	116.4	1.0	0.8
1948	135.4	1.2	133.2	0.9	0.7
1949	251.8	6.5	244.0	1.3	0.5
1950	149.1	0.0	147.8	1.3	0.8
1951	143.5	0.0	142.3	1.2	0.8
1952	142.1	0.0	140.9	1.2	0.8
1953	249.4	2.3	244.7	2.5	1.0
1954	114.0	0.0	113.1	0.9	0.8
1955	129.6	0.0	128.7	0.9	0.7
1956	245.5	2.0	241.1	2.4	1.0
1957	157.7	3.2	153.4	1.1	0.7
1958	233.4	0.0	231.4	2.0	0.8
1959	132.7	0.0	131.8	0.9	0.7
1960	226.0	1.2	222.5	2.3	1.0
1961	198.7	8.4	188.7	1.6	0.8
1962	118.2	0.0	117.2	1.0	0.8
1963	217.1	0.7	214.5	2.0	0.9
1964	159.6	0.0	158.2	1.4	0.8
1965	131.9	0.8	130.8	0.3	0.2

Appendix 10 Results of runoff, actual evapotranspiration and recharge rates of the rainfall station 0023602

Year	MAP (mm)	Runoff (mm)	Evap(mm)	Re (mm)	RE %
1947	214.2	0.5	212.0	1.8	0.8
1948	228.6	2.6	224.6	1.4	0.6
1949	322.8	8.5	309.1	5.2	1.6
1950	242.3	3.1	237.9	1.3	0.5
1951	309.1	3.0	303.5	2.6	0.9
1952	321.2	10.8	305.7	4.6	1.4
1953	396.3	14.1	376.1	6.1	1.5
1954	336.9	9.9	324.9	2.1	0.6
1955	262.0	1.7	257.2	3.0	1.2
1956	431.2	17.1	409.2	4.9	1.1
1957	245.2	8.6	234.4	2.3	0.9
1958	281.2	10.1	268.8	2.3	0.8
1959	177.0	0.0	175.6	1.4	0.8
1960	262.5	0.3	257.7	4.4	1.7
1961	333.6	10.5	319.8	3.3	1.0
1962	333.9	8.5	321.4	4.0	1.2
1963	259.9	6.1	251.5	2.3	0.9
1964	276.9	2.2	271.9	2.8	1.0
1965	196.9	0.0	196.0	0.9	0.5
1966	189.2	0.8	187.3	1.1	0.6
1967	197.3	0.1	196.2	1.0	0.5
1968	159.7	0.7	158.2	0.8	0.5
1969	216.9	0.8	213.0	3.1	1.4
1970	211.7	1.1	208.2	2.4	1.1
1971	144.1	0.0	142.9	1.2	0.9
1972	168.8	3.6	163.5	1.6	1.0
1973	380.3	12.8	364.8	2.7	0.7
1974	257.1	5.3	250.3	1.4	0.6
1975	299.1	9.5	287.1	2.5	0.8
1976	461.0	10.5	446.0	4.5	1.0
1977	176.1	0.5	174.1	1.4	0.8
1978	273.4	2.2	268.4	2.8	1.0
1979	147.8	0.0	146.9	0.9	0.6
1980	489.8	24.8	458.2	6.8	1.4
1981	262.2	1.8	257.9	2.5	0.9
1982	278.1	8.4	268.1	1.6	0.6
1983	258.8	10.7	245.7	2.3	0.9
1984	393.9	2.8	387.5	3.7	0.9
1985	288.3	3.3	283.1	1.9	0.7
1986	272.6	1.0	268.6	2.9	1.1
1987	245.7	2.4	241.2	2.1	0.9
1988	327.4	10.5	314.3	2.7	0.8
1989	261.7	2.2	258.8	0.7	0.3

Appendix 11 Results of runoff, actual evapotranspiration and recharge rates of the rainfall station 0011451

