

**GROUNDWATER RECHARGE ESTIMATION OF
TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH
CASE STUDIES**

Y Xu • Y Wu • A Duah

WRC Report No. 1329/1/07



Water Research Commission



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REPORT TO THE
WATER RESEARCH COMMISSION

by

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WRC Report No 1329/1/07
ISBN 978-1-77005-574-2

MAY 2007

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Executive Summary

The Table Mountain Group (TMG) aquifer system mainly consists of sandstone, shale and quartz arenite that are exploited extensively for agricultural purposes. The groundwater recharge process and significance in TMG region, especially in Kammanassie Mountain region are thoroughly studied for sustainable utilization purpose. In contrast to previous studies, the much conservative recharge estimates obtained from Kammanassie region imply that TMG aquifer storage at some localities may be limited but at regional scale it would remain significant. The TMG as a regional fractured rock aquifer system may still become main sources of future water supplies to augment both agricultural and urban requirements in the Western and Eastern Cape Provinces in South Africa.

Conceptualisation

Although groundwater recharge is widely recognised as the key factor in determining the sustainable management of groundwater resources, no comprehensive study of groundwater recharge of TMG aquifer systems has yet been undertaken. It is within this context that the WRC approved a research proposal on "Recharge processes in the TMG aquifer systems" with the prime objective of evaluating the processes of recharge and assessing realistic recharge rates in the TMG aquifer system.

To achieve the objective, the TMG aquifer systems commonly referred to as the east, syntax, southern, and west blocks, were critically reviewed and conceptualised into

nineteen hydrogeological units involving an area of 248 000 km² based on geologic, geomorphologic, climate and hydrogeological characteristics. These hydrogeological units are verified by interpreting Landsat imagery.

The recharge processes within the different aquifers are controlled by topography, snowmelt, fracture characteristics, texture and thickness of soil and vegetation type. Infiltration and percolation rates are also governed by fracture characteristics at depths. In most cases, the percolation rate is much smaller than infiltration rate. Interflow occurs in the weathered and preferential fracture zones above the elevation of the local stream or river due to the limitation of percolation rates.

In the TMG area, the fractures in shallow rock layers together with fissures by weathering near the surface are the most important factors for rainfall infiltration and groundwater recharge in the outcrop of the aquifers.

As the occurrence of recharge is highly erratic in the TMG, a geomorphologic investigation covering the entire nineteen units was conducted. From a viewpoint of groundwater recharge, five types of aquifer settings were recognised based on their geomorphologic characteristics. These are the horizontal terrain, the inclined strata system, the folded strata belt, the weathered rock (crust) aquifer zone and the fractured rock aquifer outcrop. The TMG aquifer system, including mainly the Peninsula and

Nardouw aquifers, are regionally separated by the Cedarberg aquitard and intensely folded. The outcrops of the TMG form the main recharge areas of the aquifers and so do the foothills and foot-slopes of the mountains where flood fans occur. The recharge areas are approximately 40.6 km². The deep seated faults linked to dense fracture intersections at shallow places appear to facilitate recharge processes in the deep TMG aquifer system.

Adapted Methodology

Within the recognised units, a water balance model can be used to provide a guide value. At local scale, six scenarios of recharge processes, namely soil zones, horizontal strata zones, vertical strata zones and three inclined strata zones are identified. In most of the TMG area, a combination of these scenarios would form more complex recharge processes where multiple methods are applied for cross check purpose. A hydrogeochemical conceptual model was developed to qualify and quantify the relationship between rainfall, interflow, springs, surface water and the water in shallow, local and regional aquifers. This approach is illustrated in the case of the Kammanassie area.

Case Studies

Among a few case studies carried out, the Kammanassie area is comprehensively investigated because of the availability of comprehensive data and a well-documented problematic groundwater supply scheme there.

Long term rainfall records from 1925-2000 in the Hoekplaas rainfall station, the Klein Karoo and the Kammanassie Dam rainfall station in the southern Kammanassie

mountain area were analysed for their variation and trend. However, rainfall records from 1994 to 2003 were analysed from four significant rainfall stations in the Kammanassie mountain area where most of the recharge occur. These are the Wildebeesvlakte, Parshall, V-notch and Purification Works East rainfall stations. The time lags between rainfall series and water level responses range from a few months to a few years, suggesting an exponential flow prevails as opposed to the piston flow postulated by previous workers.

The recharge area of the Vermaak's River wellfield lies within the outcrops of the Peninsula Formation window of approximately 40.6 km². There are 9 production boreholes, 27 monitoring boreholes and 19 boreholes with no detailed data in the study area. The water levels of the production boreholes in the Vermaak's catchment have been declining since 1984 due to high abstractions.

The data used in this project included new measurements of rainwater, soil samples, water samples from boreholes, springs, rain gauges and chloride deposits in the Vermaak's valley together with existing groundwater and surface water data from the National Groundwater Database. In general, groundwater samples from the TMG aquifers are classified as Na-Cl type water, while the concentrations of Cl and Na ions in surface water are higher than those in groundwater. TDS is generally low but TDS in the production boreholes of the Peninsula Formation are lower than that in the Nardouw. Although data of trace elements or other constituents are available, incomplete data sets limit their utilisation. The Fe and Mn could not be used as tracers because they had large variations in time and space.

The isotopic signatures in the production boreholes were different from those in the monitoring boreholes. ^2H in production boreholes is more affected by evaporation or altitude than those in monitoring boreholes. It illustrates that different sources or infiltration paths occur. Isotope patterns of hot springs differentiate from those of water samples from boreholes. The water in the borehole seems recharged from low elevation areas (more negative value), but the hot springs recharged from high elevation areas or low elevation during a colder period. This would imply the flow paths are different. The flow of the boreholes is locally bounded, while the hot spring flow is regional. Tritium concentrations demonstrate that preferential recharge paths occurred in the Vermaak's River well field. The variation of the tritium in the aquifer shows two types of deep circulation, i.e. old and recent water. The ^3H cannot be used to estimate recharge rate in the TMG fractured rock aquifer due to both insufficient data and its own limitation of short half-life time. The ^{13}C values indicate that recent water component in the boreholes is more than that of the hot springs. The residence time of groundwater based on ^{14}C can only give rough ranges of recharge estimates due to both the uncertainty of its initial level and infiltration paths.

The values of recharge estimates in the Kammanassie area are less than 5% of annual rainfall, much lower than the 15 ~ 20% by previous researchers. A 3-D numeric model simulation is used to confirm the lower recharge scenario at steady state. The obtained results are compatible with those in southern Africa.

The results of the other case studies are either documented in formally published papers or reported separately.

Recharge Estimates of TMG

For a given study area, the recharge estimation should be guided by the geomorphological types mentioned above. Many methods should be applied to cross-check to results, including the Chloride Mass Balance (CMB) mixing model, statistic methods, the water level fluctuation methods (RIB and CWD-CRD). The numerical modelling proves to be an excellent tool for use in the verification of the water balance scenarios for recharge estimation in the case of the Kammanassie. At a regional scale, a soil water balance model can be used to provide a guide value. This method is applied for the 19 hydrogeological units demarcated in TMG region during this project.

The average recharge rate of the TMG is about $30 \text{ mm}\cdot\text{a}^{-1}$. The highest recharge rate is $137 \text{ mm}\cdot\text{a}^{-1}$ associated with $1842 \text{ mm}\cdot\text{a}^{-1}$ at rainfall station No. 0022116 in the hydrogeological unit 5; the lowest recharge rate is $0.7 \text{ mm}\cdot\text{a}^{-1}$ associated with $164 \text{ mm}\cdot\text{a}^{-1}$ at station No. 0048043 in the hydrogeological unit 13. The regions with a recharge rate more than $35 \text{ mm}\cdot\text{a}^{-1}$ occur in the hydrogeological units 5, 6, 8 and part of the hydrogeological unit 17. A rough estimate for the TMG hot springs would be about $5 \text{ mm}\cdot\text{a}^{-1}$. A recharge map for TMG is compiled to provide guidelines for use and reference for future updating.

The distribution of snow occurs above 1000 m - 1200 m a.m.s.l of the TMG mountain area in some winter seasons. The recharge areas influenced by snowmelt in the TMG are the catchment areas above 1000 a.m.s.l. The hourly snowmelt rates are estimated at air temperatures 4°C and 6°C as 0.328 mm/hr and 0.689 mm/hr , respectively. These snowmelt rates are the

ceiling of the percolation. Most of snowmelt water emerges in the form of interflow. Recharge rates must be much smaller than snowmelt rates. Further research is needed to compare snowmelt recharge with that of normal rainfall conditions.

Three types of vegetation, namely the Mountain Fynbos shrub, sparse vegetation and exotic plants, are dominant in the recharge areas of the TMG aquifer. The most common vegetation is the Fynbos shrub. The exotic plants consume more water than the indigenous plants and they could also reduce recharge. Several studies focusing on the impact of the exotic plant species on runoff suggest that the study of their impact on recharge would be necessary in the future.

Uncertainty

For certain methods, attention should be paid to how the estimates (% or $\text{mm}\cdot\text{a}^{-1}$) are expressed as they may be related to dependent variables such as the storativity of the aquifer and size of recharge area involved, etc.

As a rule rather than exception, all recharge estimation methods have their uncertainties and inaccuracies arising from spatial and temporal variability in processes and parameter estimations, measurement errors and validity of assumptions. The methods adapted in this study are based on the availability of data and the reliability of the methods in space and time scales in the sub-region spanning a period of three decades. Error propagation, which stems from error inputs, would therefore be duly considered as demonstrated in the report.

Acknowledgements

This research project was funded by the Water Research Commission with the assistance from the Department of Water Affairs and Forestry (DWAF).

The project team is most grateful to the steering committee members for their invaluable inputs in steering the programme from start to finish as well as the timely editing and comments. The project team is also grateful to the chairperson of the Steering Committee, Dr. Kevin Pietersen and lately Dr. Shafick Adams for managing the project efficiently. The following organisations and their representatives are acknowledged for their inputs:

- DWAF: Mr. M Smart, Mr. P Seward and Mr. E van Wyk
- CSIR: Dr. H Beekman, Dr. J Bean, Ms. C Colvin, Mr. J Weaver
- City of Cape Town: Mr. RC Bishop

We would like to acknowledge the University of the Western Cape (UWC) for its support. The staff and postgraduate students of the Department of Earth Sciences at UWC are thanked for their constructive roles in execution of the project.

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List of Acronyms

<i>CFB</i>	Cape Fold Belt
<i>CMB</i>	Chloride Mass Balance
<i>CRD</i>	Cumulative Rainfall Departure
<i>DWAF</i>	Department of Water Affairs and Forestry, South Africa
<i>KKRWSS</i>	The Klein Karoo Rural Water Supply Scheme
<i>MAP</i>	Mean Annual Precipitation
<i>MAR</i>	Mean Annual Runoff
<i>NGDB</i>	South Africa National Groundwater Database
<i>RIB</i>	Rainfall Infiltration Breakthrough
<i>TMG</i>	Table Mountain Group
<i>WRC</i>	Water Research Commission

1. INTRODUCTION

1.1 The Background of Study

The Table Mountain Group (TMG) system is pre-dominantly a fractured, deep-seated aquifer system. For more than 30 years the TMG has been considered a potential source for bulk water supply to meet both agricultural and urban requirements of both the Western and Eastern Cape Provinces. Although groundwater recharge is widely recognised as the key factor in determining the sustainable management of groundwater resources, no comprehensive study of groundwater recharge of TMG aquifer systems has yet been undertaken. Particularly the lack of monitoring data is presently seen as the key-limiting factor in quantifying recharge of TMG aquifer systems. It is within this context that the WRC approved a research study on "Recharge processes in the TMG aquifer systems". Critical issues around the sustainable development of the TMG aquifer systems are the understanding of recharge processes and factors influencing recharge, and ultimately quantification of recharge rates under various climatic, geomorphologic and hydrogeologic settings.

1.2 The Objectives

The project forms part of the hydrogeology of the TMG aquifer studies and has the following three objectives:

- Quantification of realistic recharge rates and the processes of recharge in the TMG aquifer system in the Kammanassie area;
- Evaluation of the sensitivity of recharge rates to rainfall patterns (spatial and temporal patterns) with a focus on understanding the impacts of climatic changes on groundwater recharge; and
- The use and development of innovative techniques to assist in recharge estimation in the TMG aquifer systems.

To achieve the aims, the TMG aquifer systems were subdivided into hydrogeological units based on climate, geology and hydrogeological characteristics. For confirming the hydrogeological conceptual model within the hydrogeological units, a hydro-geochemical conceptual model was used to quantify the relationship between rainfall, surface water, springs and the water in shallow, local and regional aquifers. The focus of the study is on recharge mechanisms and recharge estimation within the TMG area. The study evaluates recharge processes and recharge estimation methods in the TMG aquifer systems. The recharge estimation was illustrated with case studies. In particular, the study addressed the following issues:

- 1) Cross-calibration of recharge estimation techniques;

- 2) Spatial and temporal variability of recharge;
- 3) Impact of climatic change on recharge in the Kammanassie area.

The results of the study attempts to improve our understanding of the functioning of the TMG aquifer systems. In this way, it contributes to the sustainable development of a potential bulk water supply source in the Western and Eastern Cape Provinces.

1.3 Site Selection

The study includes the TMG area, with special emphasis on the Kammanassie area. The Kammanassie area was selected for the following reasons:

- 1) It represents a typical area with tectonic characteristics, hydrogeologic settings and climatic regimes;
- 2) Availability of relatively comprehensive data sets; and
- 3) A relatively well-documented conceptual hydrogeological model for the area.

Several previous studies in the region have provided a good source of background information (Kotze, 2000, 2001 and 2002). The outcrops of the TMG in the study area generally occur in mountainous terrain and valleys. The cold and hot springs occurring in the Kammanassie area did assist in constructing a hydrogeological conceptual model.

Several management problems have occurred during the abstraction from the Klein Karoo Rural Water Supply Scheme (KKRWSS) since 1984. These included over-abstraction, decline in borehole efficiency and iron-oxidizing bacteriological clogging of some of the borehole screens. The main concern was that the production capacity of boreholes would decrease and nearby springs at the Cedarberg Shale (C/S) band in the Kammanassie Mountains would run dry, even the ecosystem would be shifted. It is important that the hydrogeological concepts and water balance model are better understood in order to at least attempt to answer the following questions:

- How much information about flow and recharge can be revealed from hydrogeochemistry and groundwater level fluctuation in the production boreholes?
- What are the recharge processes and the realistic groundwater recharge rates in the study area?

2. CONCEPTUALISATION OF THE TMG

2.1 *Geology and Stratigraphy of the TMG*

The sediments of the Cape Supergroup were deposited in a shallow marine environment under tidal, wave and storm influences, as well as in a non-marine, braided-fluvial environment from early Ordovician to early Carboniferous. This predominantly siliclastic sequence is exposed along the entire length of the Cape Fold Belt (CFB), the 220-280 million year old Orogenic belt straddling the west and south coasts of South Africa. The succession of quartz arenites, shales and siltstone, with minor conglomerate and a thin diamictite unit, has been subdivided into the Table Mountain, Bokkeveld and Witteberg Group (Du Toit, 1954; Rust, 1967; Theron, 1962; Theron and Looek, 1988; Broquet, 1992). For the Western and Eastern Cape, maximum thickness of the Cape Supergroup amounts to 5, 300 m and 9, 600 m respectively (SACS, 1980). The medium to coarse grain sizes and relative purity of some of the quartz arenites, together with their well indurated nature and fracturing due to folding and faulting in the fold belt, enhance both the quality of the groundwater and its exploitation potential.

The presently exposed structure and thickness of the Table Mountain Group (TMG) rocks are the result of initial deposition within an east-trending basin (Rust, 1973) along the Southern and South-western Cape regions, which was modified by two major tectonic events, namely the Permo-Triassic Cape Orogeny and the fragmentation of South-western Gondwana during the Mesozoic. The Cape Orogeny had the effect of tectonically thickening the sequence in areas of high strain like the Southern Cape. The CFB consists of two branches forming a mountain chain of about 1 200 km along the south coast and part of the west coast (De Villiers, 1944; Söhnge and Hälbig, 1983). The Cape Orogeny resulted in further deformation of the metamorphosed Neoproterozoic rocks, the Cape Granite Suite, together with its cover sequence of Ordovician to Triassic rocks (Cape Super Group and part of the Karoo Sequence). The outcrop of the southern branch is about 200 km, and of the western branch, about 150 km. Both branches are arcuate in plan view and convex towards the Karoo Basin merging with northeast-trending folds in the syntaxis of the South-western Cape. The entire geological succession in the study area includes the basement rocks, Cape Supergroup and part of the Karoo. The succession with their associated thickness and lithological compositions are summarised in Figure 1. Of the entire succession only the TMG is of relevance in this study. A geological map of the study area is presented in Figure 2.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

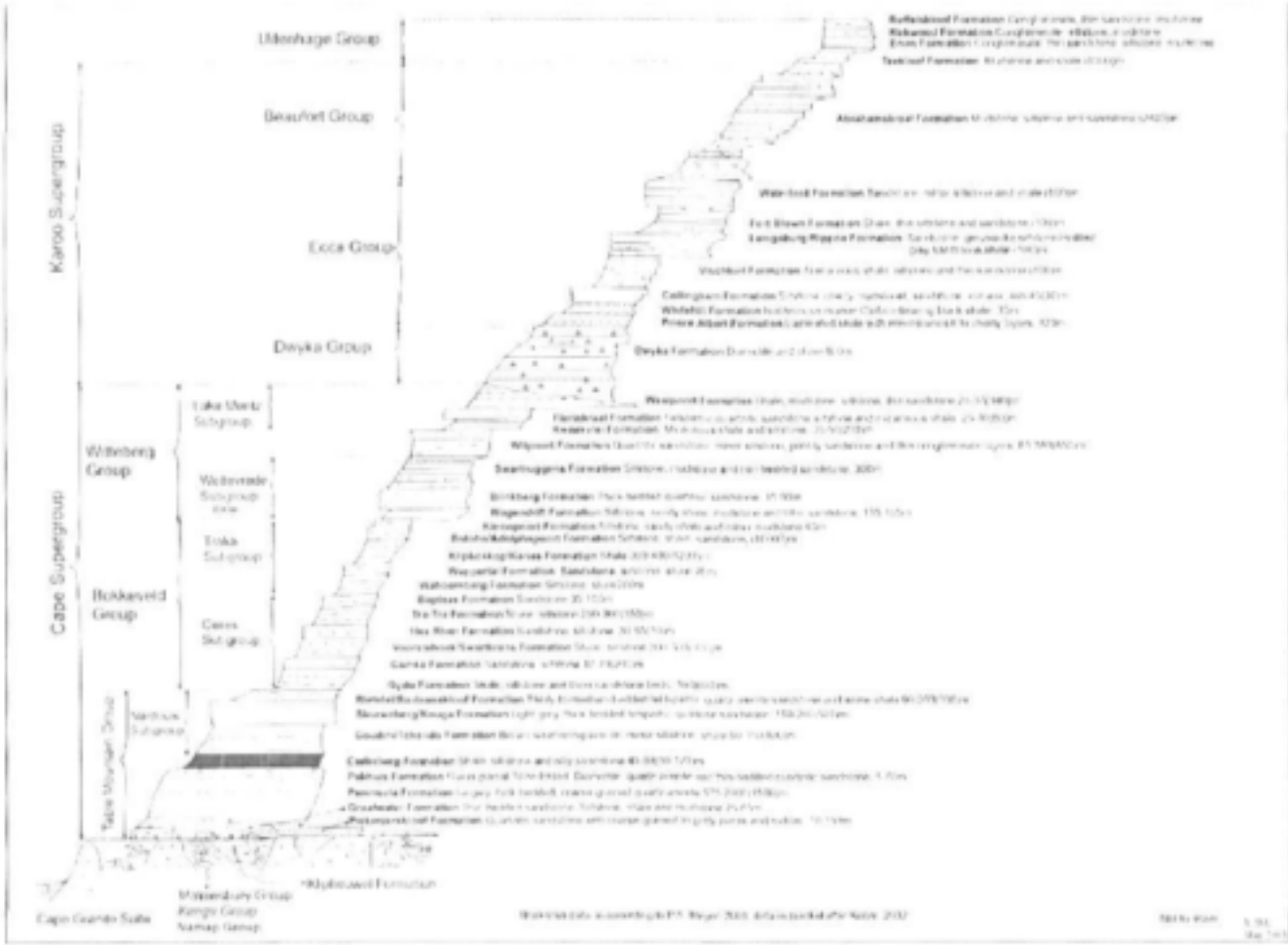


Figure 2.1 Stratigraphic sequence of the study area.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

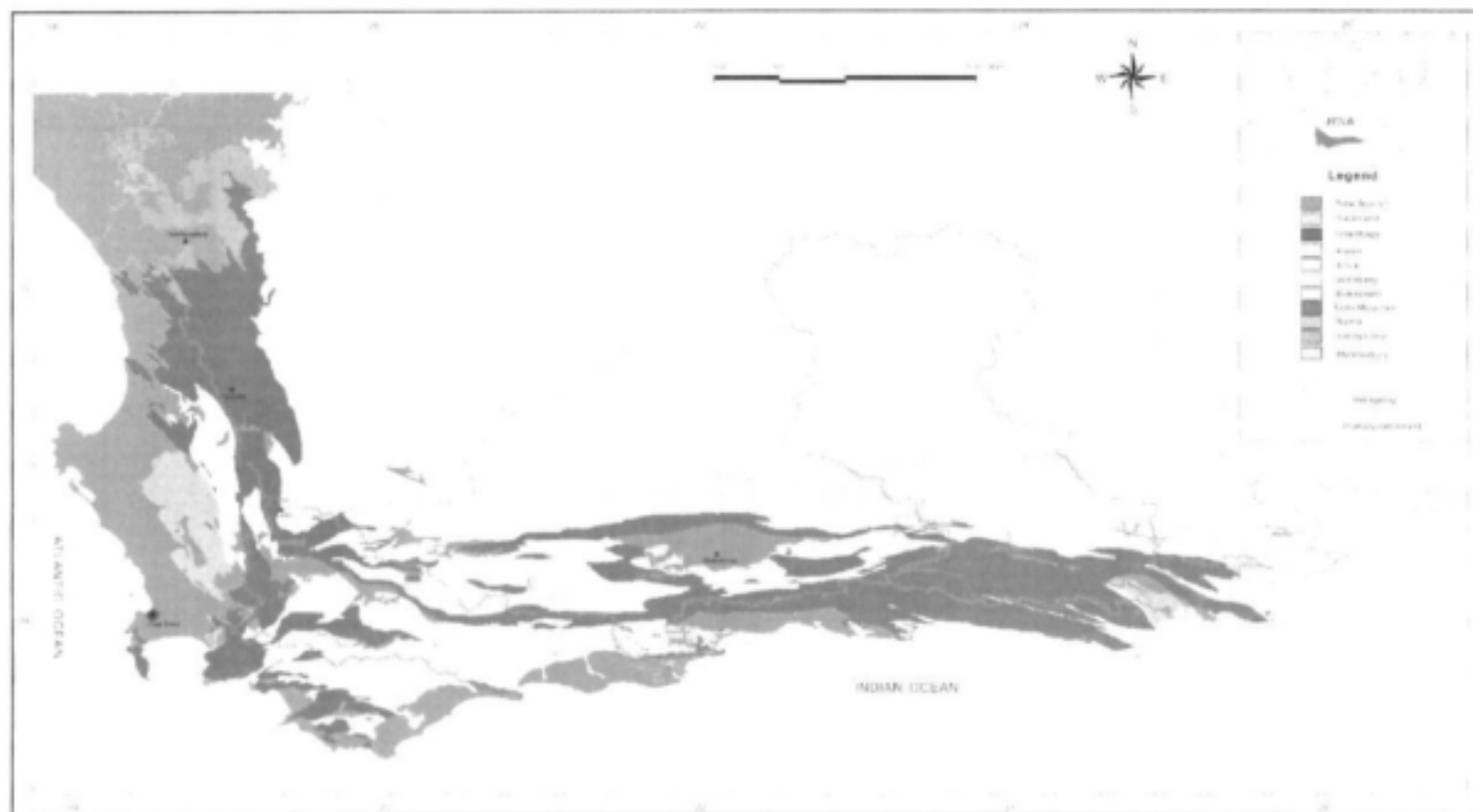


Figure 2.2 Geological map of the study area

The TMG comprises of an approximately 4 000 m thick sequence of quartz arenite and minor shale layers deposited in a shallow, but extensive, predominantly east-west striking basin, changing to a northwest orientation near Worcester. The maximum thickness in the east section (Port Elizabeth) is approximately 3010 m (Rust, 1973). The TMG comprises of the Piekenierskloof Formation, Graafwater Formation, Peninsula Formation, Pakhuis Formation, Cederberg Formation and Nardouw Subgroup. The bottom of the TMG is the Piekenierskloof Formation or Graafwater Formation. The Piekenierskloof Formation is quartzitic sandstone with coarse-grained to gritty zones, 10-150 m. The Graafwater Formation has thickness of 25-65m and comprises of thin-bedded sandstone, siltstone, shale and mudstone.