Year	MAP (mm)	Runoff (mm)	Evap(mm)	Re (mm)	RE %
1968	373.20	24.2	344.0	5.0	1.3
1969	197.30	0.7	195.6	1.0	0.5
1970	498.00	28.0	458.5	11.4	2.3
1971	408.70	5.3	389.0	14.5	3.5
1972	251.50	0.2	245.5	5.8	2.3
1973	366.50	2.6	356.6	7.4	2.0
1974	368.90	4.9	357.8	6.2	1.7
1975	392.30	6.1	380.5	5.8	1.5
1976	533.70	28.0	489.6	16.1	3.0
1977	361.90	4.5	350.4	7.0	1.9
1978	457.60	7.0	433.9	16.7	3.7
1979	277.20	0.0	271.7	5.5	2.0
1981	828.60	99.5	707.0	22.1	2.7
1981	516.00	36.2	455.8	24.0	4.7
1982	425.90	11.1	405.3	9.5	2.2
1983	390.30	5.8	375.5	9.0	2.3
1984	481.00	15.9	450.3	14.8	3.1
1985	589.90	64.1	502.3	23.4	4.0
1986	439.00	16.9	411.0	11.1	2.5
1987	250.40	0.0	246.1	4.3	1.7
1988	379.40	15.2	361.1	3.2	0.8
1989	509.50	48.6	444.1	16.8	3.3

Appendix 12 Results of runoff, actual evapotranspiration and recharge rates of the rainfall station 0008782

Year	MAP (mm)	Runoff (mm)	Evap(mm)	Re (mm)	RE %
1947	639.60	43.88	573.27	22.45	3.51
1948	593.10	30.29	540.99	21.82	3.68
1949	541.50	29.21	493.92	18.37	3.39
1950	858.60	58.41	774.16	26.03	3.03
1951	506.70	29.17	462.18	15.35	3.03
1952	726.40	48.50	664.84	13.05	1.80
1953	740.40	40.21	675.35	24.84	3.36
1954	744.80	56.98	671.71	16.10	2.16
1955	516.50	17.57	471.12	27.81	5.38
1956	804.20	63.24	720.80	20.16	2.51
1957	721.60	57.02	629.87	34.71	4.81
1958	769.00	53.91	677.11	37.98	4.94
1959	524.60	23.96	481.97	18.67	3.56
1960	652.90	32.36	595.53	25.01	3.83
1961	870.80	103.34	744.91	22.54	2.59
1962	757.40	86.46	652.65	18.28	2.41
1963	783.00	76.63	686.99	19.38	2.47
1964	569.60	27.07	521.40	21.14	3.71
1965	813.90	112.04	683.97	17.89	2.20
1966	763.50	101.98	646.21	15.31	2.00
1967	456.90	14.17	434.61	8.12	1.78
1968	491.50	20.04	464.21	7.25	1.48
1969	399.00	11.05	383.23	4.71	1.18
1970	679.30	62.58	598.13	18.59	2.74
1971	649.50	23.95	596.72	28.82	4.44
1972	499.20	13.28	479.48	6.45	1.29
1973	603.00	48.03	530.94	24.02	3.98
1974	480.10	12.99	454.54	12.56	2.62
1975	656.20	23.76	610.91	21.53	3.28
1976	876.90	79.77	769.26	27.87	3.18
1977	628.00	28.45	572.82	26.72	4.26
1978	740.20	64.92	648.50	26.78	3.62
1979	558.50	17.51	528.77	12.22	2.19
1980	947.90	106.11	819.84	21.96	2.32
1981	839.90	93.80	731.45	14.65	1.74
1982	489.20	19.43	466.52	3.26	0.67
1983	597.50	41.64	539.41	16.45	2.75
1984	670.80	66.88	585.53	18.39	2.74
1985	803.90	95.57	692.72	15.61	1.94
1986	589.70	37.93	537.89	13.88	2.35
1987	505.20	25.66	464.15	15.39	3.05
1988	641.20	73.44	544.43	23.33	3.64
1989	696.00	85.55	590.96	19.49	2.80

Appendix 13 Relationship between recharge and high rainfall events with monthly critical value of 21.1mm at the rainfall station 0024101

Year	Monthly rainfall – critical value (mm)												Sum of extra mm	High rainfall events	RE mm
	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep			
1931	16.1	-1.5	6.1	-21.1	-4.6	-21.1	-15.3	-0.5	-6.3	-11.2	-21.1	68.8	90.9	3	2.2
1932	-21.1	-13.0	-21.1	-21.1	-20.8	-7.1	-16.5	-2.0	-11.0	-0.5	35.2	-21.1	35.2	1	0.8
1933	-21.1	-17.0	-14.2	1.8	-12.7	-11.4	-10.4	-3.8	-21.1	3.8	-6.6	-9.4	5.5	2	1.1
1934	29.2	45.7	-21.1	-21.1	-9.7	-21.1	-9.4	9.4	1.7	-17.3	-10.7	-14.0	85.9	4	1.7
1935	-21.1	15.2	-12.2	-2.6	-21.1	-21.1	-18.1	-18.5	-13.7	-12.7	10.1	11.9	37.1	3	1.3
1936	-21.1	26.7	0.2	-21.1	-21.1	-10.7	-16.8	-16.5	-5.3	13.0	-21.1	0.8	40.6	4	1.1
1937	-7.6	-0.5	0.2	-21.1	-21.1	5.5	2.7	-13.7	-21.1	-19.1	-14.0	-6.6	8.3	3	1.2
1938	-12.2	12.5	-12.5	-21.1	10.6	-7.4	-5.1	-21.1	-21.1	-16.3	15.7	-21.1	38.7	3	1.3
1939	-21.1	-13.5	-21.1	-6.9	-2.1	-21.1	8.3	-21.1	-12.9	-15.6	-16.3	-7.9	8.3	1	0.9
1940	-8.7	20.3	-21.1	-5.9	-21.1	-21.1	3.7	22.6	-3.8	-6.9	-9.7	-20.3	46.5	3	1.5
1941	6.8	-12.8	-21.1	-17.5	-21.1	-21.1	-9.4	7.9	-7.3	-16.3	-21.1	-7.2	14.6	2	1.0
1942	-14.1	-17.2	16.5	12.4	-16.3	-21.1	-21.1	-5.6	-9.8	-17.0	-6.1	24.1	52.9	3	1.4
1943	-16.8	17.2	-12.0	-21.1	-21.1	-3.8	5.0	44.4	-10.4	-13.0	4.3	25.2	95.9	5	2.2
1944	-12.5	-21.1	-21.1	-21.1	-21.1	-21.1	-18.1	88.9	19.2	3.7	14.8	-15.8	126.5	4	3.5
1945	27.9	-21.1	-21.1	-21.1	-18.6	25.6	-16.8	-21.1	-18.1	-9.4	-19.6	-3.3	53.4	2	0.7
1946	-17.0	-21.1	-21.1	-21.1	-21.1	32.2	-9.2	-4.1	-13.7	31.5	-21.1	-16.5	63.6	2	1.1
1947	13.7	-17.8	-21.1	-8.9	-19.6	-12.2	-14.5	-4.6	-11.7	-5.3	-21.1	-12.7	13.7	1	1.0
1948	36.3	-13.0	-1.5	-21.1	-21.1	-21.1	-2.3	-11.7	-15.2	-19.1	-6.9	-21.1	36.3	1	0.9
1949	12.5	57.1	-18.3	-21.1	-16.3	-10.9	-1.1	-8.4	-18.6	32.2	-19.3	10.8	112.5	4	1.3
1950	-16.0	-2.4	-17.3	20.6	-21.1	-21.1	-8.9	-19.8	-1.8	17.0	-14.7	-18.6	37.5	2	1.3
1951	1.2	-21.1	-21.1	-19.8	2.3	-21.1	-8.6	-10.0	-9.9	-5.6	5.5	-1.5	8.9	3	1.2
1952	-21.1	17.5	-10.5	-21.1	-13.3	-18.1	5.9	-12.7	-4.9	-8.9	-21.1	-2.8	23.3	2	1.2
1953	7.3	-10.4	-13.5	-21.1	-21.1	-4.4	-4.1	33.4	7.4	37.4	0.9	-15.6	86.2	5	2.5
1954	-21.1	-21.1	-21.1	-21.1	8.9	-21.1	-21.1	-21.1	5.4	-1.6	16.9	-21.1	31.1	3	0.9
1955	-8.6	-21.1	-21.1	-8.1	-21.1	0.5	-14.6	0.9	-1.6	-6.1	-6.1	-16.6	1.3	2	0.9
1956	-4.1	-21.1	14.4	-21.1	-1.1	2.4	-21.1	40.9	1.9	4.4	-5.6	2.4	66.2	6	2.4
1957	-21.1	-21.1	-21.1	-21.1	-21.1	-7.6	-7.6	45.1	-11.1	-21.1	29.4	-17.1	74.4	2	1.1
1958	-5.5	-19.1	-21.1	3.4	11.4	-17.1	17.9	20.9	-19.6	20.9	-4.3	-7.6	74.3	5	2.0
1959	2.9	7.4	-21.1	-17.9	-18.6	-18.1	-9.6	-9.6	7.4	-8.6	-21.1	-13.6	17.6	3	0.9
1960	-21.1	13.4	-9.1	-13.6	-10.6	35.9	20.9	-8.6	-18.1	-8.1	-16.1	7.9	78.0	4	2.3
1961	-8.1	-7.6	-21.1	-0.1	-21.1	-21.1	-8.6	-14.6	1.4	-3.1	65.6	-16.1	66.9	2	1.6
1962	-0.1	-9.6	-21.1	-21.1	-17.1	-13.6	-11.1	17.0	-19.0	-2.6	-15.6	-21.1	17.0	1	1.0
1963	21.9	-5.2	-5.6	-15.1	-21.1	-8.6	-17.1	12.9	31.4	-16.6	3.6	-16.6	69.7	4	2.0
1964	-4.0	17.9	-21.1	-21.1	-12.1	-11.6	1.7	18.7	-18.2	-12.1	-10.6	-21.1	38.2	3	1.4
1965	32.9	-6.1	0.9	-21.1	-21.1	-21.1	-14.6	-9.6	-21.1	-10.4	-21.1	-8.9	33.7	2	0.3