The name of the Peninsula Formation is derived from the Cape Peninsula (Rust, 1967), where the full succession (550 m) is exposed against Table Mountain. It comprises a succession of coarse-grained, white quartz arenite with scattered small pebbles and discrete thin beds of a matrix-supported conglomerate. The formation reaches a thickness of about 1800 m at Clanwilliam in the west (Rust, 1967), but is reportedly much thicker (Rust, 1973) in the Eastern Cape. The exact thickness is difficult to measure due to the severe folding and unknown amounts of thickening due to thrusting (Booth and Shone, 1992) within the formation. It is noticed that the residual thickness of the TMG (as stated above) is very different from borehole to borehole. The determination of formation thickness is also impeded by general lack of marker beds.

The Pakhuis Formation occurs above the Peninsula Formation and comprises a 40 m which succession of glacially derived sediments, but is restricted to the South-western Cape (Broquet, 1992; Rust, 1967). The Cederberg Formation with a thickness of 50-120 m comprises shale, siltstone and silty sandstone. It is prominently a marker band between the Peninsula Formation and the Nardouw Subgroup. It is a hydrodynamic and recharge boundary.

The Nardouw Subgroup, with its three subdivisions, the Goudini, Skurweberg and Rietvlei (Baviaanskloof in the Eastern Cape) Formations, is another thick (maximum 1200 m) unit of sandstone that varies between quartz arenite, silty and feldspathic arenites, accompanied by minor inter-bedded conglomerate and shale. This lithological diversity, together with textural, grain size and bedding thickness differences, lead to pronounced differences in weathering, structural and hydrogeological characteristics. The basal unit, the Goudini Formation, is characterised by reddish weathering, thin sandstone beds with common shale intercalations. The Skurweberg Formation is thick-bedded arenite. The topmost unit, the Rietvlei Formation contains more feldspar and is characterised in the field by growing vegetation on it. The contact with the overlying dark shale of the Bokkeveld Group is usually abrupt.

2.1.1 Regional Structure

The entire geological succession of the CFB was subjected to severe north-south orientated compressive stresses during the Cape Orogeny. This resulted in a variety of geological features and structures. Figure 3 shows the tectonic pattern and outline of structures. The CFB consists of two branches, namely western branch and southern branch, which form the characteristic mountain chain along the southern and part of the western coasts of South Africa. The western

and southern branches of the CFB meet in a 100 km wide syntaxis area, comprising NE-trending folds situated between Ceres and Gansbaai (approximately 200 km to the west of the study area).

In the Southern Cape and Klein Karoo, the southern branch of the CFB comprises a 900 km long gently arcuate belt. The more resistant strata of the group form prominent E-W trending mountain ranges, which are Swartberg, Roolberg, Gamkaberg, Kammanassie and Outeniqua Mountains.

The above-mentioned mountain ranges are separated by synclinal intermountane basins. The resistant quartzites are exposed in the cores of anticlinal mountains and form a rugged topography. The valleys are broad and due to the underlying less resistant Bokkeveld Group sediments.

The compressional deformation during the Cape Orogeny was followed by extensional tectonics, during which the Uitenhage Group was deposited within a number of fault-bounded basins, reaching a thickness of more than 2000 m in places. The examples of the extensional tectonics include the following (Duvenhage *et al.*, 1993):

- Reverse faults associated with over-folding, during the Cape Orogeny;
- The Congo Fault (CF), a reverse fault on a previous thrust fault plane; and
- Several normal faults of the post-Cretaceous age.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

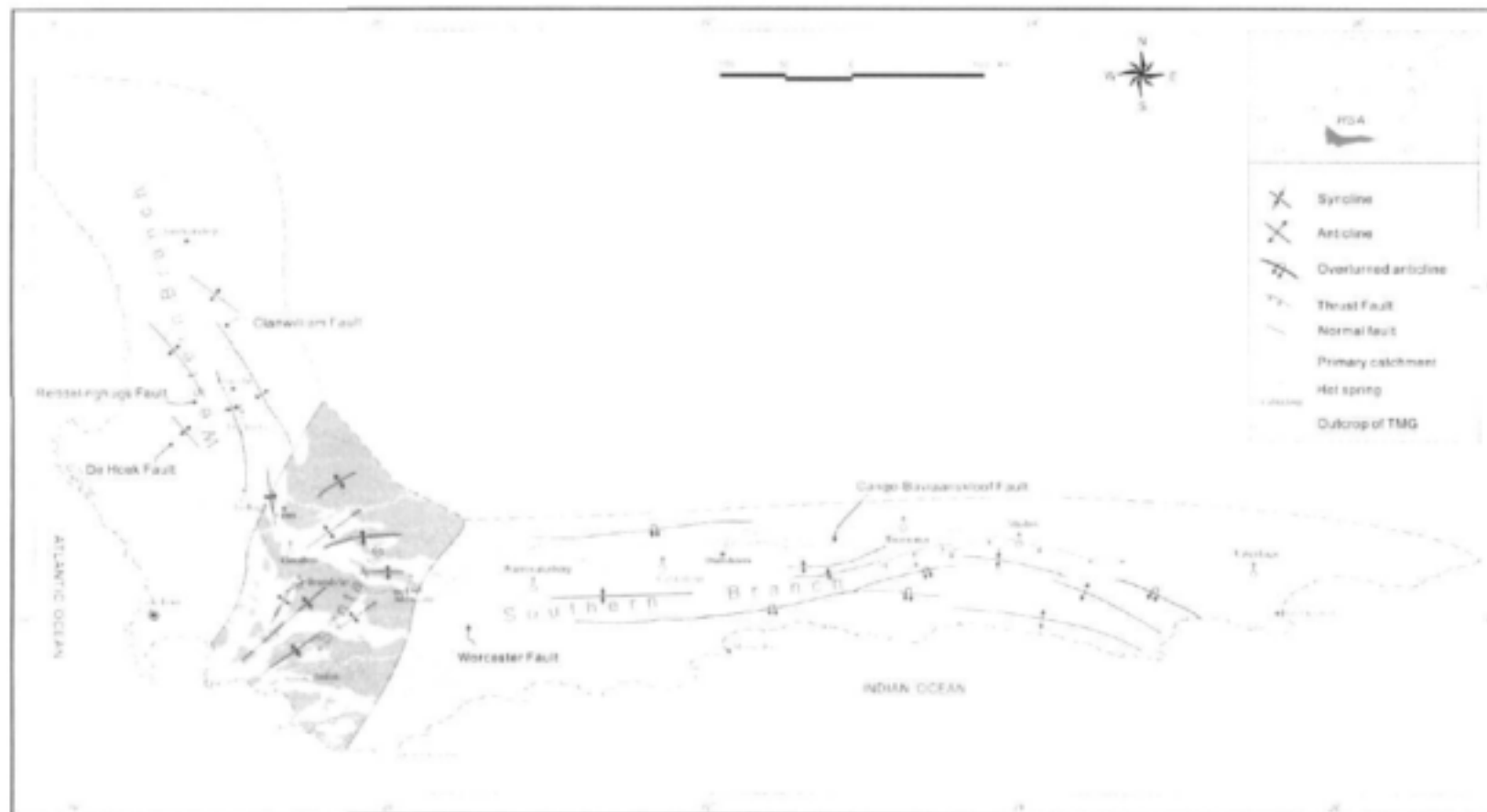


Figure 2.3 Sketch of the structure in the study area
(Faults are adapted from Council for Geoscience, 1997)

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

Several examples of recent tectonic activity (neotectonics) exist in the south-eastern Cape (Andreoli *et al.*, 1989; Hill, 1988; Hattingh and Goedhart, 1997; Hartnady, 1998) and Karoo (Woodford and Chevallier, 1998). The tectonic activity suggests that an extensional tectonic regime continues to prevail, with an extension in a NNE-SSW direction and compression in a WNW-ESE direction. These were associated with the following events:

- Cape Orogeny: N-S, NW-SE, NE-SW, E-W (thrusting) systems.
- Gondwana break-up (extensional tectonics): development of E-W oblique shears.
- Extension with a right-lateral shear component and reactivation of earlier fractures (E-W, WNW, and N-S).
- Tertiary to present time: continuation of the extensional stress regime in a NNE-SSW direction.

The E-W striking system consists of long, continuous E-W and WNW trending faults extending across the entire study area, of which only the most prominent faults, with large displacements, are shown on the 1:250 000 scale geological maps. The satellite lineament map, therefore, provides a much more complete representation of the most predominant fracture sets, which are E-W, WNW and ENE trending fracture sets (more fractures and longer extent) in the east section of the TMG.

The N-S trending fracture system consisted of shorter, more discontinuous fractures, which generally corresponded to a dense network of NNW-SSE, NNE-SSW trending joints (forming a conjugate) and less prominent N-S trending joints, often not showing any displacements on geological maps or satellite images. These fractures or joints can be very penetrative and form prominent morpho-tectonic features.

Extensive field mapping and construction of balanced cross-sections was the only way to determine the abundance of thrust faulting relative to normal faulting. Normal faulting seemed to have controlled the emplacement of dolerite dykes in the Western Karoo and deposition of the Uitenhage Group and even high-level terrace deposits.

North-south trending fractures display large variations in geographic distribution. The Kammanassie and Rooiberg Anticlines have by far the highest density of fractures and joints, of which the majority were most probably related to the Cape Orogeny. The NNW, NNE and NS trending fractures have left a strong overprint over the CFB structures. Some of the lineaments crossing the Uitenhage Group outcrops were older than the fractures.

2.2 Hydrogeology of the TMG

Approximately 37 000 km² of outcrop area of the TMG is spread within an area of 248 000 km² in the Western and Eastern Cape Provinces. The TMG comprises about 4 000 m thick sequence of quartz arenite, which form the TMG Super Aquifer. Information on climate, hydrology, geomorphology and geology are used to develop conceptual models of recharge in the TMG system.

The TMG area has been subdivided into groundwater regions on the basis of both lithostratigraphy and physiography, namely crystalline, metamorphic and igneous, intrusive, extrusive, sedimentary and composite (Vegter, 2001). There are twenty one groundwater regions, in which six regions in the north of the TMG may have hydraulic connectivity at depth, namely at Bushmanland Pan Belt (35), Hantam (36), Tanqua Karoo (37), Western Great Karoo (41), Southern Highland (40) and Eastern Great Karoo (42).

In earlier studies, four sections of the entire TMG Super Aquifer were outlined, based on different thickness of geological successions or their positions in the CFB and on associated tectonic regimes present (Figure 2.4):

- a) The Western TMG Section (A), the CAGE study – Hartnady and Hay (2000).
- b) The Agter-Witzenberg and Ceres-Hex River (B) – Central TMG Section, Rosewarne (1984), Kotze and Rosewarne (1997) and Weaver *et al.* (1999).
- c) The TMG Aquifer in the Klein Karoo Section (C), Kotze (2001).
- d) The Eastern TMG Section, Murray (1996).

The above-mentioned case studies have indicated that significant differences in fracturing and faulting styles, lithological sequences, recharge and natural discharge (springs) prevail in each section of the TMG Aquifer. Informally named aquizones (Al-Aswad and Al-Bassam, 1997; Kotze, 2002), they were made apart from the formal hydrogeological hierarchy.

The interpretation of satellite imagery, structural, hydrochemistry and hydrogeological data were collated to define the conceptual hydrogeological model for the Klein Karoo Section of the TMG aquifer. For delineation of fractured rock aquifer at a local scale, the term "Keystone Block" was put forward by Kotze (2002).

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

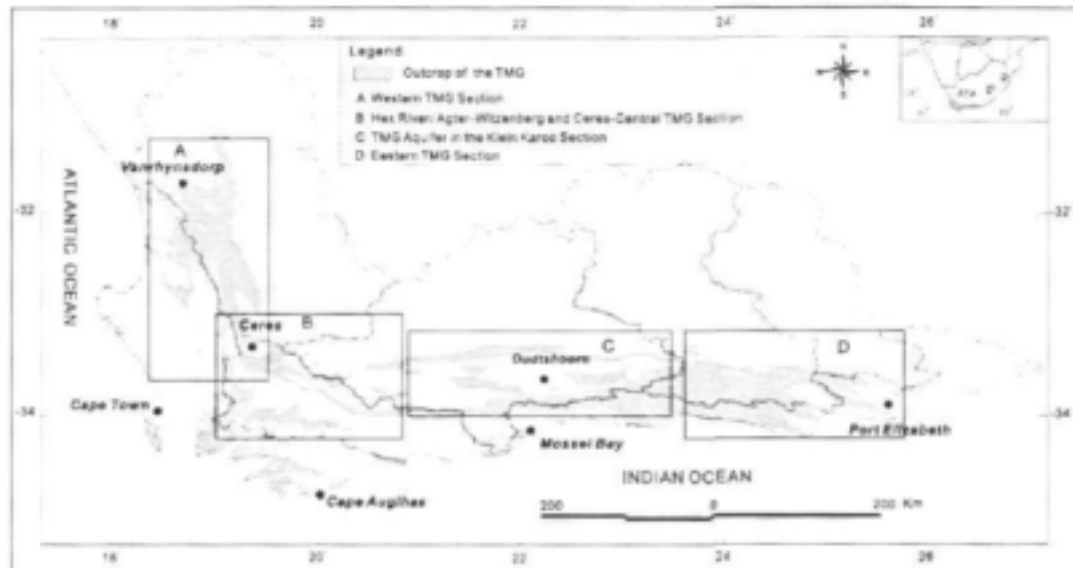


Figure 2.4 Four sections of the TMG aquifer in early studies

2.2.1 Aquifer Systems of the TMG

Generally a hydrodynamic aquifer system has three elements, namely aquifer, aquitard and hydraulically impermeable boundaries, and possesses a recharge area, a flow area and a discharge area as well. In some cases, the discharge area is very close to the recharge area or even overlaps, so the flow path is short.

Folded and fractured rock aquifer systems dominate in the TMG area. The TMG aquifers, i.e. geological units with economically exploitable groundwater quantities, are classified as secondary aquifers. Extensive folding and faulting gave rise to the secondary porosity during several tectonic events. The style of fracturing and folding depends on the matrix and regional tectonics of the fracture pattern. The following TMG aquifer systems are presented in the TMG area:

- Horizontal terrain aquifer system
- Inclined strata aquifer system
- Folded strata aquifer system
- Syncline aquifer sub system
- Anticline aquifer sub system
- Fracture zone aquifer system
- Weathered rock (crust) aquifer system (not discussed here)

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

2.2.1.1 Horizontal terrain aquifer

Two types of aquifer occur in a horizontal terrain, namely perched and unconfined aquifers. An example is Platteklip Gorge Spring as can be seen in Figure 2.5. The issue point of the spring is on an interface of the Peninsula and Graafwater Formation.

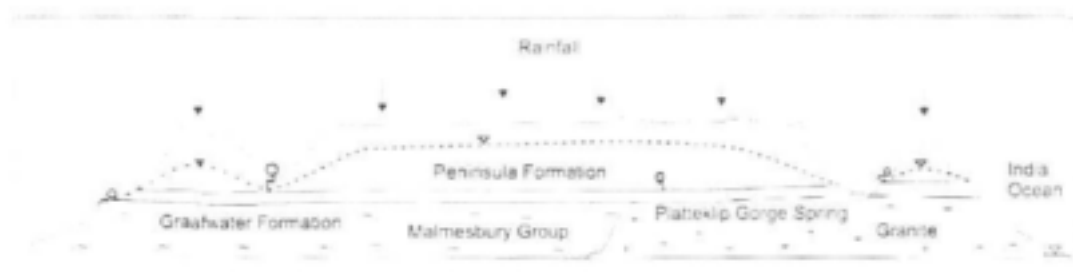


Figure 2.5 Horizontal terrain aquifer in the Table Mountain

2.2.1.2 Folded strata aquifer system

The folded TMG aquifer system includes syncline and anticline sub systems. The syncline sub system located in basin area is a permeable rock layer within low permeable rock layers. Four typical configurations, namely fault-fold-lithology controlled, fold-lithology controlled, leakage and overflow system, are shown in Figure 2.6. The anticline aquifer sub system distributes a core of the anticline with wide-open fractures, such as the Kammanassie mega-fold.

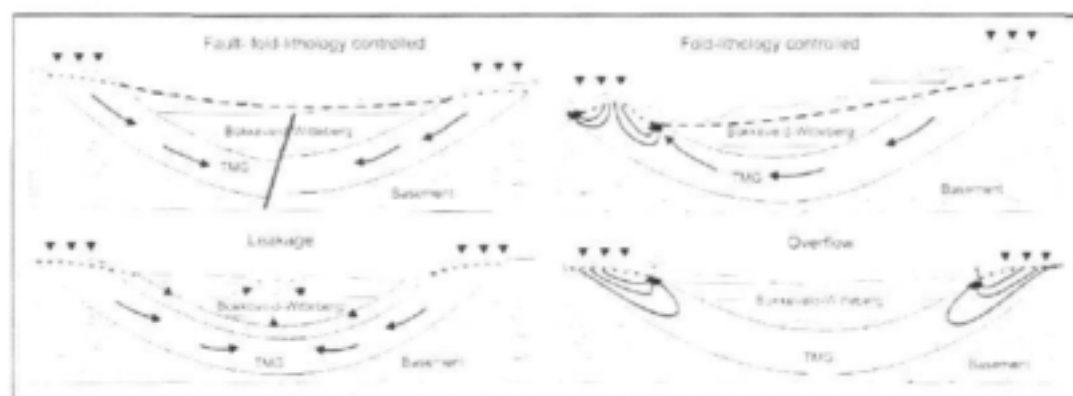


Figure 2.6 Models of syncline aquifer systems

2.2.1.3 Fracture zone aquifer system

According to the hydrogeological function of a fracture, five types are classified as follows (Sheng et al., 1985):

- Fracture: Tensile, the permeability of this fault is larger than the hanging wall and footwall.
- Fracture conduit: Connecting different aquifers, e.g. the Brandvlei Fault.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

- Impermeable fracture, such as Wemmershoek Fault, Worcester Fault, Cango Fault. The permeability of the faults is much smaller than that of the hanging wall and footwall.
- Storage fracture is defined as a closed storage space in a fault zone.
- Water-scarce fracture, a fault with no water-runs where there is no storage space and very low permeability in these fault zones.

The heterogeneous and anisotropic fractured rock aquifer system comprises an open fracture, conduit and hydraulic boundary (impermeable fault) as can be seen in Figure 2.7. The following points are illustrated:

- 1) Fractures are divided into open, closed and infilled fractures.
- 2) Fracture density is related to the thickness of the layer of lithology.
- 3) Conduits, storage as well as exclusion of groundwater can be present in a fracture.
- 4) There are weathered and tectonic fractured rock aquifer systems.
- 5) Hydraulically connected aquifers occur in different aquifers by conduit fractures.
- 6) In a fracture zone aquifer system, groundwater can be obtained at another side of the mountain and the opposite bank of the river as well.
- 7) Cool water occurs in local flow, but thermal water can occur in regional flow, which may flow along the TMG buried by thick younger formations in basin area.