Appendix 14 Comparison with other results in the Western Klein Karoo area

(unit: mm)

Quartenary catchment No	Source					
	This study		Jia and Xu (2005)		Fortuin (2004)	
	MAP	Recharge	MAP	Recharge	MAP[1)]	Recharge
H20A	289.3	7.0	394.4	82.1	440.8	19.0
H20B	286.7	8.0	525.3	148.5	737.4	73.8
H20F	288.4	9.5	576.3	176.1	950.3	108.5
H20G	261.9	12.5	578.7	144.3	914.5	110.0
H30A	274.5	12.0	518.6	157.2	545.6	32.6
H30B	494.2	7.5	401.3	5.6	465.0	21.1
H30C	293.3	3.5	400.9	61.9	605.2	41.1
H30D	290.4	7.5	449.1	86.6	496.3	30.3
H40A	169.0	4.5	405.0	70.5	539.0	31.5
H40B	172.9	8.5	474.9	91.0	788.7	82.3
H70C	280.9	17.5	575.2	176.2	468.1	23.1
H70D	764.4	17.5	628.9	127.4	725.9	57.9
H70E	682.8	20.6	748.0	194.9	805.1	74.2
H80A	643.1	20.0	597.0	57.9	733.2	54.3
H80B	999.0	20.0	743.1	183.7	953.9	102.6
H90A	635.6	22.5	596.9	86.3	736.2	57.6
H90B	643.1	15.0	543.5	62.8	760.1	58.8
J12A	420.3	6.0	437.0	136.8	552.3	38.9
J12B	418.4	2.0	300.0	63.7	331.2	10.5
J12D	437.2	4.0	400.0	107.1	346.5	11.8
J12F	228.7	2.5	368.2	33.2	305.4	10.6
J12J	232.0	7.5	303.6	26.8	313.7	9.9
J12K	212.3	9.5	375.7	41.0	250.2	6.4
J12L	208.4	7.5	506.4	116.5	384.5	17.0
J12M	200.8	20.0	477.0	78.9	350.9	14.2
J13A	241.2	7.5	404.7	16.2	369.4	14.8
J13B	143.6	17.5	407.5	47.6	368.9	14.1
J13C	138.4	15.0	427.1	70.9	421.6	17.5
J40A	132.4	17.5	428.2	70.9	508.6	26.0
J40B	129.5	30.0	409.5	62.7	517.7	27.2
J40C	130.4	25.0	485.7	74.0	622.2	38.7

Appendix 15 Statistical results of stream flow in the study area (Mm^3)

Gauge	item	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Total
H2H001	Min	0.2	0.1	1.2	0.9	0.3	0.2	0.0	0.0	0.0	0.0	0.0	0.0	2.8
	Max	22.5	22.0	6.6	48.2	4.9	7.4	19.4	56.4	73.0	69.7	108.1	50.2	488.4
	Median	5.8	3.9	2.5	2.0	1.6	1.6	2.1	4.4	8.3	11.1	13.9	10.1	67.2
	Mean	7.2	4.8	2.9	3.1	1.8	1.8	2.8	9.1	12.6	15.4	18.4	11.8	91.7
	Stedv	5.1	3.5	1.4	6.0	0.9	1.1	3.2	11.8	0.0	14.9	18.6	8.8	75.0
H2H004	Min	0.8	0.5	0.5	0.4	0.2	0.2	0.2	0.2	0.4	0.2	0.1	0.6	4.3
	Max	8.4	16.9	4.0	14.0	3.0	3.0	2.0	15.7	21.9	19.0	29.2	15.6	152.6
	Median	1.7	1.0	2.0	2.9	1.3	0.5	0.6	1.2	2.9	3.6	3.7	2.9	24.2
	Mean	2.4	2.0	2.0	3.0	1.3	0.7	0.8	3.0	3.9	5.1	6.5	4.3	35.1
	Stedv	1.8	3.3	1.0	2.5	0.8	0.7	0.6	4.1	0.4	4.8	7.4	3.8	31.1
H2R001	Min	0.7	0.4	0.0	0.0	0.0	0.0	0.2	0.2	0.2	0.0	0.9	1.0	3.7
	Max	3.8	3.2	1.8	4.0	3.2	1.3	1.8	9.2	14.1	14.7	15.5	6.8	79.5
	Median	1.5	0.9	0.3	0.2	0.4	0.5	0.5	0.8	1.9	2.4	2.6	1.5	13.5
	Mean	1.7	1.1	0.5	0.7	0.9	0.6	0.6	1.9	3.3	3.2	3.9	2.5	20.9
	Stedv	0.9	0.8	0.6	1.1	1.1	0.4	0.5	2.5	0.2	3.7	3.9	1.8	17.3
H3R001	Min	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
	Max	3.0	1.1	0.9	3.2	0.7	0.7	6.8	7.2	12.4	6.1	13.9	6.6	62.5
	Median	0.1	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.1	0.1	0.3	0.3	1.1
	Mean	0.3	0.2	0.1	0.2	0.1	0.1	0.6	0.6	0.7	0.7	1.5	0.7	5.9
	Stedv	0.6	0.3	0.2	0.6	0.2	0.2	1.5	1.5	0.0	1.3	3.0	1.3	10.8
H4R002	Min	0.2	0.1	0.0	0.0	0.0	0.0	0.0	0.1	0.2	0.3	0.2	0.2	1.4
	Max	11.1	2.8	1.4	17.5	6.3	1.7	3.6	1.9	3.4	3.7	18.0	7.2	78.6
	Median	0.4	0.4	0.3	0.3	0.2	0.3	0.3	0.4	0.5	0.6	0.9	0.5	5.0
	Mean	1.0	0.5	0.3	0.8	0.4	0.4	0.5	0.5	0.9	1.0	1.9	1.0	9.2
	Stedv	2.0	0.5	0.3	3.0	1.1	0.4	0.6	0.5	0.2	0.8	3.7	1.5	14.5
H7H004	Min	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
	Max	2.1	2.5	2.3	4.3	1.6	2.1	4.6	3.0	1.8	2.6	6.7	2.8	36.4
	Median	0.2	0.1	0.0	0.0	0.1	0.1	0.0	0.1	0.1	0.1	0.3	0.3	1.6
	Mean	0.5	0.5	0.2	0.2	0.2	0.2	0.4	0.5	0.3	0.4	0.8	0.5	4.7
	Stedv	0.6	0.7	0.4	0.7	0.3	0.4	0.9	0.7	0.0	0.6	1.5	0.6	7.4
H7R001	Min	0.0	0.0	0.2	1.0	0.4	0.6	0.0	0.0	0.0	0.0	0.0	0.0	2.2
	Max	41.5	52.7	34.5	70.1	30.2	20.3	64.6	51.2	10.0	36.5	50.6	34.1	496.3
	Median	5.2	6.9	4.2	3.8	5.9	6.0	5.8	3.4	2.6	3.1	5.9	7.7	60.4
	Mean	10.1	12.7	7.2	8.0	8.9	7.2	12.5	7.8	3.4	7.1	11.9	10.4	107.2
	Stedv	11.8	14.7	7.9	14.4	8.6	6.0	16.2	11.7	0.0	9.5	14.5	10.5	125.8
H8R001	Min	0.5	0.2	0.3	0.1	0.1	0.2	0.2	0.2	0.0	0.0	0.0	0.0	1.7
	Max	7.1	14.9	11.5	12.9	6.4	5.7	23.0	11.0	5.5	6.5	13.1	8.9	126.6
	Median	2.0	2.3	1.0	0.8	1.4	1.2	1.1	0.9	1.1	1.3	1.5	2.4	17.1
	Mean	2.8	3.1	1.7	1.6	2.1	1.9	3.2	2.3	1.5	1.9	2.9	3.0	27.9
	Stedv	2.1	3.3	2.4	2.6	1.8	1.6	5.2	2.7	0.0	1.8	3.3	2.4	29.1

Appendix 15 Statistical results of stream flow in the study area (Mm^3) (continued)