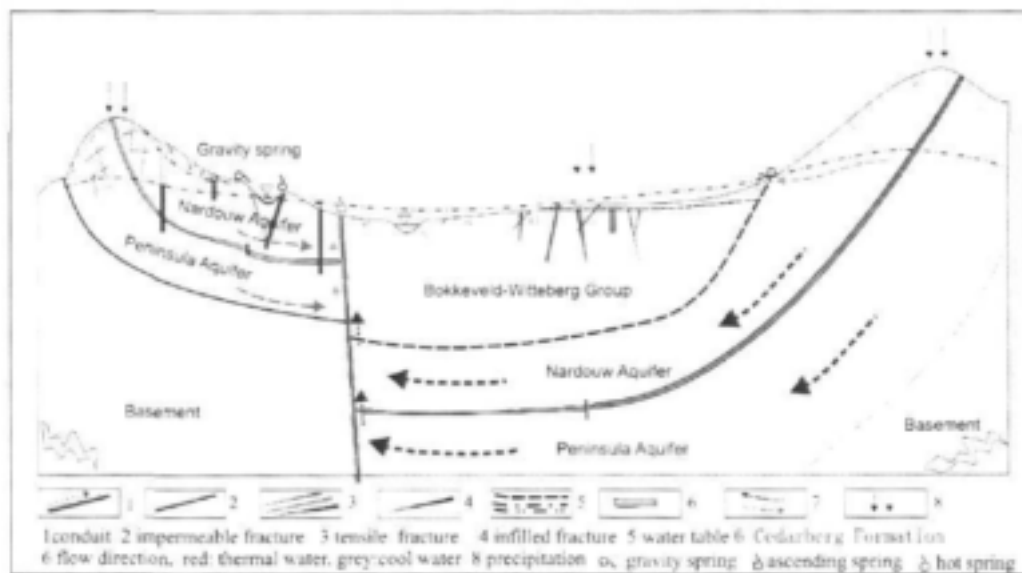


Figure 2.7 A model of the TMG fractured rock aquifer system

2.2.1.4 Composite aquifer system

An independent aquifer system usually occurs locally, but regional aquifer systems are always complex and synthetically produced. Different aquifer systems may connect to each other through the faults or leakage layers. The composite aquifer systems are:

- The monocline -fault aquifer system [Cape Peninsula].
- The syncline-fault aquifer system [Montagu, Witzenberg].
- The anticline-fault aquifer system [Vermaak's River Valley].
- The syncline-anticline-fault (fold-fault) aquifer system [Brandvlei].

2.2.2 Hydrogeological Boundaries of the TMG Aquifer

2.2.2.1 Lithology

Lithology may control springs occurrences due to the impedance or presence of impermeable layers. Springs discharge at contact zones with interbeds and lithological boundaries, such as the Bokkeveld, the Malmesbury, the Kango and the Namap Group and the Cedarberg Formation. There are a few well-known Cedarberg Formation related springs, such as Hoeksberg Spring (south of McGregor), Vermaak's Spring, Marnewicks Spring, the Meirings, the Swartberg Spring (feeding the Dorps River at Price Albert) and the Humansdorp Spring.

2.2.2.2 Faults

A regional fault is normally a structural boundary as well as a hydrogeological boundary. Most of the regional faults are impermeable boundaries, but fault branches are tensile faults, which are excellent infiltration paths.

There are fault-controlled hot springs, occurring from Citrusdal in the west to Uitenhage in the east. The alignment of these springs follows a remarkable curved line, which appears to mimic the shape of CFB (refer to Figure 2.3). The alignment of these springs may be meaningful in terms of the deeper CFB structure (Meyer, 2002). The major fault controlled springs have the following characteristics:

- They all come out of conduits in competent, fractured and fissured quartzitic-sandstone of the TMG, which are in all instances faulted against, or are in contact with the incompetent Cedarberg Formation and Bokkeveld and Uitenhage Groups aquicludes.
- They are hypothermal springs, with water temperatures in excess of 20°C. They are thus all relatively deep circulating springs. The yields range from 9 to 127 l·s⁻¹ with limited seasonal fluctuations.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

- The EC of the groundwater does not exceed $35 \text{ mS}\cdot\text{m}^{-1}$. The dominant chemical determinants of the groundwater are sodium and chloride. Hydroxides of iron and/or manganese, ranging between 0.1 and $2.0 \text{ mg}\cdot\text{l}^{-1}$, have been recorded in most of them.

2.2.2.3 Drainage system

A watershed is usually a groundwater recharge boundary, but such a boundary can be removed by faults cutting through the watershed, causing recharge area extended into a neighbouring surface water catchment. In most cases, primary and secondary catchment boundaries are related to structures or are one of hydrodynamic boundaries in the TMG area. Major rivers are very active hydrogeological boundary, as they are not only the discharge corridors but could also become sources of recharge depending on the river conditions. Most of the perennial rivers occur in the discharge areas of the TMG aquifer, such as the Olifants River (Olifants Doorn area), Great Berg River, Breede River, Gourits River. The Indian and Atlantic Oceans provide an important opportunity for deep groundwater discharge of the TMG aquifer systems. Groundwater near the coastal areas is likely to discharge into the oceans in most of the southern coast from Strand to Port Elizabeth. Some important hydrogeological boundaries that are used in identifying hydrogeological units, namely fault, lithology, rivers and oceans, are listed in Table 2.1.

Table 2.1 Types of hydrogeological boundaries in the TMG area

Boundary	Class	Name / Characteristics
Fault	Regional, >50km,	Clanwilliam Fault, Leipoldtville fault, Redelinghuis Fault, Theerivier faults, De Hoek Fault, Touwsriver fault, Worcester Fault, Cango-Baviaanskloof Fault, Olifants fault system, Suurbraak fault, Outeniqua mountains fault, Genadendal fault, Hawston - Middleton fault Zone, Baviaanskloof Mountains thrust fault etc.
Lithology/ Stratigraphy	Group, Formation	Group: Malmesbury, Kango, Namap, Bokkeveld, Witteberg, Dwyka, Eccca, Beaufort, Uitenhage Formation: Cedarberg
River	Main branch, perennial	Olifans (Western coast), Great Berg, Breede River, Sonderend, Touws, Groot, Gourits, Gamka, Olifans (Calitzsdorp), Kammanassie, Gamtoos, Swartkops.
Drainage	Primary catchment	E, F, G, H, L, M, N
Oceans		Atlantic Ocean and India Ocean

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2.2.3 Hydrogeological Units

Nineteen hydrogeological units were identified in the TMG based on their hydrogeological properties, covering an area of 248 000 km² in the Western and Eastern Cape Provinces (Figure 2.8). The general characteristics of the hydrogeological units are listed in Table 2.2.

Table 2.2 Hydrogeological units in the TMG area

Unit no	Name	Boundary	Characteristics and Stratigraphy
0	Karoo	North Zero deposit boundary of the TMG, South End of the outcrop of the TMG	TMG aquifers covered by younger Formations Recharge from outcrop of the TMG in the south Stratigraphy involved: Df, Dbi, C-Pd, Pt, Ppw, Pr, Pa, Pf, Pwa, Ope, Sn
1	Vanrhynsdorp	North End of outcrop of the TMG, South Clarewilliam Fault, Congo-Baviaanskloof Fault, Baviaanskloof T/Fault	Anticline is dominant Stratigraphy involved: C-P, De, Dw, Tra-Tra, Ope, Sn
2	Clarewilliam	North Clarewilliam Fault, South Leopoldville Fault	Anticline-syncline-anticline-Fault system. The strata are NW-SE trend Stratigraphy involved: Ng, Ckn, De, Dbi, Tra-Tra, Op, Ope, Sn
3	Piketberg	North Leopoldville Fault (NW-SE), South De Hoek Fault (NW-SE)	Anticline-Fault fracture aquifer system Stratigraphy: Np, Npo, Nbi, Ck, Op, Ope, Sn
4	Wellington	North De Hoek Fault (NW-SE), Worcester Fault (NE-SE to E-W) South Moeresburg-Wellington ((NW-SE) - Paarl-Franschoek (NWW-SEE) East Brandvlei (NE-SW)	Fault-anticline aquifer system Stratigraphy involved: Nno, Nbi, Ny, Nko, N-Cw, N-Cm and Ope, Sn
5	Stellenbosch	North Moeresburg-Wellington ((NW-SE) - Paarl-Franschoek (NWW-SEE), South Table Mountain East Franschoek-Stellen Fault (NE-SW)	Fault aquifer system Stratigraphy involved: Cg, N-Ck, N-Cs, Nf, Ck, Nty, Ope
6	Cape Peninsula	Cape Peninsula, Atlantic Ocean	Monocline aquifer system, TMG is located on top of Malmesbury Group (Ccc) and granite
7	Ceres	North Theeriver Fault (NW-SE) and Touwsriver Fault (E-W), South Worcester Fault (NE-SE to E-W), East Hex River Fault (NE-SW)	Fault controlled aquifer system Stratigraphy involved: Npo, Nbi, C-Pd, Dbi, De, Dw, Ope, Sn

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OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES**

Table 2.2: Hydrogeological units in the TMG area (continued)

Unit No	Name	Boundary	Characteristics and Stratigraphy
8	Worcester	North: Worcester Fault (NE-SE to E-W) West: Brandvlei Fault (NE-SW) and Frans-Stell Fault (NE-SW) East: Betty's Bay-Botriver-Berea Fault Zone (NE-SW) and Robertson Fault (NE-SW)	An antic-syn-fault aquifer system. Stratigraphy involved: N-C, NE, Ppw, C-Pd, Ope, Sn
9	Montagu	North: Touwsriver - Herbertsdale Fault Zone (NW-SE) South: Worcester- Swellendam- Herbertsdale Fault Zone West: Hex River Fault (NE-SW)	Fault-syncline aquifer system. Stratigraphy involved: N - C Dbi, Dc, Di, Dw, Ope, Sn
10	Bonnievale	North: Worcester- Swellendam Fault South: Berea-Greyton-Storm Fault (E-W) and Wydgelee Fault. East: Bontebok Fault (NW-SE) West: Robertson Fault (NE-SW)	Fault-syncline aquifer system. Stratigraphy involved: N- C, NE, Dbi, Dw, Je, Ope, Sn
11	Caledon	North: Berea-Greyton - Storm Fault (E-W) and Wydgelee Fault (NW-SE). East: Bontebok Fault (NW-SE) West: Betty's Bay-Botriver-Berea Fault Zone (NE-SW) South: Coast line	Fault controlled aquifer systems. Stratigraphy involved: N- C, Nty, Dbi, Dc, Je, T-Qb, Qst, Ope, Sn
12	Riversdale	North: Worcester- Swellendam- Herbertsdale Fault Zone South: Coast line West: Bontebok Fault (NW-SE) East: drainage boundary	Fault aquifer system and monocline aquifer system. Stratigraphy involved: Dc, Dbi, Dc, Dw, Je, J-K, T-Qb, Ope, Sn
13	Swarthberg	North: End of outcrop of the TMG. South: Ladismith Fault (E-W) and Congo Fault (E-W)	Fault - overturned anticline aquifer. Stratigraphy: Nk, Dt, Je, Ope, Sn
14	Oudshoorn	North: Ladismith Fault (E-W) and Congo Fault (E-W). Northwest: Touws Fault zones Northeast: Baviaanskloof Fault zone South: watershed	Fault- fold aquifer system. Stratigraphy involved: Dbi, Dc, Dt, Dw, J-K, Kb, Ope, Sn
15	George	North: Watershed, South: Coast line, West: Drainage, East: Outeniqua Fault zone	Fault-monocline aquifer system Stratigraphy involved: N- Cg, Nka, Tg, T-Qb, J-K, Ope, Sn
16	Sturis	North: End of outcrop of TMG (Baviaanskloof Trust Fault). South: Kouga Fault zone	Fault- syncline aquifer system. Stratigraphy: Nk, Je, Ope, Sn
17	Humansdorp	North: Kouga Fault zone South: Coast line	Fault-anticline aquifer system. Stratigraphy: Dc, Je, J-K, Ope, Sn
18	Uitenhage	North: End of the outcrop of the TMG. South: Coast line	Fault-anticline aquifer system. Stratigraphy: Nga, Ks, Qsc, Ope, Sn

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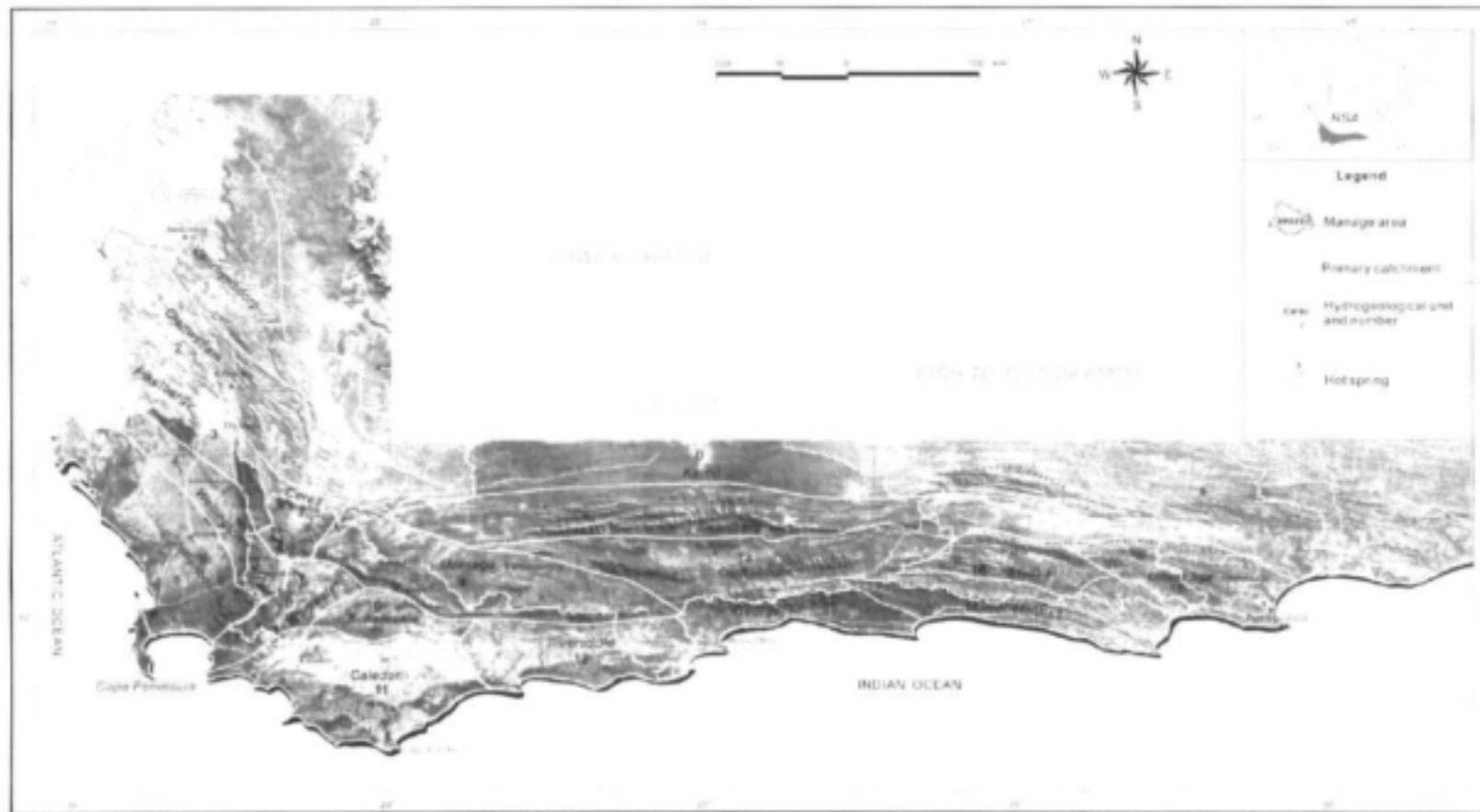


Figure 2.8 Hydrogeological units in the TMG area

2.2.4 Hydrogeochemical Characteristics of the TMG Aquifer System

The statistical analysis was based on chemical data of 98 samples distributed throughout the whole TMG area (collected from NGDB and Vegter, 1995 a, b), in which 89 samples were taken from the Peninsula Aquifer. The major ions in the groundwater are shown in Table 2.3. A predominantly Na-Cl type water occurs based on the Piper hydrogeochemical classification system (Figure 2.9).

Table 2.3 Water quality in the TMG aquifer system

(unit: mg l⁻¹; EC: mS m⁻¹)

Item	EC	TD	pH	K ⁺	Na ⁺	Ca ²⁺	Mg ²⁺	F ⁻	Cl ⁻	HCO ₃ ⁻	NO ₃ ⁻	SO ₄	PO ₄	SiO ₂
Minimum	5.0	21.0	4.2	0.1	3.9	0.5	0.7	0.01	7.4	0.4	0.01	0.2	0.01	4.9
Average	14	83.4	6.0	2.0	16.8	3.8	2.6	0.15	28.6	16.14	1.88	5.4	0.03	13.4
Maximum	53	189	7.6	10.0	46.0	18.0	7.1	0.49	82.0	85.44	17.90	18.9	0.16	37.9
Stand. dev	8.4	42.6	0.8	2.5	9.6	4.0	1.5	0.13	17.1	17.9	3.42	4.5	0.03	8.4

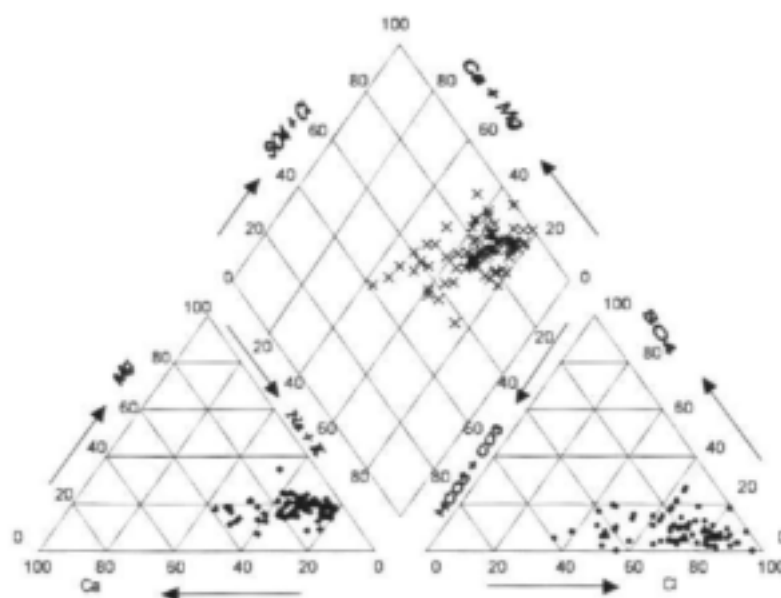


Figure 2.9 Piper diagram of hydrochemistry in the TMG aquifer system

2.2.4.1 Chloride concentration in the TMG aquifer

The results of analyses of 98 water samples were used to calculate the background values of chloride in the TMG aquifers. Figure 2.10 shows the trend of chloride concentration. Most of them are less than 40 mg l⁻¹. The distribution of the concentration is shown in Figure 2.11. The frequency is 32.7%, 24.5%, 15.3% for concentrations of 10-20 mg l⁻¹, 20-30 mg l⁻¹ and 30-40 mg l⁻¹, respectively. The frequency is 8.2% for both 50-60 mg l⁻¹ and 60-70 mg l⁻¹ ranges. The concentrations can be grouped into two levels, that is 10-40 mg l⁻¹ and 50-70 mg l⁻¹.

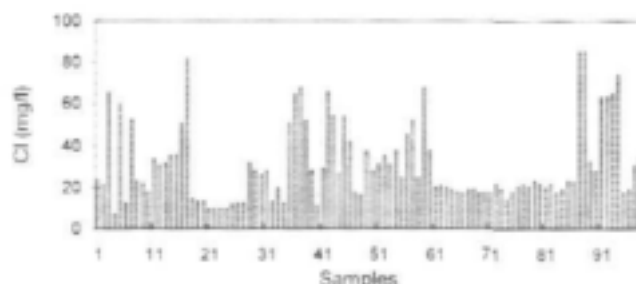


Figure 2.10 Chloride concentrations in the TMG aquifer

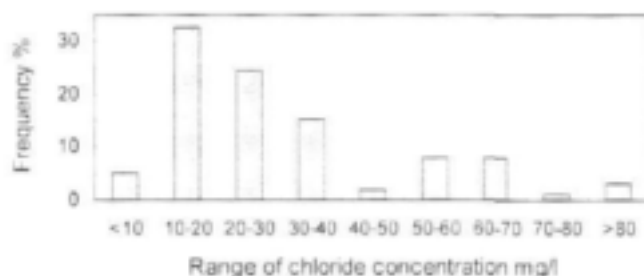


Figure 2.11 Frequency of chloride concentrations in the TMG aquifer

The statistical analysis results of the 98 chloride concentrations are listed in Table 2.4. The discrepancies between arithmetic and harmonic, arithmetic and geometrical means are 38% and 18% respectively. Therefore errors would be introduced in the recharge estimation using the different mean values. There are large errors when the resultant concentration lies between $\bar{x}' + \sigma$ to $\bar{x}' + 3\sigma$, so the chloride mass balance method cannot be used to estimate the recharge rate under these conditions.