Gauge & items		Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Total
H9H002	Min	0.3	0.2	0.2	0.2	0.1	0.3	0.1	0.2	0.2	0.3	0.4	0.2	2.6
	Max	7.6	13.7	8.4	14.3	5.4	4.1	13.2	10.8	4.0	4.7	14.2	11.0	111.3
	Median	1.1	1.3	0.8	0.7	0.9	0.9	1.0	0.8	0.8	1.0	1.0	1.4	11.7
	Mean	2.1	2.3	1.4	1.4	1.4	1.5	2.2	1.7	1.1	1.4	2.0	2.2	20.6
	Stedv	1.9	2.9	1.6	2.6	1.5	1.2	3.0	2.6	0.2	1.3	2.7	2.3	23.9
H9H004	Min	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.2	0.3	0.3	0.4	0.2	1.5
	Max	4.8	3.7	5.2	9.0	3.7	2.5	5.0	6.9	2.8	3.6	9.0	7.1	63.2
	Median	1.0	0.9	0.6	0.5	0.8	0.9	0.6	0.4	0.6	0.8	0.9	1.4	9.5
	Mean	1.5	1.3	1.1	1.1	1.2	1.0	1.4	1.2	0.8	1.2	1.6	1.8	15.1
	Stedv	1.3	1.1	1.2	1.9	1.1	0.8	1.5	1.8	0.3	0.9	1.9	1.6	15.4
H9H005	Min	0.3	0.2	0.1	0.1	0.1	0.1	0.0	0.0	0.1	0.1	0.4	0.2	1.6
	Max	21.0	25.3	23.4	40.7	14.5	10.4	21.7	30.6	11.4	12.4	40.8	31.4	283.4
	Median	2.2	3.2	1.4	0.6	1.9	1.6	1.6	0.6	1.1	1.9	1.6	3.3	21.0
	Mean	5.2	5.0	3.5	3.4	3.6	3.3	4.7	4.1	2.0	3.5	5.5	6.2	50.0
	Stedv	5.9	6.1	5.1	8.6	4.4	3.5	6.3	8.2	0.1	3.8	9.4	7.6	69.0
H9R001	Min	0.2	0.1	0.1	0.1	0.2	0.1	0.1	0.1	0.1	0.2	0.1	0.1	1.5
	Max	3.8	11.7	4.1	6.1	2.8	2.3	3.9	5.2	2.1	2.3	5.3	3.7	53.3
	Median	0.6	0.8	0.6	0.4	0.5	0.7	0.6	0.7	0.5	0.7	0.9	0.9	7.8
	Mean	1.1	1.5	0.8	0.8	1.0	0.8	1.2	1.0	0.6	0.8	1.2	1.1	11.9
	Stedv	1.0	2.4	0.8	1.2	0.8	0.7	1.3	1.2	0.1	0.6	1.2	0.9	12.1
J1R002	Min	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
	Max	8.9	5.0	2.9	3.3	3.7	5.1	1.2	2.3	2.2	1.7	4.6	1.5	42.3
	Median	0.0	0.1	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.3
	Mean	0.3	0.4	0.1	0.2	0.3	0.2	0.1	0.1	0.1	0.1	0.2	0.1	2.3
	Stedv	1.3	0.9	0.4	0.5	0.8	0.7	0.2	0.3	0.0	0.3	0.7	0.2	6.5
J1R003	Min	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
	Max	2.1	17.2	8.7	144.6	47.1	7.4	34.6	18.2	26.3	46.3	23.7	9.1	385.3
	Median	0.3	0.3	0.3	0.3	0.3	0.3	0.2	0.1	0.3	0.1	0.3	0.4	3.1
	Mean	0.5	1.3	1.3	6.4	4.2	1.4	2.5	1.8	2.8	3.2	3.4	1.3	30.0
	Stedv	0.6	3.3	2.2	26.7	11.2	2.1	6.8	4.2	0.0	9.2	6.3	2.3	75.0
J1R004	Min	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.1	0.0	0.0	0.2
	Max	2.2	0.6	0.6	6.3	1.6	2.0	6.0	6.3	4.0	2.9	7.4	6.3	46.0
	Median	0.2	0.3	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.3	0.4	0.6	2.3
	Mean	0.6	0.3	0.2	0.8	0.3	0.3	0.9	1.6	0.6	0.8	1.1	1.6	9.1
	Stedv	0.7	0.2	0.2	2.0	0.5	0.6	2.0	2.5	0.0	0.9	2.4	2.4	14.6