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Table 2.4 Chloride concentrations of the TMG aquifer (mg l^{-1})

Type	Mean	$\bar{x} + \sigma^{1)}$	$\bar{x} + 2\sigma$	$\bar{x} + 3\sigma$	$\bar{x} - \sigma$
Arithmetic	30.9	50.1	69.1	88.1	7.12
Geometrical	26.2	45.2	64.3	83.4	11.9
Harmonic	22.5	41.6	60.6	79.7	3.43

¹⁾ σ - standard deviation

The population mean and the interval of the chloride concentration for 95% and 90% confidence are listed in Table 2.5. The table shows that the values of confidence of chloride concentration increases with decreasing sample size. For instance, the interval is 22.4-29.9 mg l^{-1} for a sample size of 100, but 4.56- 47.8 mg l^{-1} for a size of 3. In other words, the error $[(\text{max-min})/\text{max}]$ is 25% when the sample size is 100, but the error is 90% when the sample size is 3.

Table 2.5 Confidence of the chloride concentration

Conditions	$\alpha=0.05, \sigma=19.1, \text{mean}=30.9$						$\alpha=0.1, \sigma=19.1, \text{mean}=30.9$					
Sample size	100	50	20	10	5	3	100	50	20	10	5	3
Confidence	3.74	5.29	8.37	11.8	16.7	21.6	3.14	4.44	7.02	9.93	14	15.9
Maximum (mg l^{-1})	29.9	31.5	34.5	38	42.9	47.8	29.3	30.6	33.2	36.1	40.2	42.1
Minimum (mg l^{-1})	22.4	20.9	17.8	14.3	9.44	4.56	23	21.7	19.2	16.2	12.1	10.3

2.2.4.2 Chloride in the other aquifers of the study area

In comparison to chloride concentration in the TMG aquifer, the chloride concentrations in the other aquifers of the study area are illustrated in Table 2.6. Large differences of the concentration occur between the TMG aquifer and the aquifers around the TMG due to lithology. A very low recharge rate would be produced based on the chloride concentration presented in Table 2.6 if equation (3.1) is adopted. This may be the reason why the CMB method cannot be used to estimate the recharge rate in the aquifer with high background concentration values of chloride, which comes from multiple sources. A contour map of chloride concentration of rainfall is constructed as shown in figure 2.12. If the chloride concentration of rainfall is taken from the map and the concentration of groundwater taken from the Table 2.6, large difference in recharge rate should occur in the same contour of chloride concentration due to different chloride concentration of groundwater in the aquifers. It is worth mentioning that the contour of chloride concentration of rainfall is not quite accurate.

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Table 2.6 Cl concentration in the surrounding aquifers of the TMG

Aquifer (Group or Formation)	Maximum (mg l ⁻¹)	Minimum (mg l ⁻¹)	Mean (mg l ⁻¹)	Sample size
Malmesbury	3196.0	37.0	636.6	9
Cape Granite	688.0	86.0	365.6	5
Bokkeveld	4393.0	112.0	1092.7	9
Witteberg	1638.0	71.0	595.4	7
Dwyka	1984.0	53.0	673.0	7
Ecce	1550.0	104.0	519.2	6
Beaufort	394.0	67.6	249.3	6
Saarberg	864.0	829.0	846.5	2
Unenhage	2537.0	336.0	1497.3	6
Algoa	856.0	615.0	735.5	2
De Hoopvlei	150.0	140.0	145.0	4
Coastal sand	745.0	106.4	431.4	4
Alluvial	2400.0	16.0	656.7	7

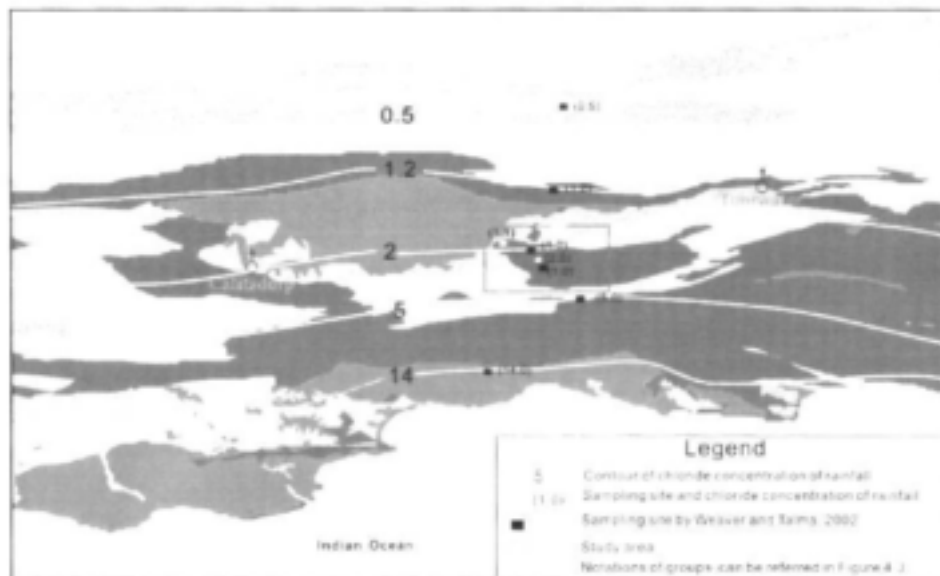


Figure 2.12 Contour of Cl conc of rainfall in the Kammanassie area

2.3 Role of Geomorphology in Recharge Processes

Geomorphology plays a vital role in the formation of groundwater recharge. The recharge areas of the TMG aquifer lie primarily in the rugged mountainous areas with altitudes ranging from 200 m to 2 250 m. To the northeast of the mountains lie the relatively high altitude plateau of the Karoo at an elevation of between 800 m and 2 000 m. In the west section, the NW-SE trend of the mountains run parallel to the coastline and consist of the Bokkeveld, Piketberg, Cedarberg, Skurweberg, Cold Bokkeveld and Witsenberg Mountains. In the syntaxis area, the ranges strike northeast, such as the Hex River Mountains, Riversonderend and Overberg Mountains. The Swartberg Mountains lie in the north, and the Langeberg Range and Outeniqua Mountains in the southern E-W striking range, which runs parallel to the coastline with the Kammanassie Mountain range, lying in the east between them (refer to Figure 54). The topography of the eastern section is, to a large extent, characterised by a number of NW-SE striking mountain ranges and ridges, such as the Kouga, Tsitsikamma-Kareedouw, Baviaanskloof, Grootrivier Mountain and Great Winterhoek Mountains. The interaction of climate, geology, morphology, soil condition, and vegetation determines the recharge process. An understanding of the geomorphology contributes to a better delineation of the recharge processes of the TMG aquifer system and an improved appreciation of recharge estimation. Field investigations were done to develop conceptual models of recharge processes in the TMG area.

2.3.1 Topography

The topography of the TMG aquifer is controlled by lithology and structure. The mountain ranges are consistent with the orientation of faults. Gentle slope topography is characterized in southern Grabouw. Gorge topography predominates along river valleys in the outcrop areas of the TMG. Fault controlled topography is typified by mega faults, such as the De Hoek Fault, Worcester Fault and Cango Fault.

Four types of topography, controlling different recharge processes, are identified in the TMG area, namely mesa (Figure 2.13), cuesta (Figure 2.14), fold-controlled topography (Figure 2.15) and fault-controlled topography (Figure 2.16). Vertical infiltration controls the recharge process in the mesa area. Percolation occurs along the bedding fractures and recharges vicinity catchment in the cuesta area. Recharge processes are more complex in the folded and faulted rock area where more effective recharge may occur in the area with high density of fractures and percolate to local or other catchment via fracture networks.



Figure 2.13 Horizontal topography (mesa) in the Table Mountain



Figure 2.14 Cuesta (incline) topography in the Rooiels Bay



Figure 2.15 Fold controlled topography in Montagu



Figure 2.16 Fault controlled topography in Rawsonville

2.3.2 Influence of Fractures on Recharge

The processes of recharge are specifically associated with open fractures. These fractures, when near the surface, are often totally or semi-infilled by soil and vegetation. A typical example is the fractures in the Table Mountain (Figure 2.17). In this figure the ongoing infiltration process is observed through bubble formation on the water surface. If the infiltration rate is smaller than the intensity of rainfall, surface flow occurs.

High infiltration rates may occur in non-filled fractures. The infiltration may be influenced by soils due to reducing aperture of fractures and vegetation due to increasing evapotranspiration in semi-filled fractures. A good example is the Blikhuis area shown in Figure 2.18.

The outcrops are highly fractured in most areas of the TMG. Stress relief fractures are often seen in the near land surface, especially in steep slopes. Rainfall can result in very high infiltration rate near land surface, but the fractures tend to close at depths, therefore, most of infiltrated water comes out as interflow. An example is the Vermaak's Valley (Figure 2.19).

Preferential fractures are linked to the aquifer directly even though some paleo-fractures are sealed. Young fractures are generally not cemented. Examples of these fractures in Rooi's Bay are shown in Figure 2.20. The fractures at Dunn's Castle are also typical examples (Figure 2.21). The tectonic fractures together with weathering and stress relief fractures can form the preferential fracture zones, in which effective recharge will be increased. This type is characterised at Platteklip Gorge (Figure 2.22). The open fracture density near the surface is the key factor controlling the infiltration rate, but recharge is constrained by characteristics of fractures at depth. The wide-open fractures not only enhance rainfall infiltration but also form preferential recharge.

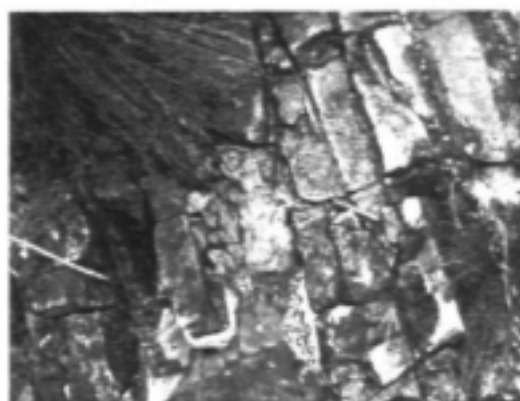


Figure 2.17 Fractures clogged with infillings near the surface in the Table Mountain

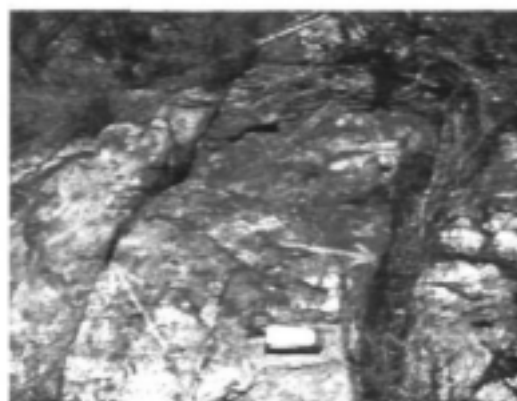


Figure 2.20 Open and semi infilled fractures in the Blikhuis



Figure 2.18 Stress-relief fractures in the Vermaak's valley

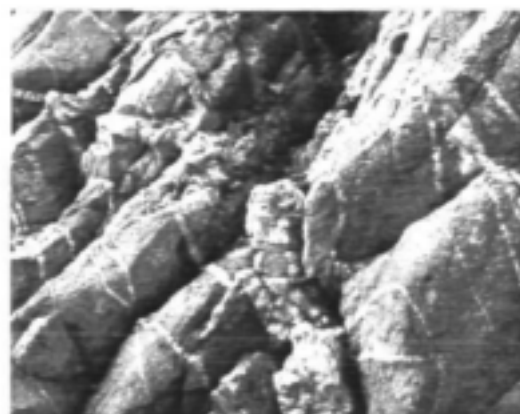


Figure 2.21 Fissure without displacement in the Rooi els Bay

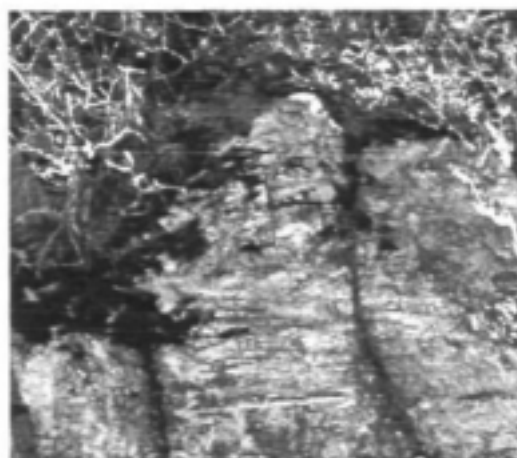


Figure 2.19 Preferential fractures at Dunn's Castle

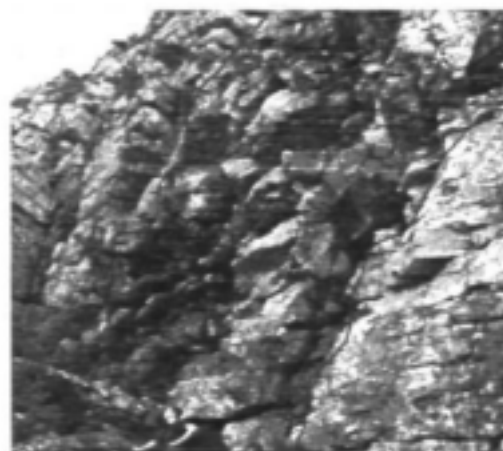


Figure 2.22 Preferential fracture zone at Platteklip Gorge, Cape Town

2.3.3 Impact of Slopes on Recharge

2.3.3.1 Slope type

Recharge is related to the slope of mountains. For instance, rainfall has sufficient time to infiltrate into soils and rocks if the slope is gentle; rainfall is unlikely to infiltrate effectively if the slope is steep. The relationship between the slope and the dip of strata, together with the occurrence of fracture networks, is the main factor influencing the recharge rate. A gentle slope with the strata dipping away is good for recharge if the slope angle is less than the dip angle (Figure 2.23), but an escarpment with the strata dipping in the opposite direction hampers recharge, which is illustrated by the snow cover on the slope in Boland area. Snow accumulates on a moderate slope, but is hardly retained on steep slopes (Figure 2.24). A slope surface with strata forming the surface is the worst-case scenario of recharge if there are no open fracture networks for infiltration. The Ladismith area shown in Figure 2.25 is a typical example.



Figure 2.23 Moderate slope near the Theewaterskloof Dam



Figure 2.24 Slope and snow in Boland (Cape Nature Reserve II)



Figure 2.25 Layer slope in Ladismith

2.3.3.2 Form of slopes

Estimating the impact of the concavity, convexity, regularity or warp of a slope on recharge is a very delicate procedure. These factors are often neglected giving rise to conflicting results. The longer the slope the more rainfall that will infiltrate on a convex slope. The outcrops of the TMG usually have a convex slope; an example is in Citrusdal (Figure 2.26). The presence of concave slopes in a landscape indicates that there must be trapping, siltation and colluvial deposits in the valley. A concave ground surface or slope surface can often hold a certain amount of water after rainfall. For instance, it can hold water in hollows such as on the top of the Table Mountain (Figure 2.27). It is advantageous for infiltration if micro fractures occur under hollows that are often filled with water.



Figure 2.26 Convex slopes in Citrusdal



Figure 2.27 Water (brown) held in hollows on the top of Table Mountain

As the gradient increases, the transporting water accelerates towards the foot of the slope. This occurs when the slope (S) exceeds 15% (Roose, 1996). Roose (1996) also mention that runoff does not necessarily increase with slope. In general, there are mesa on the top of mountain and long cuesta with steep transection valley in the outcrop of the TMG area. A gentle watershed occurs in most of the outcrop area of the TMG in the western and eastern section, for example, the Witteberg Mountain (Figure 2.28). In the syntax area, cliffs and narrow valleys are dominant; the slope of the banks of valleys are greater than 30° in most cases, but slopes are gentle on the foot of the valleys (Figure 2.29). Slopes are wide on the gentle side where they coincide with strata dip and are narrow on the steep side where they are opposite to the strata dip in the cuesta. A typical example is the Langeberg Range (Figure 2.30). More complex slopes occur in some areas, such as the Swartberg Pass (Figure 2.31). In the hydrogeological unit 3 and 6, there is a broad and gentle land surface with an almost horizontal dip of strata. Based on the slope character, three grades are classified, namely $10-20^\circ$, $20-30^\circ$ and $>30^\circ$. More effective recharge occurs if the slope is less than 30° (gentle) in the outcrops of the TMG area.



Figure 2.28 Slope gradient in the Witteberg Mountain



Figure 2.29 Slope gradient in the Hex River Mountain

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Figure 2.30 Slope gradient in the Langeberg Mountain



Figure 2.31 Slope gradient in the Swartberg Mountain

2.3.4 Influence of Vegetation on Recharge

2.3.4.1 Vegetation types

Influence of vegetation on recharge can be observed in many ways, such as interception and transpiration. The vegetation can consume water from deep soils and fractures. Three types of vegetation are dominant in the recharge area of the TMG aquifer, namely indigenous Mountain Fynbos, exotic plants and sparse vegetation. The most common of which is the Fynbos, an example of which can be seen in Figure 2.32. The sparse vegetation can often be seen in the outcrop of the Skurweberg Formation, an example of which can be witnessed in the Mitchell's pass area (Figure 2.33). Exotics, such as *Pinus* and *Eucalyptus* species are planted at the foot of the mountains or gentle slopes (Figure 2.34 and Figure 2.35). Due to natural spreading some areas later become invaded. For the outcrops of the TMG, there is a high vegetation cover rate on the gentle slopes and sparse vegetation on the steep slopes because of the soil thickness.



Figure 2.32 Mountain Fynbos in the Mont Rochelle Nature Reserve



Figure 2.34 Sparse vegetation in the Mitchell's Pass



Figure 2.33 Exotic-*Pinus* in Grabouw



Figure 2.35 Exotic-*Eucalyptus* in the Sir Lowry's Pass

The first robust estimate of the impacts of plantations at a national scale, was based on the forestry impact curves and catchment mean annual runoff (MAR) calculated by HRU (1981). Plantations were estimated to be using an additional $1.284 \times 10^6 \text{ m}^3$ of water in 1980, which is approximately 7.9% of the usable portion of total water resources (DWA, 1986). Given a total

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plantation area of 1.13×10^6 ha (DWA, 1986), an average reduction in annual runoff from afforested land of 113.6 mm (Scott *et al.*, 1998).

The estimated impact of invading exotics on the water resources of South Africa, at approximately 6.7% of the total surface runoff, is substantial. Even at a conservative rate of new invasion of 5% per year, the size of the invaded area could double in 15 years, potentially doubling the water use (Versfeld *et al.*, 1998). There is clear evidence that exotic species will increase in extent and density if given time and sufficient area to invade to use an even greater proportion of the available water in catchment areas (Le Maitre *et al.*, 2000).

About 10 million ha (8.28%) of South Africa's land area have been invaded to some degree by a wide range of exotic species. If the invaded area is 'condensed' to adjust to 100% of the cover, then the equivalent of approximately 1.7 million ha (1.39%) has been fully invaded. The invaded area in the TMG area is presented in Table 2.7.

Table 2.7 Areas invaded by exotic plants in the different primary catchments
(After Le Maitre *et al.*, 2000)

Primary catchment	Drainage system	Area (ha)	Total area invaded area		Condensed invaded area	
			(ha)	%	(ha)	%
E	Olifants (W Cape)	4 906	528 972	10.78	37 623	0.77
G	Agulhas coast	2 524	1 597 036	63.26	384 636	15.24
H	Breede, SW Cape	1 551	741 408	47.78	84 398	5.44
J	Gouritz	4 513	711 436	15.76	59 399	1.32
K	S Cape coast	716 816	132 030	18.42	52 993	7.39
L	Gamtoos	3 473	277 428	7.99	34 289	0.99
M	Port Elizabeth region	261 156	70 376	26.95	11 358	4.35
N	Sundays	2 122	25 191	1.19	3 964	0.19

The top 10 invading species or groups of species in South Africa ranked by condensed invaded area are listed in Table 2.8. Habitat indicates the main habitats invaded by the species, which are: landscape (l), riparian (r) and alluvial plants (r(a)). The condensed area is the total area adjusted to bring the cover to the equivalent of 100%. Density is the estimated mean cover over the total invaded area.

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Table 2.8 Top 10 invading species or groups of species in South Africa
(ranked by condensed invaded area. After Le Maitre et al., 2000)

Species	Habitat	Condensed	Total invaded	Density
<i>Acacia Cyclops</i>	l	339 153	1 855 792	18.28
<i>Prosopis</i> spp.	R (a)	173 149	1 809 229	9.57
<i>Acacia mearnsii</i>	r, l	131 341	2 477 278	5.30
<i>Acacia saligna</i>	l, r	108 004	1 852 155	5.83
<i>Solanum</i>	r, l	89 374	1 760 978	5.08
<i>Pinus</i> spp.	l	76 994	2 953 529	2.61
<i>Opuntia</i> spp.	l	75 356	1 816 714	4.15
<i>Melia azedarach</i>	r l	72 625	3 039 002	2.39
<i>Lantana camara</i>	r	69 211	2 235 395	3.10
<i>Hakea</i> spp.	l	64 089	723 449	8.86

The impact of invading exotic plants on mean annual runoff is 5.58 % and 15.82% of MAP for water use in the Eastern and Western Cape Provinces, respectively. The impact of water use by invading exotic plants on the mean annual runoff (MAR) in primary catchment areas of the TMG is present in Table 2.9 The highest impact appears on the Agulhas coast in the hydrogeological unit 11.

Table 2.9 Impact of water use by invading exotic plants on MAR in primary catchment areas of the TMG
(After Le Maitre et al., 2000)

Primary Catchment	Drainage system	MAR invaded ($\times 10^6 \text{ m}^3$)	Condensed water use area (ha)	Incremental use ($\times 10^6 \text{ m}^3$)	Water use MAP (mm)	Reduction Equivalents (mm)
E	Olifants Sout, Doring	1 008.35	37 623	35.52	3.52	94
G	Agulhas coast	2 056.75	384 636	646.50	31.43	168
H	Breede, Riversdale	2 088.35	84 398	181.63	8.70	215
J	Gouritz	670.63	59 399	74.79	11.15	126
K	S Cape coast	1 297.30	52 993	134.46	10.36	254
L	Gamtoos	494.71	34 289	96.53	19.51	282
M	PE Coast Swartkops,	150.04	11 358	40.18	26.78	354
N	Sundays	279.89	3 964	8.34	2.98	210

The ranks of the invading species are presented in Table 10. As a group the acacias, with seven species in the top 25 -notably *A. mearnsii*, *A. cyclops* and *A. dealbata* - are the prime water users, accounting for 55% of the total water used. They are followed by the *Pinus*, *Eucalyptus*, *Prosopis* species and *Melia azedarach*. The top ten accounts for 81% of all water used by invading exotics. The impact of the latter species (11-25 in Table 2.10) on surface runoff is limited because of its coastal distribution pattern, except where communities depend on groundwater supplies, as in the case of the town of Atlantis in the Western Cape (Le Maitre et al., 2000).

The effects on stream flow by *E. grandis* were assumed to be representative of all eucalyptus species planted in plantations at present. An analysis of some measurements of transpiration by poplars suggested that their annual water use was similar to pines (Scott *et al.*, 1998). Nevertheless, because all hardwoods (e.g. poplars and oaks) were grouped with eucalypts in the National Forestry Coverage Map, the flow-reduction curves for Eucalypts were used for all hardwoods except Wattles. The inclusion of Poplars and other species under Eucalypts is unlikely to have a significant impact on the estimates because the areas planted with these species only amount to about 0.6% of the total afforested area (DWAF, 1996). The impact of Eucalyptus and Pines on flow is illustrated in Figure 2.36. An adult tree would use more water and reduce flow more.

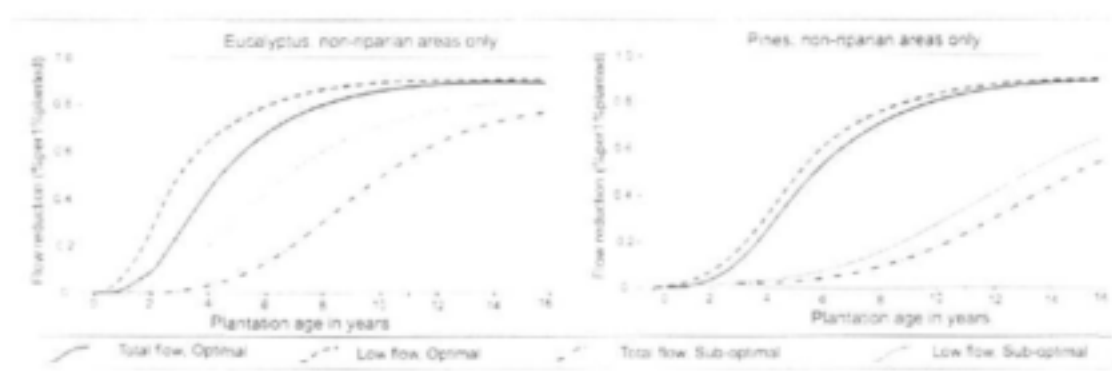


Figure 2.36 Flow reduction curves for Eucalyptus and Pines (Scott *et al.*, 1998)

Several studies concentrated on the impact of the exotic species on runoff, but the impact of the exotic species on recharge has not yet been discussed. Although the invading exotic plants consume large amounts of water in the whole TMG area, the cover rate and density of the exotic plants in the outcrop of the TMG is still small. From the water balance point of view, the exotic plants consume more water than indigenous plants and this can reduce the recharge rate as well. Further studies of the impact of exotic species on recharge are necessary.

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Table 2.10 Rank of the mean annual water use of invading species in South Africa
(After Le Maitre et al., 2000)

No.	Species	Water Use (millions of m ³)	Cumulative percentage of the total water use
1	<i>Acacia mearnsii</i>	576.58	17.46
2	<i>Acacia Cyclops</i>	487.63	32.22
3	<i>Acacia dealbata</i>	248.32	39.74
4	<i>Acacia mixed spp.</i>	242.63	47.08
5	<i>Pinus spp.</i>	231.53	54.09
6	<i>Eucalyptus spp.</i>	213.98	60.57
7	<i>Prosopis spp.</i>	191.94	66.38
8	<i>Acacia saligna</i>	171.13	71.56
9	<i>Melia azedarach</i>	164.91	76.56
10	<i>Solanum</i>	139.97	80.79
11	<i>Lantana camara</i>	97.14	83.73
12	<i>Chromolaena</i>	68.26	85.80
13	<i>Hakea spp.</i>	66.30	87.81
14	<i>Populus spp.</i>	53.83	89.44
15	<i>Jacaranda</i>	48.40	90.90
16	<i>Sesbania punicea</i>	42.57	92.19
17	<i>Rubus spp.</i>	41.33	93.44
18	<i>Acacia longifolia</i>	38.73	94.62
19	<i>Psidium guajava</i>	37.31	95.75
20	<i>Caesalpinia</i>	33.82	96.77
21	<i>Salix spp.</i>	33.21	97.78
22	<i>Acacia melanoxylon</i>	32.20	98.75
23	<i>Acacia decurrens</i>	9.83	99.05
24	<i>Quercus spp.</i>	7.24	99.27
25	Other spp.	17.27	100.00

2.3.5 Structural Influences on Recharge

Recharge processes are related to topography, structure and texture of soils and rock fractures and rainfall patterns in the TMG area. The rainfall intensity is one of the important controlling factors. Infiltration is reduced due to the surface sealing effect resulting from high rainfall intensities and rain splash. The infiltration depth is slightly deeper along the roots of plants and macro pores. The soil moisture is also one of the factors which influence the infiltration depth. Based on geomorphology and hydrogeology, six models of the recharge processes in the TMG area are conceptualised, namely (1) recharge processes in soil zones, (2) recharge processes in horizontal strata zones, (3) recharge processes in upright strata zones, (4) recharge processes in inclined strata zones, including three different cases.

2.3.5.1 Recharge processes in soil zones

The recharge process in zones with soil cover is depicted in Figure 2.37. This model contains:

- 1) The unsaturated zone, which is composed of soil and rock materials. The infiltration processes in the unsaturated zone are much more complex. Four zones can be characterised: the leach zone, the illuvial zone, the root zone and a no impact zone. Moisture in the illuvial and root zones is lost by direct evaporation or transpiration of plants.
- 2) Rainfall which contributes to both runoff and infiltration. During a rainfall event the rate of precipitation can exceed the infiltration capacity of soils, in which case excess water will pond on the soil surface. It is the ponded water that is available for overland flow to surface streams.
- 3) The concentration of runoff that varies from place to place, depending upon the surface environments. Its concentration C_r is greater than C_i .
- 4) The concentration of infiltration water noted as C_1 , which is greater than C_r . When the infiltration water emerges as interflow, its chloride concentration is termed C_{11} with $C_{11} \geq C_1$.
- 5) The concentration in the illuvial zone C_2 , with $C_2 > C_1$. When the infiltration water emerges as interflow, its chloride concentration is termed C_{21} with $C_{21} \geq C_2$. There is a maximum concentration (C_{21}) in the scenario.
- 6) The concentration of chloride from the illuvial zone is C_{22} with $C_{22} \geq C_2$. When the infiltration water emerges as interflow, its chloride concentration is termed as C_{22} .
- 7) The infiltration water recharges groundwater with concentration C_{22} .
- 8) The chloride concentration of interflow (C_{in}) should satisfy

$$C_{in} \times V_{in} = C_{11} \times V_{11} + C_{21} \times V_{21} + C_{22} \times V_{22}$$

$$V_{in} = V_{11} + V_{21} + V_{22}$$

The concentration of surface water should satisfy:

$$C_s \times V_s = C_{in} \times V_{in} + C_{01} \times V_r$$

$$V_s = V_{in} + V_r$$

The mixture concentration of groundwater should satisfy

$$C_g \times V_g = C_{g0} \times V_{g0} + C_{22} \times V_{22}$$

C_{g0} is the chloride concentration of old storage water in aquifer.

The subscript of volume (V) is corresponding to the components of concentration.

- 9) The infiltration water, which temporarily stayed in the unsaturated zone, would have a constant concentration. This water may be flushed into the water table by sequential infiltration events.
- 10) In any given topography, only one or two types of interflows can prevail.

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- 11) The concentration C_{21} is similar to C_{22} in most scenarios because it originates from the same zone.

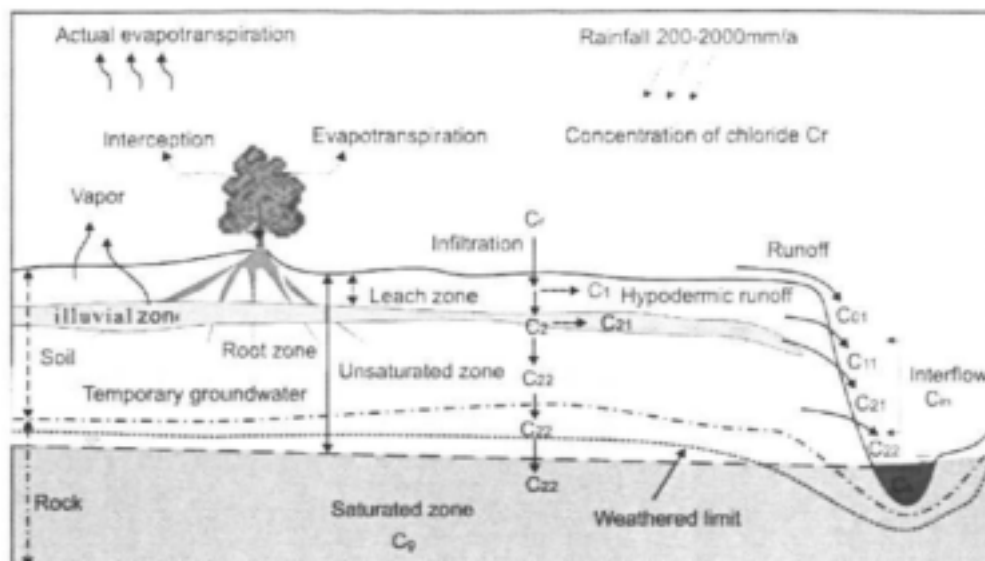


Figure 2.37 Sketch of recharge processes: case 1 with soil cover

C_r - chloride concentration of rainfall

C_s - chloride concentration of surface water

C_g - chloride concentration of groundwater

C_{in} - chloride concentration of interflow

C_{01} - chloride concentration of runoff (sheet flow)

C_1 and C_{11} - chloride concentration of hypodermic runoff (Hypodermic runoff is one of the runoff within the loosen materials near the land surface, such as worm pipe etc.)

C_2 - illuvial (salt deposit) zone

C_{21} - interflow from illuvial zone

C_{22} - chloride concentration of infiltrating water



The model depicted in Figure 2.37 would occur mostly in valleys in the TMG area. For example, the Vermaaks valley fits the model as shown in Figure 2.38. The seepage zone with growing grass is the start point of the Vermaaks valley.

Figure 2.38 Interflow originated from soils zone in the Vermaaks valley

2.3.5.2 Recharge processes in horizontal strata zones

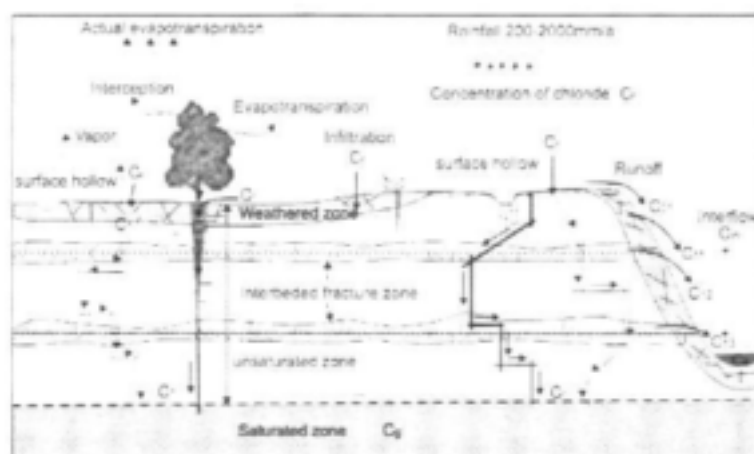


Figure 2.39 Recharge processes: case 2 with horizontal strata

- C_i - chloride concentration of rainfall
- C_s - chloride concentration of surface water
- C_g - chloride concentration of groundwater
- C_e - chloride concentration of sheet flow
- C_m - chloride concentration of interflow
- C_{01} - chloride concentration of runoff (sheet flow)
- C_1 and C_{11} - chloride concentration of water from weathered zone
- C_{12} and C_{13} - chloride concentration of water from preferential bedding fractures

Figure 2.39 depicts the recharge process in horizontal strata zones (mesa). In this model, the following assumptions are made:

- 1) The strata are almost horizontal. The weathered zone only occurs in a certain area.
- 2) The plants thrive along fractures or weathered zones.
- 3) Hollows occur on ground or rock surface. Evaporation is considerable if water is held in hollows.
- 4) There is a preferential fracture zone between strata layers and many preferential fractures in a rock mass.
- 5) Evaporation occurs near the surface and evapotranspiration occurs along the fracture through plants.
- 6) The chloride concentration of preferential flow stays stable; it is similar to that of rainfall.
- 7) The concentration of runoff is C_{01} with $C_{01} > C_i$.
- 8) Concentration of interflow is the max value of the output of the weathered zone.

- 9) The concentration of infiltration water in fractures with plants is C_1 with $C_1 > C_r$, but the recharge volume can be smaller than that through preferential fractures.
- 10) When the infiltration water seeps as interflow through the weathered zone or preferential fracture zone, the chloride concentrations denote as C_{11} , C_{12} and C_{13} . The concentrations are almost similar in most scenarios.
- 11) The concentration of groundwater is a mixture of rainfall, interflow and regional flow. They satisfy

$$C_g \times V_g = C_{g0} \times V_{g0} + C_1 \times V_1 + C_r \times V_r$$

C_{g0} is the chloride concentration of old storage water in aquifer. The subscript of volume (V) is corresponding to the components of concentration.

Typical physical model expressed in the conceptual model (Figure 2.39) can be found near the Maclear's Beacon on Table Mountain (Figure 2.40).



Figure 2.40 Preferential fractures (Karst) with plants near the Maclear's Beacon on Table Mountain

2.3.5.3 Recharge processes in upright strata

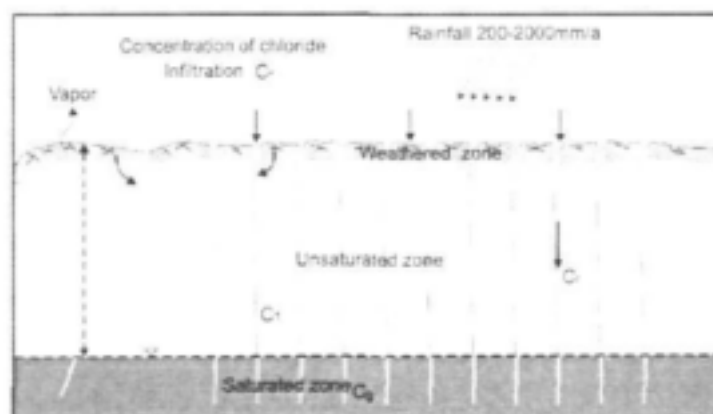


Figure 2.41 Recharge processes: case 3 with upright strata

C_r - chloride concentration of rainfall

C_g - chloride concentration of groundwater

C_1 - chloride concentration of water leakage from weathered zone

Figure 2.41 shows the model of the recharge process in almost upright strata as in karst areas. In this model, the assumptions are the following:

- 1) The strata are almost upright. The weathered zone only occurs in a certain area. There is a weathered zone with potential high hydraulic conductivity.
- 2) There is preferential fracture zone between strata layers and many vertical preferential fractures in the rock mass.
- 3) Evaporation occurs on the surface of a weathered zone. Salinity can accumulate in weathered zones by evaporation.
- 4) The chloride concentration of preferential flow remains stable; it is similar to rainfall. The chloride concentration of infiltration water in the weathered zone is C_1 with $C_1 > C_r$. The concentration of groundwater is a mixture of those of rainfall, interflow and old aquifer storage. They satisfy

$$C_g \times V_g = C_{g0} \times V_{g0} + C_1 \times V_1 + C_r \times V_r$$

C_{g0} is the chloride concentration of old storage water in aquifer. The subscript of volume (V) is corresponding to the components of concentration.

The typical topography expressed in the conceptual model (Figure 2.41) occurs in the core of anticline, such as Old English fort in Montagu and Uniondale Poort- Upright strata with open fractures shown in Figure 2.42 and Figure 2.43.



Figure 2.42 Upright strata with open fractures in Montagu



Figure 2.43 Upright strata with open fractures in the Uniondale Poort

2.3.5.4 Recharge processes in incline strata zones

In this scenario, three recharge processes related to dip and slope in cuesta or inclined topography areas are considered, these are as follows:

- the angle of dip is less than the angle of slope (case 4, Figure 2.44 A),
- the angle of dip exceeds the angle of slope (case 5, Figure 2.44 B) and
- the angle of dip is opposite to the slope (case 6, Figure 2.44 C)

From the above models, some common features can be drawn as follows:

- 1) The strata are monocline, and the dip angle could be less than or greater than the slope angle.
- 2) The weathered zone only occurs in a certain area. Stress relief fractures are located in relief. There is a preferential fracture zone between strata layers and many vertical preferential fractures in a rock mass.
- 3) Infiltration water may recharge the aquifer in the opposite catchment area through preferential fracture networks.
- 4) Part of the rainfall intercepted by plants infiltrates into soil along trunks of plants. Evaporation occurs on the surface. Evapotranspiration occurs along the fractures with plants.

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Figure 2.44 Recharge processes in the incline area

C_r - chloride concentration of rainfall, C_s - chloride concentration of surface water,
 C_g - chloride concentration of groundwater, C_e - chloride concentration of sheet flow,
 C_m - chloride concentration of interflow, C_{r1} - chloride concentration of runoff (sheet flow),
 C_1 - chloride concentration of water from weathered zone.
 A: case 4 with strata dip smaller than slope angle

B: case 5 with strata dip bigger than slope angle

C: case 6 with strata dip opposite to the slope

- 5) The chloride concentrations of the preferential flow are almost similar to that of rainfall. The chloride concentration of water in the weathered zone is C_1 with $C_1 > C_r$. The concentration of groundwater is a mixture of rainfall, interflow and old storage water. They satisfy:

$$C_g \times V_g = C_{g0} \times V_{g0} + C_1 \times V_1 + C_r \times V_r$$

C_{g0} is the chloride concentration of old storage water in aquifer. The subscript of volume (V) is corresponding to the components of concentration.

- 6) Interflow occurs in the weathered and preferential fracture zones above the elevation of local stream or river. The chloride concentrations of interflow are almost same in most scenarios. Interflow occurs more easily in case 4 than in the other two cases because bedding fractures are cut by valley.

The typical scenarios of these cases occur in a limb of a fold. Typical examples of case 4 and case 5 are shown in Figure 2.45 and Figure 2.46. The case 6 can be referred to Figure 2.14.



Figure 2.45 Dip angle less than the slope angle in the Bain's Kloof



Figure 2.46 Dip angle bigger than the slope angle in the Piketberg Pass

3. METHODOLOGY OF RECHARGE ESTIMATION

A comprehensive approach to understanding the TMG recharge was adopted. The research dealt with both temporal and spatial variability of groundwater recharge. Existing recharge estimation methods were reviewed in the light of their applicability to the TMG aquifers. The methodology of recharge estimation has incorporated the following main steps:

- Desk study
- Data collation
- Fieldwork
- Interpretation of hydrogeological and hydrogeochemical data
- Conceptualisation and classification of recharge processes at different scales
- Identification of appropriate estimation methods
- Calculation of recharge rates and comparison of results within a case study

3.1 Review of Historical Data

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

- A literature review on recharge mechanisms and estimations in fractured rock aquifer.
- A literature review of previous studies of the TMG aquifer includes geological and hydrogeological maps and reports, remote sensing images, recharge researches and other databases, which are Department of Water Affairs and Forestry (DWAF), South Africa National Groundwater Database (NGDB) and REGIS Database, Geographical Information Systems (GIS) of Cape Nature Reserve (CNC) and Water Research Commission (WRC) WR90 Flow data.

3.1.1 Data Collection and Analysis

Time series data were captured from National Groundwater Database (NGDB) and REGIS Database at the 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.

Rainfall samples were collected from three new monitoring stations at the Purification Works in Dysselsdorp for determination of chloride concentrations between 2003 and 2004. The samples of borehole water, surface water, interflow, spring flow and soil were collected from the field study. The chemical constituents of the samples were analysed at the laboratory. The statistical

analyses methods used in the study included the arithmetic, geometrical and harmonic mean, standard deviation and coefficient of variation.

3.2 Field Work

- The detailed fieldwork done during this study is categorised as follows:
- Geomorphological investigations using satellite images
- Hydrogeological surveys
- Borehole water, surface water, spring sampling in the Vermaaks River Well field
- Soil sampling
- Setting up monitoring rainfall stations
- Rain water collection

3.3 Construction of Conceptual Recharge Models

At different scales, the structural geology and topography play a significant role in defining the recharge processes and mechanisms. A framework for assessing recharge processes in the study area is illustrated in Figure 3.1.

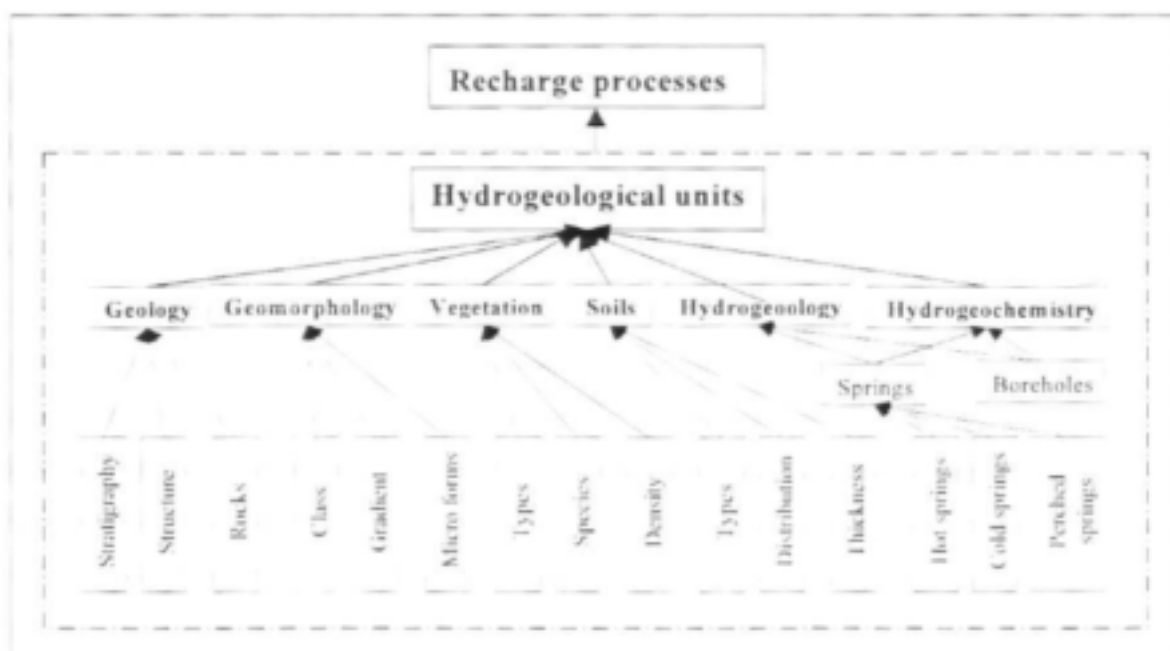


Figure 3.1 A framework for study of recharge processes

3.4 Recharge Estimation

An approach of cross-calibration of the following set of methods was adopted to quantify groundwater recharge rates:

- Tracer techniques: Chloride Mass Balance (CMB) method;
- Catchment Water Balance techniques, e.g. the Spring Volume Fluctuation (SVF) method; and
- The Cumulative Rainfall Departure (CRD) method by Bredenkamp *et al.* (1995) and Xu and Van Tonder (2001) and the Extended model called Rainfall Infiltration Breakthrough (RIB) by Xu and Beekman (2003);
- Verification of a numeric model simulation

A comparative study in the case study area enabled the testing of various methods and approaches to recharge estimation. A framework for the recharge estimation is illustrated in Figure 3.2. In this study, the CMB method is discussed in detail because the method has been widely applied to estimate groundwater recharge in semi-arid regions, especially in South Africa.

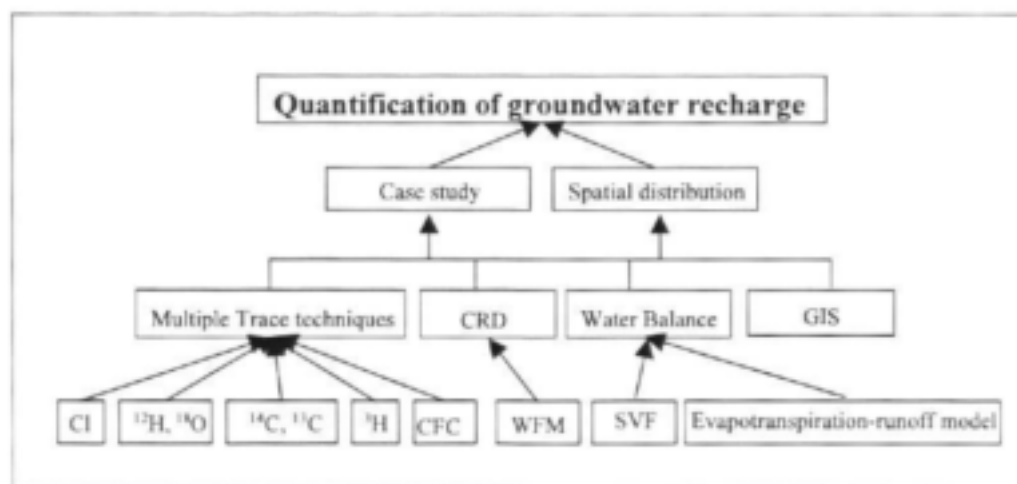


Figure 3.2 A framework for study of recharge estimation

A mixing model based on the CMB method was constructed for estimating recharge rate. Figure 3.3 shows the framework for the mixing model based on the study of hydrogeochemistry. The framework includes four steps, which are literature study, field study, interpretation of chemical data and constructing conceptual model and modelling.

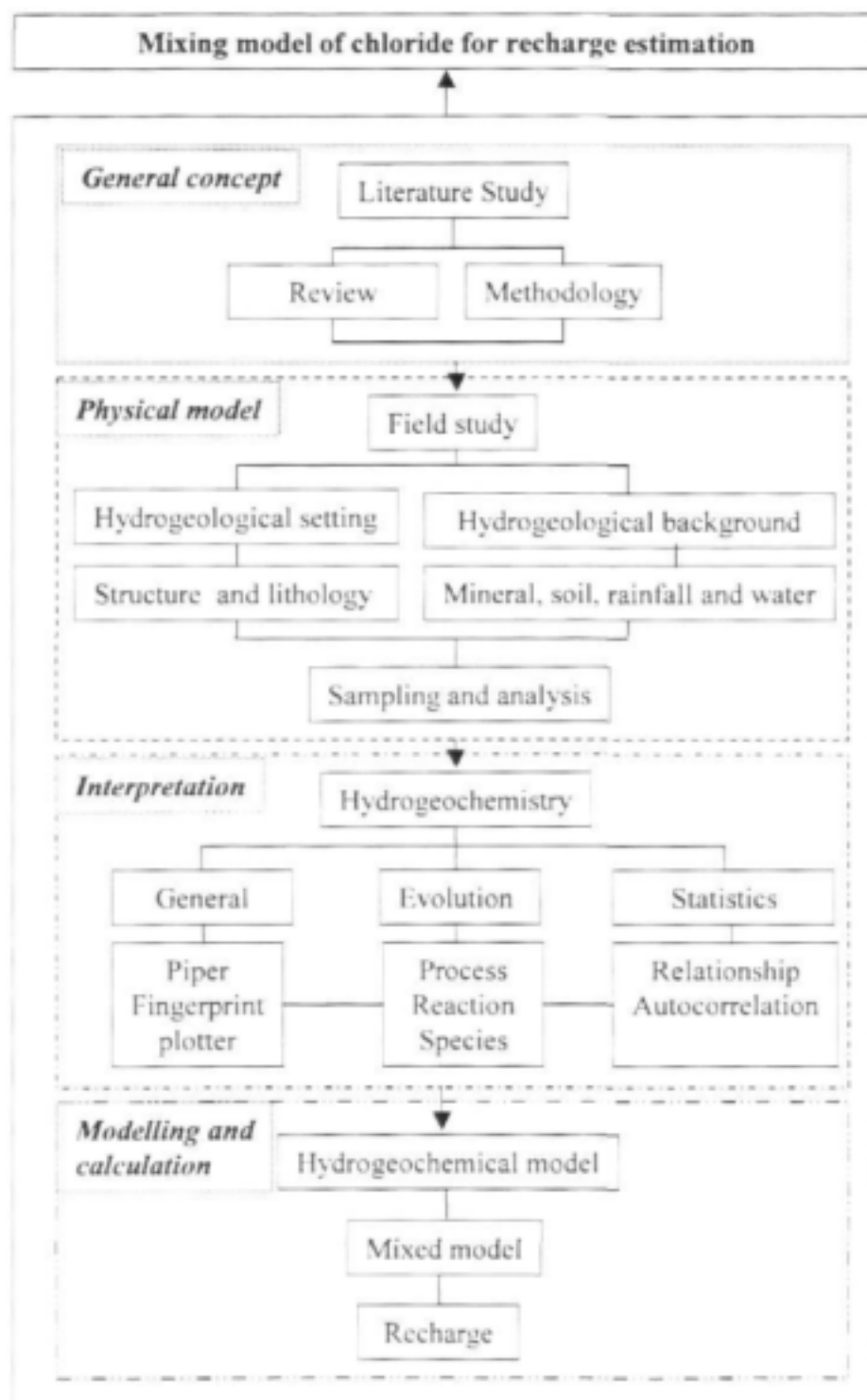


Figure 3.3 A framework for study of mixing model

3.5 Sensitivity of Recharge Estimates

A research methodology was developed to address the impact of climatic change in areas with various geomorphological characteristics, in the Kammanassie area. Geomorphological characteristics, relevant to groundwater recharge, of different areas of the TMG aquifer systems were classified within a GIS framework. Changes in climate and rainfall patterns and frequencies were related to aquifer recharge within the Kammanassie region. Influence factors of recharge such as temperature, soil, vegetation and rainfall intensity were examined and analysed. A framework for the sensitivity analysis of the recharge estimates is illustrated in Figure 3.4.

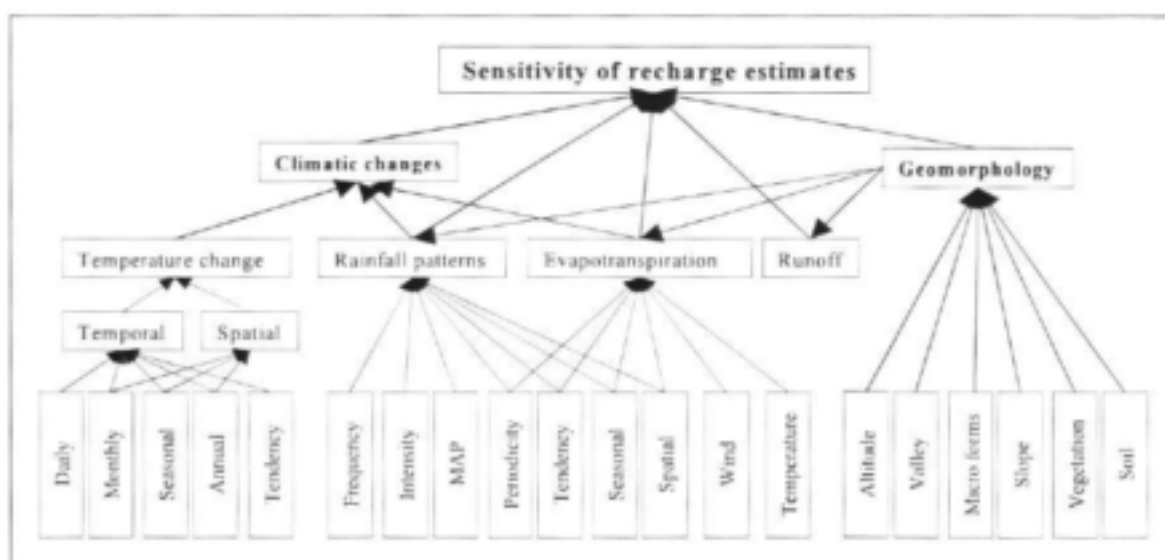


Figure 3.4 A framework of approach to sensitivity analysis of recharge estimates

3.6 Literature Review

Quantification of recharge is a pre-requisite for the assessment of groundwater resources. Yet it is impossible to measure recharge directly at a catchment and regional scale, especially in hard rock areas. Although there are various well established methods for the recharge estimation (Lerner *et al.*, 1990), few can be applied successfully to the TMG area. Assessment of groundwater recharge is one of the key challenges in determining the sustainable yield of the TMG aquifers, especially in the Kammanassie area where people are concerned about the negative impact on the National Park.

3.6.1 Overview of Recharge Estimation Methods

Most of the classification methods focus on the unsaturated and saturated zones and approach to the recharge, including direct measurement and indirect estimation, which are constructed by

Darcy's Law or water balance. Beekman and Xu (2003) classified the recharge estimation methods based on spatial scale and technique method as follows:

- Hydrogeological provinces: regions of similar climate and geology with similar geomorphic history. For example, alluvial fans and riverbeds, sand and sandstone and volcanic areas (Lerner *et al.*, 1990);
- Hydrological zones: atmosphere, surface water; and unsaturated and saturated zones (Bredenkamp *et al.*, 1995; Beekman *et al.*, 1999 and Scanlon *et al.*, 2002); or
- Physical and Tracer approaches: direct versus indirect, water balance and Darcyan physical methods; and chemical, isotopic and gaseous tracer methods (Lerner *et al.*, 1990; and Kinzelbach *et al.*, 2002).

An overview of commonly used recharge estimation methods is given in Table 3.1. The methods are grouped according to hydrological zones. A brief description of the principles, formulae, limitations and references is given for each method. Methods referring to surface water and the unsaturated zone estimate the potential recharge, whereas methods referring to the saturated zone estimate the actual recharge. Methods that have great potential to forecast recharge are those which have established relationships between rainfall, abstraction and water level fluctuations, such as the CRD, EARTH, Auto Regression Moving Averages and empirical methods. Critical in reliable forecasting of recharge is the accuracy of forecasting rainfall in terms of frequency of events, quantity and intensity. Note that the accuracy of forecasting recharge is further complicated by the non-linearity of groundwater resources in their response to rainfall (Beekman and Xu, 2003).

A review of commonly used methods in semi-arid Southern Africa is presented in Table 3.2. Methods were evaluated in terms of applicability and ratings (Beekman and Xu, 2003). The ratings themselves are mainly based on the authors' experience. They should be treated with caution since they are subjective by nature. Table 11 highlights the following:

All methods can be used in large ranges of flux. Most methods are only applicable at small scales. Some methods are used for short periods of time, which may not reflect the groundwater regime related to recharge.

The CMB method can be used in the unsaturated and saturated zones at scale with a range from 0.1m² to 10 000 m² from years to thousands of years (spatial scales range from ~200 m to several kilometres, Scanlon *et al.*, 2002).

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Table 3.1 Outline of estimation methods of recharge in literature

Zone	Method	Principle	Formula ¹⁾	Limitations	Ref.
Surface water	Hydrograph separation	Stream hydrograph separation outflow balances recharge minus evapotranspiration and abstraction	Groundwater recharge (Q_R) is function of input (I) and output (O), the general form is $Q_R = f(I, O)$	Ephemeral rivers, related to numerical model. Storativity is sensitive. Water table contour is not easy to obtain. O and I values are difficult to control	11, 31, 32
	Channel Water Budget (CWB)	Recharge derived from difference in flow upstream and downstream accounting for evapotranspiration, in and outflow and channel storage change	$RE_{eff} = R_t - RO_{eff} - ET_{tot}$	Flow measurements are inaccurate.	7
Unsaturated zone	Lysimeter	Difference in flux between rainfall and seepage proportional to recharge	$R = DS - FC$, $R = DPF \cdot (DS - FC)$, $R = CREC \cdot (S - FC) \cdot DS/SC$, $R = C \cdot \{b \cdot (DS - FC)\}$, $R = C \cdot (DS - FC) \cdot b$, $R = I \cdot \left\{1 - \frac{sc - DS}{sc - fc}\right\}$; HILACRES model: $R = P \cdot (DS - FC) \cdot C \cdot Q$, $R = CRAK \cdot (P - Q - E) \cdot DS/SC$, GSPB model, ARBM model, $R = BFI(1 - BFI) \cdot Q$	Many estimation models must be compared. It is difficult to control data quality and their spatial variability. This is ineffective for hard rock area.	13-29
	Unsat. Flow Mod. (UFM) / Richards' Equation	Unsaturated flow simulation e.g. by using numerical solutions to Richards equation	$\frac{\partial \theta}{\partial t} = \left[K(\theta) \left(\frac{\partial \psi}{\partial z} - 1 \right) \right]$	Poorly known relationship between hydraulic conductivity - moisture content	2
	Zero Flux Plane (ZFP)	Changes in soil moisture storage below Zero Flux Plane (-zero vertical hydraulic gradient) proportional to recharge	$F(z) = \int_0^z q(z) dz = \int_0^z [\theta(t1) - (\theta(t2))] dz$	Subsurface heterogeneity, periods of high infiltration, there is a problem when flux is low. Poorly known hydraulic conductivity of soil	2, 8, 9

**GROUNDWATER RECHARGE ESTIMATION
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Table 3.1 Continued

Unsaturated zone	Environmental (CMB)	Chloride Mass Balance - Profiling; drainage inversely proportional to Cl in pore water	$R_{mci} = \frac{TD \cdot TM}{TC}$	Long-term atmospheric deposition unknown; Chlorine recycling in biosphere and atmosphere is not known	1,2,8,9, 35
	Historical: Tritium profiles	Vertical distribution of tracer as a result of activities in the past (3H , 3Cl)	$RE(\%) = \frac{^3H_{ex}}{^3H_{st}} \cdot 100$	Present 3H levels are almost undetectable; 3H is lost in other ways; or 3H water has penetrated beyond the soil	1,2, 8,9
Saturated-unsaturated zone	Cumulative rainfall departure (CRD)	Water level response proportional to cumulative rainfall departure	$R_t = r(CRD)_t = S_y \{ \Delta h_t + [Q_p + Q_{out}] / (\Delta S_y) \}$ with $(CRD)_t = \sum_{i=1}^t P_i - \left(2 - \frac{1}{P_{eq}} \sum_{i=1}^t P_i \right) P_t$	Deep (multi-layer) aquifer, sensitive to S_y	2, 3, 4
	Water Balance (WB)	Based hydrologic cycle water balance and climate elements	$Re(t) = P(t) - Et(t) - Q(t) - \Delta\theta(t)$	Et and $Q(t)$ are theoretical; $\Delta\theta(t)$ should be equal to zero	30
	EARTH	Lumped distributed model simulating water level fluctuations by coupling climatic, soil moisture and groundwater level data for estimating recharge	$RE = f(\text{Maxil, SOMOS, LINRES, Satflow})$ $RE = S(h_t - c_s h_{t-1})$ $P_s + R_s + RE + Q_d = P$	11 parameters are required; Poorly known S_y	5, 8
	Water Table Fluctuation (WTF)	Water level response proportional to recharge/discharge	$RE = a \cdot Rf + b$	In/outflow and S_y usually unknown	2

**GROUNDWATER RECHARGE ESTIMATION
OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES**

Table 3.1 Continued

Saturated-unsaturated zone	Environmental (CMB)	Chloride mass balance: amount of Cl into the system balanced by amount of Cl out of the system for negligible surface runoff/runon	$R_i = \frac{T_d}{Cl_{ss}}$ $\frac{dF_x}{dx} = Q + D \cdot \nabla \cdot F_x = \frac{dF_x}{dx} = R$ $\nabla \cdot F_x = \frac{\partial F_x}{\partial x} + \frac{\partial F_y}{\partial y} = P - E - S = R \cdot F_x = \int_0^h \nabla c_x dz$	Long-term atmospheric deposition unknown. Chlorine recycling in biosphere and atmosphere is not known.	1, 2, 4, 6, 8, 9, 32
	Groundwater Modelling (GM)	Recharge inversely derived from numerical modelling groundwater flow and calibrating on hydraulic heads / groundwater ages	$RE_{average} = T \cdot \frac{W}{A} \cdot \frac{dh}{dt} \times 365 \times 1000 \text{ mm / a}$	Time consuming, poorly known transmissivity, sensitive to boundaries. T and S values are usually derived by assuming steady state conditions	2, 8
Saturated zone	Inverse of the dynamic model (Model flow)		$S_{x,y} \frac{\partial h}{\partial t} = \frac{\partial}{\partial x} \left(T_x \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(T_y \frac{\partial h}{\partial y} \right) - RE_{(x,y,t)} + Q_{(x,y,t)}$		
	Sat. Volume Fluctuation (SVF)	Water balance over time based on averaged groundwater levels from monitoring boreholes	$S \cdot \frac{\Delta V}{\Delta t} = I - O + RE - Q_e$	S is not easy to decide. I value is often assumed equal to O value. Flow-through region, multi-layer aquifers	2
	Water balance: EV-SF	Water balance at catchment scale	$RE = \frac{Q_s}{A}$	Shallow groundwater, regional flow cannot be considered	2
	Groundwater dating (GD)	Age gradient derived from tracers, inversely proportional to recharge; Recharge unconfined aquifer based on vertical age gradient (³ H, CFCs, ³ H/ ³ He)	$t_s = \frac{nH}{RE_s} \ln \left(\frac{H}{h} \right), t_a = \frac{1}{\lambda} \ln \left(1 + \frac{n_0 D}{RE_s \lambda} \right)$	¹⁴ C, ³ H/ ³ He, CFC: poorly known porosity / correction for dead carbon contribution. Poorly known porosity. Recharge confined aquifer based on horizontal age gradient (¹⁴ C and ³⁶ Cl)	1, 9, 33, 34, 36, 37

**GROUNDWATER RECHARGE ESTIMATION
OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES**

Table 3.1 Continued

Saturated zone	Stable isotopes	$^{18}\text{O}, ^2\text{H}$	$Q_e = Q_s \left(\frac{\delta^{18}\text{O}_s - \delta^{18}\text{O}_e}{\delta^{18}\text{O}_e - \delta^{18}\text{O}_g} \right)$	There is no monogamous relationship between rainfall, rivers and groundwater. Flow paths are not easily known. Various mixing ratios could yield the same final isotopic value, lack of a distinct signature.	12
	Groundwater temperature		$T(z,t) = T_0 + T_{10}(z - Ut) + \frac{1}{2}(b + T_{10}U)/2U \{ [(z + Ut)e^{z^2/4at}] \times \text{erfc}\{(z + Ut)/2(at)^{1/2}\} + (Ut - z)\text{erfc}\{z - Ut/2(at)^{1/2}\}\}$ $U = vc_0\rho_0/c\rho$	The thermal diffusivity and the heat capacity of the aquifer are not easy to obtain.	

1 Beekman et al., 1996	7 Lerner et al., 1990	13 Chapman and Malone, 1999	19 Boughton, 1993b	25 Chiew et al., 1993	31 White et al., 2003
2 Bredenkamp et al., 1995	8 Gieske, 1992	14 Aston and Dunin, 1977	20 Fleming, 1975	26 Lopes et al., 1982	32 Cook et al., 2001
3 Xu & van Tonder, 2001	9 Sclaolo, 1998	15 AWRC, 1969	21 Chapman and Malone, 1999	27 Dawdy and O'Donnell, 1965	33 Cook and Solomon, 1997
4 Weaver & Talma, 2002	10 Sami & Hughes, 1996	16 Black et al., 1969	22 Huggins and Monke, 1967	28 Ye et al., 1997	34 Cook et al., 1998
5 Lee and Gehrels, 1997	11 Xu et al., 2002	17 Boughton, 1993 a, b	23 Chapman, 1970	29 Evans and Jakeman, 1998	35 Cook et al., 1994
6 Beekman and Sanguro, 2002	12 Fritz et al. 1976	18 Boughton, 1993b	24 Jakeman and Hornberger, 1993	30 Tyner et al., 2000	36 Cook and Solomon, 1995
					37 Weaver et al., 1999

1) Symbols are designated in general notation

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

Table 3.2 Review of common recharge estimation methods used for semi-arid Southern Africa
(Beekman and Xu, 2003)

Zone	Method	Applicability ²			Rating ³		
		Flux (mm-yr ⁻¹)	Area (km ²)	Time (yrs)	Acc.	Ease	Cost
SW	Baseflow	400-4000 (0.1-1000)	10 ⁴ -1300 (10-1000)	0.3-50 (1-100)	2	2	1-2
	CWB	100-5000	10 ³ -10	1d-1yr	2-3	2	3
	WM	1-400	10 ¹ -5×10 ⁵	1d-10yr	2	2-3	3
Unsaturated ¹	Lysimeter	1-500	0.1-30m ²	0.1-6	1-2	3	3
	UFM	20-500	0.1-1m ²	0.1-400	3	2	2
	ZFP	30-500	0.1-1m ²	0.1-6	3	2	2
	CMB	0.1-300 (0.6-300)	0.1-1m ²	5-10000	2	1	1
	Historical	10-50 (10-80)	0.1-1m ²	1.5-50	2-3	2-3	3
Saturated - Unsaturated	CRD	(0.1-1000)	(1-1000)	(0.1-20)	1-2	1-2	1
	EARTH	(1-80)	(1-10m ²)	(1-5)	1-2	2	1
	WTF	5-500	5×10 ⁻⁵ ->10 ⁻³	0.1-5	2	1	1
	CMB	0.1-500	2×10 ⁻⁶ - >10 ⁻²	5->10000	2	1	1
Saturated	GM	(0.1-1000)	(10 ⁻⁶ -10 ⁻⁵)	(1d-20yr)	2	3	2-3
	SVF	(0.1-1000)	(1-1000)	(0.1-20)	1-2	1-2	1
	EV-SF	(0.1-1000)	(1-100)	(1-100)	1-2	1-2	1-2
	GD	¹⁴ C: 1-100 ³ H/ ³ He, CFC: 30-1000	¹⁴ C, ³ H/ ³ He ³ , CFC: 2*10 ⁻⁶ ->10 ⁻³	¹⁴ C: 200-200000 ³ H/ ³ He ³ , CFC: 2-40	3	2-3	3

HS: Hydrograph Separation – Baseflow	EARTH: Extended model for Aquifer Recharge and Moisture Transport through Unsaturated Hardrock
CWB: Channel Water Budget	
WM: Watershed Modelling	WTF: Water Table Fluctuation
UFM: Unsaturated Flow Modelling	GM: Groundwater modelling
ZFP: Zero Flux Plane	SVF: Saturated Volume Fluctuation
CMB: Chloride Mass Balance	EV-SF: Equal Volume - Spring Flow
CRD: Cumulative Rainfall Departure	GD: Groundwater Dating

¹All methods for estimating fluxes through the unsaturated zone assume diffuse vertical flow whereas in reality flow along preferred pathways is the rule rather than the exception. These methods therefore tend to overestimate the diffuse flux.

²Data in brackets are estimates from Southern Africa; Rainfall may be up to 2000 mm-yr⁻¹.

³Ratings for methods applied to semi-arid southern Africa.

Rating: Accuracy:1-Difference from true value within a factor of 2, 2-within a factor of 5 and 3-within a factor of 10 or more. Ease of Application is related to data requirements and data availability and is rated from 1-easy to use to 3-difficult to use. Cost is rated from 1-inexpensive to 3-expensive.

3.6.2 Recharge forecast

Forecasting groundwater recharge has become increasingly important, particularly with regard to the envisaged impacts of climatic change in short and long term on the limited water resources in the TMG area. Some guidelines for recharge estimation are given in Lerner *et al.* (1990) and Scanlon *et al.* (2002). The combined chemical and isotopic mass balance approach is based on dating moisture and groundwater using the CMB and ^{14}C groundwater dating methods.

Some methods can be applied with greater certainty in the arid and semi-arid regions of Southern Africa, such as CMB, CRD, EARTH, WTF, GM and SVF (Xu and Beekman, 2003). Three of these methods: WB, CRD and CMB are widely applied and will be discussed in more detail. They represent an increasing complexity in their use and data requirements. It is worth mentioning that the approach based on integrating spatial climatic and hydrogeological datasets in a GIS environment, namely DRASTIC approach (Aller *et al.*, 1987), may have the potential to become quantitative once it is combined with recharge estimation methods given in Table 3.1.

3.6.3 Recharge Study in the TMG Area

The first systematic recharge studies carried out in South Africa date back to the early 1970's and were undertaken in the Western Transvaal (Bredenkamp and Vogel, 1970 and Bredenkamp *et al.*, 1974) and the Northern Cape (Smit, 1978). Recharge studies were mostly carried out on a local scale and as part of a larger groundwater resource assessment project. A manual, namely "Preparation of a Manual on Quantitative Estimation of Groundwater Recharge and Aquifer Storativity" (Bredenkamp *et al.*, 1995), presents a great variety of well-tested (semi-empirical) methods that are widely employed in South Africa. In addition, it contains a wealth of recharge case studies and data covering the last 30 years. Of particular interest is that a first attempt was made to rate the different estimation techniques in terms of ease of application, reliability and availability of data.

In the last 10 years, some recharge studies were conducted in the TMG area. The results are listed in Table 3.3. The following become clear:

- CMB is the most independent method to estimate recharge in the TMG area. There are large differences in the recharge estimates using different methods by different researchers in the same area. A typical example is the Vermaak's River Catchment area.
- High recharge rates were obtained by using GIS but no detailed algorithm documented.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

Table 3.3 Review of the recharge estimates in the TMG area

Place		Method	Aquifer	MAP	RE %	Source
Vermaak's River		Base flow	Peninsula	560	8.9	Bredenkamp, 1995
		Unknown	Peninsula		17	
		CMB	Nardouw Peninsula /	299-714	43.5 /11.1	Kotze, 2000 and 2002
		SVF			3.1 /19.7	
		SVF (fit)			3.4 /16.4	
		CRD			3.2 /14.4	
		Base flow			21.4 /12.5	
		EARTH			2.9 /23.9	
		² H displacement			0.3 /2.7	
		¹⁴ C age			1.8 /2.0	
		GIS raster	TMG	2.5-4.8	Woodford, 2002	
		CMB		5	Weaver et al., 2002	
		Langebaan Road Field		CMB	Bredasdorp	396.4-648.9
CMB	11.2					
CMB	13.5					
CMB	11.5					
Greater Oudtshoorn Districts		Empirical	Peninsula	165-1049	14	Hartnady and Hay, 2002
		Empirical	Nardouw		7	
Struibaai		CMB	TMG	436	17.4	Weaver and Talma, 1999, 2002
Agter Witzenberg		CMB	Nardouw	579-777	50.44	
Botriver			Nardouw	477.2-1546.2	20	
Hermanus		¹⁴ C		635	22 (20-24)	Rosewarne and Kotze, 1996
Uitenhage Artesian Basin		Spring flow		298.2-1202.6	11	Xu and Maclear, 2003
		CMB	Groot Winterhoek		25	Maclear, 1996
		CMB			55	Kok, 1992; Parson, 2002
Whole TMG (within a radius of 200km from Cape Town)			13200km ² outcrop	600-2.020	33	Weaver and Talma, 2002
CAGE	Low rainfall	GIS	Low lying	285.8-371.2	8	Hartnady and Hay, 2000 and 2002
	High mountain				23	
	Mountain area		TMG			
				50	Weaver et al., 1999	
Uitenhage, Coega aquifers		WB	TMG	460	83	Kok, 1992
		CMB	TMG	250-850	24-55	Maclear, 1996
		WB			11	Bredenkamp, 2000; Xu and Maclear, 2003

3.6.4 Chloride Mass Balance Method

The CMB method has been extensively applied to estimate groundwater recharge in arid regions (Eriksson and Khunakasem, 1969; Allison *et al.*, 1985; Johnston, 1987; Phillips, 1994; Wood and Sanford, 1995; Sami and Hughes, 1996; Tyner *et al.*, 2002 and Cook *et al.*, 1994 and 2001). It equates the mass flux of chloride at the ground surface to that of chloride beneath the depth at which evapotranspiration takes place. At the soil surface chloride occurs naturally from precipitation and dry fallout. Slightly higher recharge rates can be estimated using Cl in groundwater than in soil water because extraction of water from the soil generally requires additional dilution (Scanlon *et al.*, 2002). The use of methods that involve chemical tracers provide more accurate estimates of groundwater recharge in arid to semi-arid regions (Bazuhaire and Wood, 1996). The method is applicable in areas of high evapotranspiration where the infiltrating water becomes concentrated (Johansson, 1987).

The CMB is often used as a first approximation of recharge due to its simplicity and cost effectiveness. Preferably, results should be verified with other methods. There are three approaches to estimate recharge using the CMB method:

- The first approach involves the chloride concentration of rain and soil moisture (unsaturated zone) and gives an estimate of the moisture flux.
- The second approach involves the chloride concentrations of groundwater (saturated zone), whereby the total average recharge is estimated.
- The third approach involves the comparison of the first two approaches to define the recharge mechanism.

3.6.5 Groundwater Level Fluctuation Methods

Hydrographs are charts that display the change of a hydrologic variable over time. Although they are usually generated for streams, hydrographs can also be made for lakes, water wells, springs and other water bodies. A hydrograph provides a way of seeing seasonal and yearly changes in the water level of an aquifer related to rainfall. The Water-table Fluctuation Method (WFM) may be the most widely used technique for estimating recharge. The advantages of the WFM include its simplicity and insensitivity to the mechanism by which water moves through the unsaturated zone. Uncertainty in estimates may be generated by the limitation of specific yield (Healy and Cook, 2002).

The Cumulative Rainfall Departure (CRD) method is based on the premise that water level fluctuations are caused by rainfall events (Bredenkamp *et al.*, 1995). The method is applied extensively in South Africa. The method was revised to accommodate for trends in rainfall time series (Xu and Van Tonder, 2001; Xu and Beekman, 2003). It could be used to obtain the variability of recharge even without a recharge coefficient or storativity. The method cannot be applied in areas without groundwater level fluctuations or water levels not responding well to rainfall. In the above form it should only be applied to unconfined aquifers. Groundwater levels of fractured aquifers with small storativity are particularly sensitive to estimates resulting from the method.

A method, namely Rainfall Infiltration Breakthrough (RIB or recharge) was used to estimate recharge, in which only part of infiltrated rainfall breaks through the zone where evapotranspiration occurs and percolates to the groundwater table (Xu and Beekman, 2003). The arrival time of this water at the water table is delayed due to the 3-dimensional spreading of moisture. The records of rainfall events are expressed as a time series: $P_1, P_2, P_3, \dots, P_m, P_{m+1}, P_{m+2}, \dots, P_n, P_{n+1}, P_{n+2}, \dots, P_i$, scenario *a* (a combination of point and diffuse recharge): recharge is the cumulative result of all previous rainfall. Scenario *b* (quick infiltration process): recharge results from the previous *n* rainfall events immediately before the current rainfall event. Scenario *c* (a piston flow in unconsolidated aquifers): recharge results from a limited rainfall series between *m* and *n*.

3.6.6 Water Balance Approach to Recharge

The hydrological water balance for an area can be defined as:

$$R_e(t) = P(t) - E_t(t) - Q(t) - \Delta\theta(t)$$

where $R_e(t)$ is recharge, $P(t)$ is precipitation, $E_t(t)$ is evapotranspiration, $Q(t)$ is runoff, and $\Delta\theta(t)$ is change in volumetric water content. The term (t) designates that the terms are distributed through time. A common approach for approximating $E_t(t)$ involves first, the estimation of a reference crop, evapotranspiration as a crop coefficient, soil and rocks types. A common method to estimate $Q(t)$ is to apply the SCS runoff equation (USDA-SCS, 1985). $E_t(t)$ and $Q(t)$ were initially estimated based on meteorological and site-specific. Assuming $\Delta\theta(t)$ is negligible between subsequent years, an annual estimate of $R_e(t)$ can be calculated. The method can be applied in most recharge scenarios. The accuracy is related to the recharge process and impact of soils and vegetation. Monthly or daily rainfall and temperature data, together with coefficients of soil and vegetation are required.

4. CASE STUDY OF THE KAMMANASSIE AREA

Although a couple of case studies are carried out during the project (Sun, 2005; Jia, 2006), the case study in Kammanassie was thoroughly executed and hence reported here.

4.1 *Spatial and Temporal Rainfall Variability and Trends*

The Kammanassie area is located in the eastern section of the Klein Karoo, which forms part of the semi-arid region. The annual average precipitation of semi-arid regions is between 200 and 500 mm (Lloyd, 1986). There has not been a distinctive bimodal seasonal cycle and a general positive rainfall trend since 1925, but a general negative trend in short period appears from 1971 to 2001 (Cleaver *et al.*, 2003). The low precipitation and large spatial and temporal variation lead to more complexity of recharge estimation in the study area.

4.1.1 Precipitation

The 15 standard homogeneous rainfall districts over the TMG area defined by the South African Weather Service (SAWS) are shown in Figure 4.1. The Kammanassie area belongs to the temperate-maritime type and is located in districts 10 and 11. This topography of the Kammanassie Mountain limits the spatial distribution of the rainfall resulting in a more complicated spatial distribution of recharge (Figure 4.2).

The average monthly rainfall data captured from the WR90 for 111 rainfall stations over the Kammanassie and surrounding region from 1920 to 1989 has been used in a spatial distribution analysis. The dots marked in the Figure 4.2, namely Parshall, V-notch, Wildebeesvlakte and Purification Works East rainfall stations, were used to analyse rainfall pattern in the Kammanassie mountainous area for in the period 1994-2003. Approximately 400 to 600 mm of the mean annual precipitation (MAP) is observed from the contour map made with the Kriging method in the Kammanassie area (Figure 4.3). The rainfall in the northern area of the Kammanassie Mountain is lower than that in the southern area.

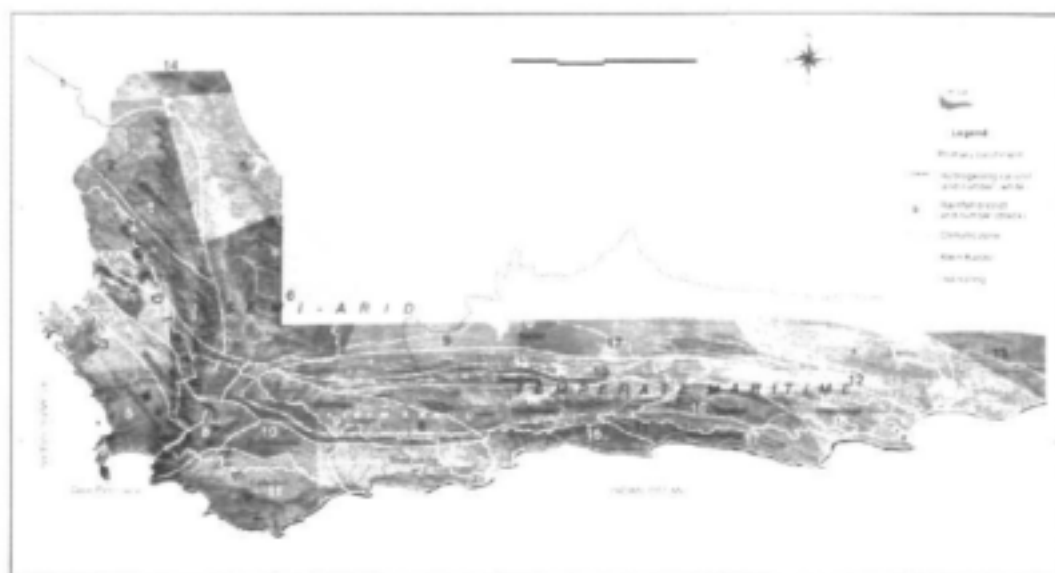
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Figure 4.1 Homogeneous rainfall districts over the TMG area

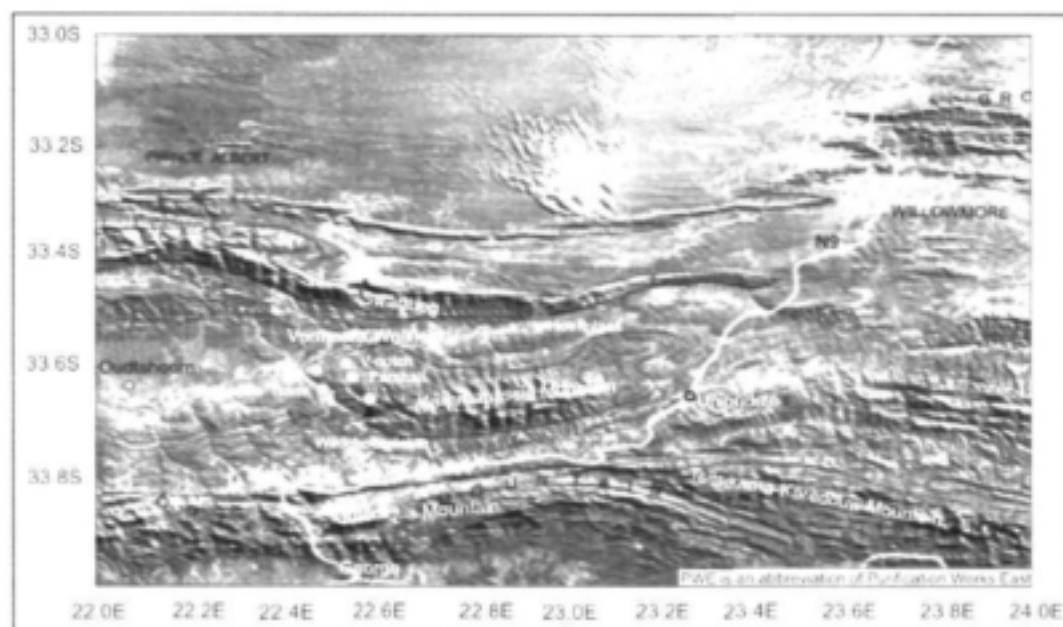


Figure 4.2 Selected area for rainfall analysis

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

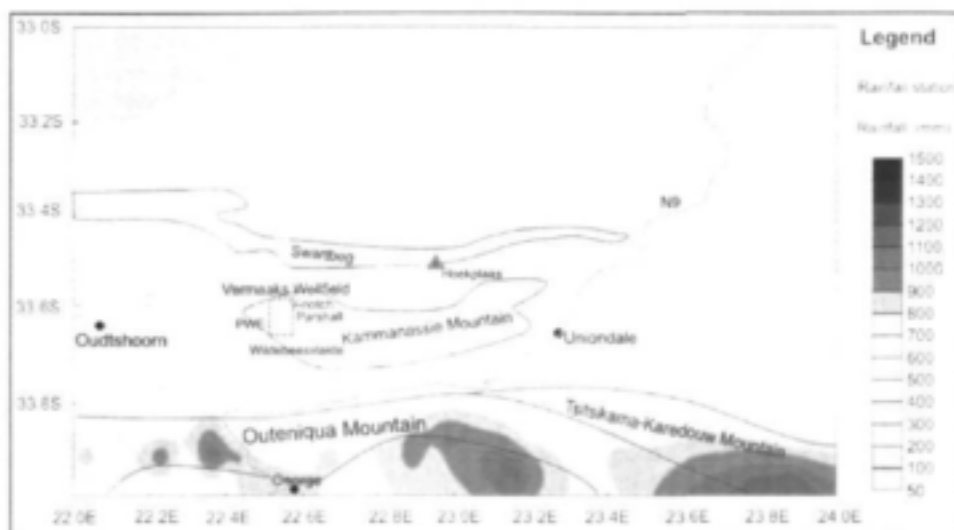
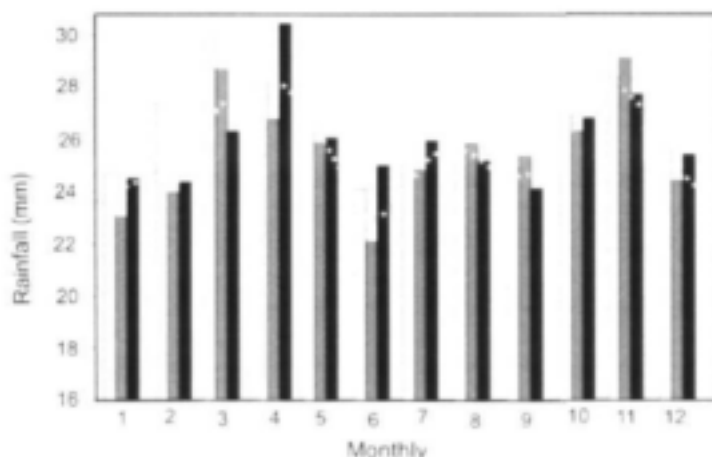


Figure 4.3 Contour of the mean annual precipitation in the Kammanassie area

In order to discuss the impact of seasonal rainfall on recharge, the monthly average rainfall over a period of 76 years (1921-1996) is shown in Figure 4.4 (Cleaver *et al.*, 2003), in which the light grey is 1921-1996 for district 10, the grey is 1971-2000 for rainfall stations 1, 3, 4, 10, 24 and 31 (see Figure 4.3); the black is 1926-1966 for Oudtshoorn. The relative effective recharge may mainly occur in March, April and November because of its higher average monthly rainfall. This may result in limited recharge, but rainfall including high rainfall events in the other months can contribute to the recharge as well.



Fresh grey -district 10; Grey-stations 1, 3, 4, 10, 24 in Figure 8; Black-Oudtshoorn

Figure 4.4 Monthly average rainfall calculated for district 10 from 1925 to 1996

(After Cleaver *et al.*, 2003)

4.1.2 Rainfall Variation

Rainfall variation with time scales is another important factor for recharge. The rainfall records (1925-2000) of Hoekplaas rainfall station for the Klein Karoo, the records (1925-2000) in the Kammanassie Dam for the southern Kammanassie Mountain area, and the records (1994-2003) in Wildebeesvlakte, Parshall, V-notch and Purification Works East rainfall stations especially for the Kammanassie Mountain area, are selected to address the long-term and short-term scenarios (see Figure 4.2).

4.1.2.1 Long-term tendency

Hoekplaas rainfall station is located at 33° 33' 03"S and 22° 58' 30" E (a private site owned by Mr. I Barnard, which is indicated by a triangle in Figure 4.2). The salient features of rainfall in this site are depicted in Figure 4.5.

For the Kammanassie Dam rainfall station, a gradual positive trend with a slope of +0.6156 in the monthly rainfall series was recorded from 1925 (Figure 4.6). The coefficient of correlation and the one-tailed probability from the F-Test between the period (1925-2002) and MAP are 0.14 and 5.41×10^{-29} , respectively. Thus, there is no significant correlation. The high rainfall values measured in 1980/1981, 1992/1993 and 1996/1997 indicate an increase in the magnitude of rainfall totals implying an increase in the magnitude of extreme rainfall events. The CRD seems to have the period of 43 years from 1948/1949 to 1990/1991. The new increasing period began in 1990/1991 although an extremely low rainfall event occurred in 1997/1998.

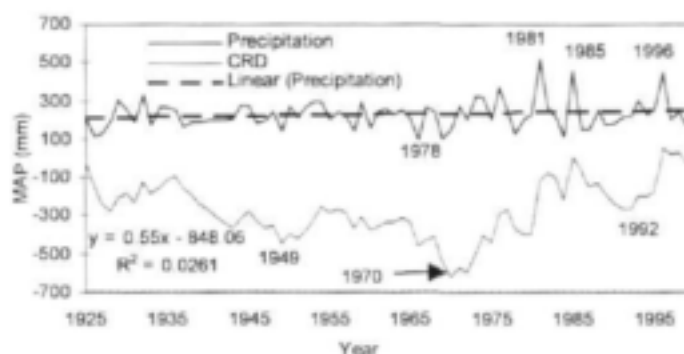


Figure 4.5 Rainfall and CRD patterns of the period 1925-2000 in the Hoekplaas

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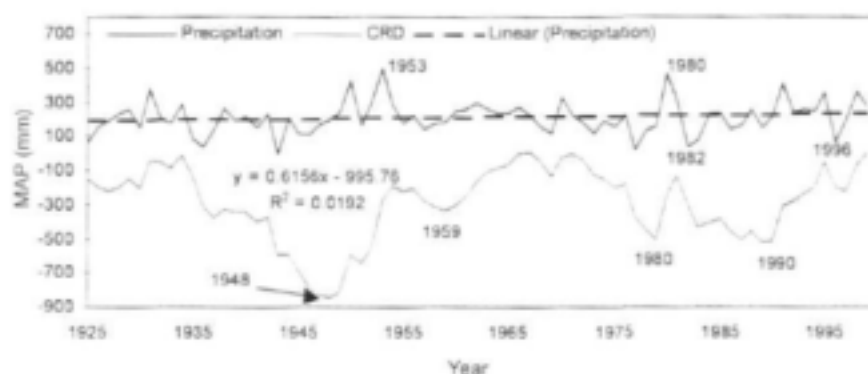


Figure 4.6 Rainfall trend in the Kammanassie Dam rainfall station (1925-2001)

4.1.2.2 Seasonal variation

Recharge may be significantly influenced by seasonal variation of rainfall. Four seasons have been distinguished (Cleaver *et al.*, 2003) in the Klein Karoo, namely December-January-February (DJF), March-April-May (MAM), June-July-August (JJA) and September-October-November (SON), as listed in Table 14. The lowest precipitation with mean values from 11.9 to 12.9 mm·a⁻¹ occurs in the DJF season. The highest one is from 19.1 to 21.5 mm per month in MAM season.

Table 4.1 Seasonal rainfall patterns at the Kammanassie Dam rainfall station (1925-2001)

Statistical analysis	SON			DJF			MAM			JJA		
	SEP	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG
Size	75	75	74	74	74	74	74	74	74	75	75	75
Min (mm)	0	0	0	0	0	0	0	0	0	0	0	0
Max (mm)	104	77	137	82	114	94	83	140	114	86	74	118
Mean (mm)	17.8	19.2	20.3	12.2	11.9	12.9	21.5	21.1	19.1	16.4	19.1	20.3
Stand. deviation	20.4	19.1	27.1	17.0	20.2	17.9	19.8	26.0	21.0	17.0	17.6	23.6
Coef. of variation	114.9	99.4	133.3	139.0	170.6	138.7	92.3	122.9	110.0	103.6	91.9	116.4

The rainfall values in any singular month have a long-term variation. This could lead to more complex recharge processes. For any month, the increased monthly rainfall in a time series would potentially increase recharge in the period. The period with increasing or decreasing rainfall is identified through the pattern of the CRD. Increasing CRD indicates relatively more rainfall than monthly average rainfall. The decreasing CRD pattern implies less rainfall than monthly average rainfall. The records in the Kammanassie Dam rainfall station were used to perform the CRD analysis. The monthly CRD (1925-2000) in the different seasons is shown in Figure 4.7 to Figure 4.10. The following observations were made:

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- In December, the decrease in the CRD was sustained from 1933 to 1992.
- In January, the first continuous decreasing CRD period was in 1933 to 1958; the second period started in 1964 and ended in 1998 after a short period of increase.
- In February, the lowest CRD was in 1938, 1951 and 1998; the longest sustaining period of unusually scarce rainfall was for 1959 to 1997.
- The period of more rainfall started in 1992 with increasing CRD in the DJF season.
- A similar tendency occurred in March and May. The period of increase for CRD was from 1944 to 1979 and from 1992 to 2001. The period of decrease in the CRD was for 1979 to 1991. Relatively more rainfall occurred from 1944 to 1979 and 1992 to 2001 but unusually scarce rainfall occurred from 1979 to 1991 in March and May.
- The longest period of decrease in the CRD occurred from 1925 to 1980 in April. The CRD increased a little and reduced again from 1964 to 1980. The period of increasing CRD occurred after April 1980. An anomalous scarce rainfall period lasted for 65 years in April. More rainfall has been occurring since 1980 in the MAM season.
- A period of increase in the CRD started in the 1990's in the MAM season.
- The period of increase in the CRD was from 1951 to 1992 in June, and from 1949 to 1972 in July. The period of decrease in the CRD occurred after 1972 in July. Relatively more rainfall was recorded in 1951 to 1992 in June and unusually scarce rainfall occurred after 1972 in July. A period of decreasing CRD started since 1992 in the JJA season.
- For September, the decreasing period of the decreasing CRD patterns had been occurring since 1943. For November, the period of decreasing CRD pattern was from 1966 to 1994. An unusually scarce rainfall occurred in these years, and relatively more rainfall occurred in November after 1994, but up to now the scarce rainfall period occurs in September.
- The points of the lowest CRD patterns occurred in 1942 and 1988 in October. In the 1925 to 2001 period, two scarce rainfall periods and two high rainfall seasons occurred in October.

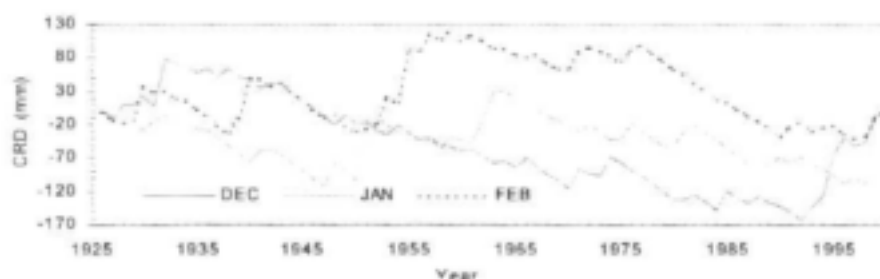


Figure 4.7 CRD patterns in the DJF season

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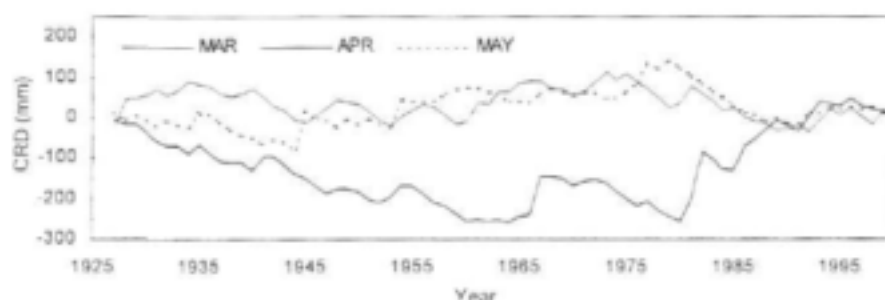


Figure 4.8 CRD patterns in the MAM season

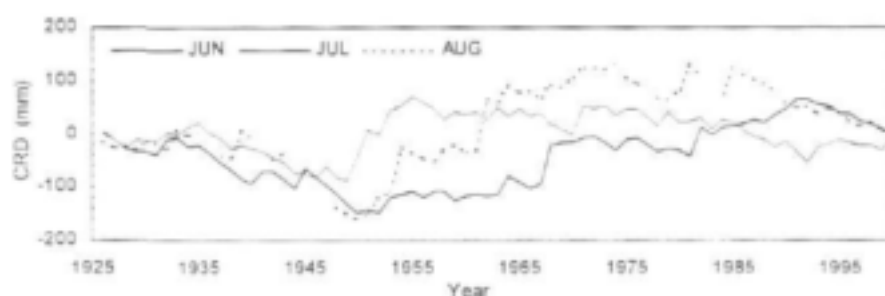


Figure 4.9 CRD patterns in the JJA season

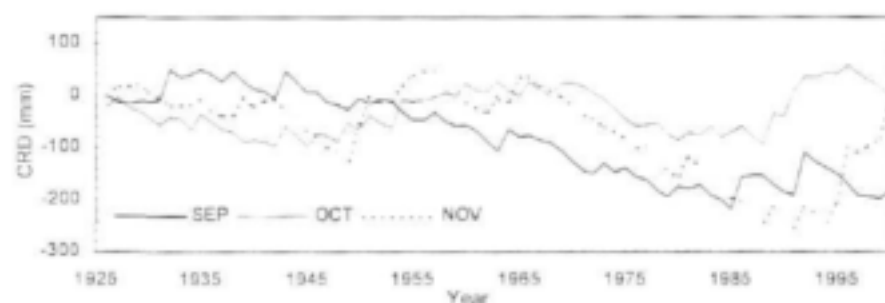


Figure 4.10 CRD patterns in the SON season

4.1.2.3 Extreme rainfall event

The extreme rainfall events play a major role on recharge in the study area. According to Table 4.2), the average value for the Hoekplaas rainfall station is 232.1 mm a^{-1} ; the mean minimum of the rainfall is less than the average value, which is 181.3 mm a^{-1} ; the mean maximum of the rainfall is higher than the total average values, which is 295.5 mm a^{-1} . The anomalous events with 16 extremely low rainfall events and 12 extremely high rainfall events are identified (Table 4.4 and Table 4.5) using the critical values (Table 4.3).

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Table 4.2 Statistical results of precipitation in the Kammanassie area (mm a^{-1})

Rainfall station	Hoekplaas (I. Barnard)			Kammanassie Dam		
Item	Total	Below total average	Above total average	Total	Below total average	Above total average
Years	75/76	42	34	75	38	37
Minimum	98.5	98.5	240.8	0.0	0.0	214.0
Maximum	517.3	231.3	517.3	498.0	208.0	498.0
Mean	232.1	181.3	295.5	209.3	138.8	283.9
Geomean	225.5	176.4	300.3	191.0	115.4	275.6
Standard dev.	75.3	35.2	65.8	97.1	54.2	72.6

Table 4.3 Critical value of anomaly

Site	Extremely high rainfall events (mm a^{-1})		Extremely low rainfall events (mm a^{-1})	
	Average + Standard deviation	Average above total average	Average - Standard deviation	Average below average
Hoekplaas	307.3	295.5	156.8	181.3
Kammanassie Dam	306.3	283.4	112.4	138.8

Table 4.4 Extremely low rainfall events in Hoekplaas rainfall station

Year	1926	1927	1933 ¹⁾	1937 ¹⁾	1949	1958	1960 ¹⁾	1966	1969	1970	1978	1984	1986	1987	1989 ¹⁾	1999
P(mm)	116.8	128.0	172.8	171.0	145.5	143.3	161.0	98.5	110.3	151.5	132.5	114.5	152.0	154.8	174.8	176.8

¹⁾ The events are identified by critical value of 181.3 mm a^{-1} .

Table 4.5 Extremely high rainfall events in Hoekplaas rainfall station

Year	1929 ²⁾	1932 ²⁾	1953 ²⁾	1954 ²⁾	1959 ²⁾	1973	1974	1976	1981	1985	1993 ²⁾	1996
P(mm)	302.3	334.0	293.3	300.0	294.0	330.5	322.0	375.5	517.3	454.3	303.8	452.5

²⁾ The events are identified by critical value of 295.5 mm a^{-1} .

A major variation in the precipitation occurs in the two rainfall stations although they were in the same "homogeneous" rainfall districts. For the Kammanassie Dam rainfall station, there were 8 extremely low and 8 high rainfall events (Table 4.6). The years with extremely high rainfall events

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

may produce more recharge, but the extremely low rainfall years result in less recharge on an annual scale. The individually high rainfall event must however, not be ignored.

Table 4.6 Extreme rainfall events in the Kammanassie Dam rainfall station

Low events	Critical value 112.38 mm·a ⁻¹	Year	1925/ 1926	1935/ 1936	1936/ 1937	1943/ 1944	1977/ 1978	1982/ 1983	1984/ 1985	1997/ 1998
		P (mm)	67	85	36	0	21	41	82	68
High events	Critical value 306.28 mm·a ⁻¹	Year	1950/ 1951	1953/ 1954	1970/ 1971	1980/ 1981	1981/ 1982	1992/ 1993	1996/ 1997	1999/ 2000
		P (mm)	430	498	331	470	325	411	355	361

4.1.3 Rainfall in Mountainous Areas

The rainfall pattern in the mountainous area is the most important issue for the recharge, as almost the entire recharge areas of the TMG aquifer lie in the mountainous areas. Four rainfall stations, namely Parshall, V-notch, Wildebeesvlakte and Purification Works East, were used to perform the CRD analysis for the pattern identification. There are similar rainfall patterns but different in magnitude from September of 1993 to March of 2003 (Figure 4.11). The highest rainfall occurs in the Wildebeesvlakte rainfall station, and the lowest one in the Purification Works East rainfall station.

The CRD patterns as shown in Figure 66 are observed as (1) The patterns of the CRD are similar but they have different magnitudes for the four rainfall stations; (2) The sequence of the CRD patterns from maximum magnitude to the minimum is from Wildebeesvlakte, Parshall, V-notch to Purification works East rainfall stations; (3) The wave patterns occurred in September of 1993 to September of 1996 and (4) The positive trend pattern occurred from September of 1996 to June of 1997, while negative patterns occurred from July of 1997 to May of 2002.

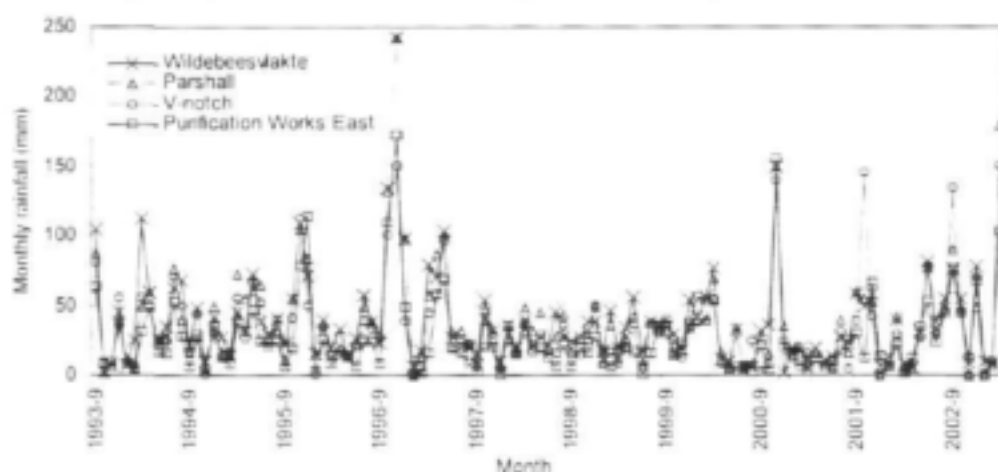


Figure 4.11 Rainfall patterns in the Kammanassie region

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

High correlation coefficients occur between the four rainfall stations, but the F-Test probabilities are low except for those between the Parshall and Wildebeesvlakte rainfall stations (Table 4.7). The significant correlation may occur between the Parshall and Wildebeesvlakte rainfall stations because of the highest probability of 55.8%. From this point of view, the selection of the rainfall station should be worked out carefully if rainfall records are used to estimate recharge. For the Parshall and Wildebeesvlakte rainfall stations, a similar distribution of rainfall frequency occurred in 1994-2003 (Figure 4.13); the frequencies of rainfall events were 20.7% and 26.1% for rainfall higher than 50 mm per month, and 50.5% and 52.3% for rainfall higher than 30 mm per month, respectively.

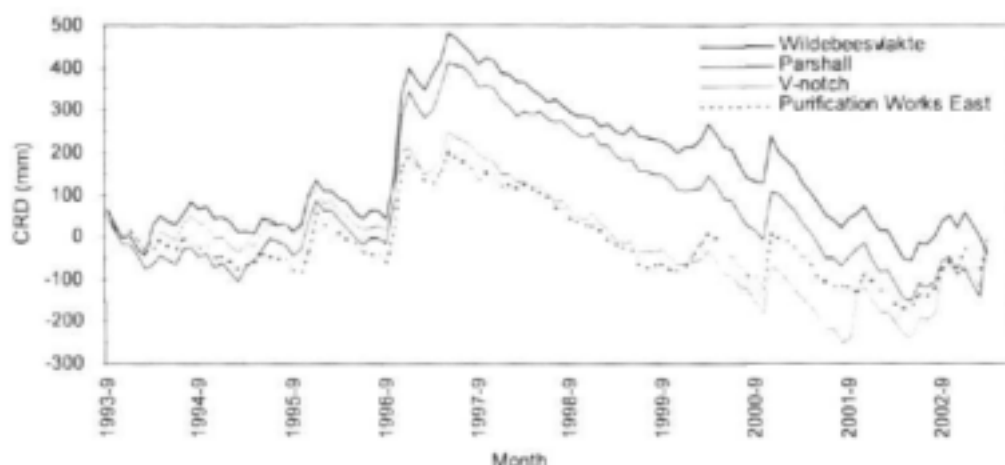


Figure 4.12 CRD patterns in the Kammanassie area

Table 4.7 Cross correlation analysis and F-Test

Rainfall station	Parshall VS Wildebeesvlakte	Parshall VS V-notch	Parshall VS Purification Works East	Wildebeesvlakte VS Purification Works East	Wildebeesvlakte VS V-notch	V-notch VS Purification Works East
Coeff. of correlation	0.98	0.75	0.86	0.81	0.68	0.92
F-Test probability	5.58E-01	8.06E-03	7.32E-09	1.66E-07	3.84E-02	1.20E-03

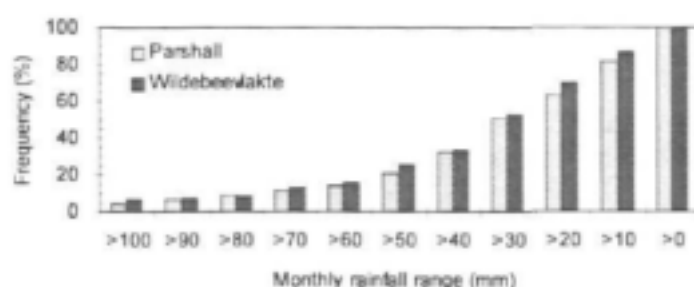


Figure 4.13 Rainfall event frequencies of the Parshall and Wildebeesvlakte rainfall stations (April 1994 to March 2003)

4.1.4 Rainfall and Temperature Relationship

Recharge may be affected by temperature through evapotranspiration. The patterns of monthly minimum and maximum temperatures in the Kammanassie mountainous area are similar to each other. The highest temperature occurs in February, and the lowest one occurs in June. The relationship between rainfall and temperature is illustrated in Figure 4.14. A lower difference in the temperature but higher rainfall occurs in March, April and November, thus the recharge indicated by comparison may also be higher. Although the lowest temperatures occur in June and July, the recharge may be low because of low rainfall.

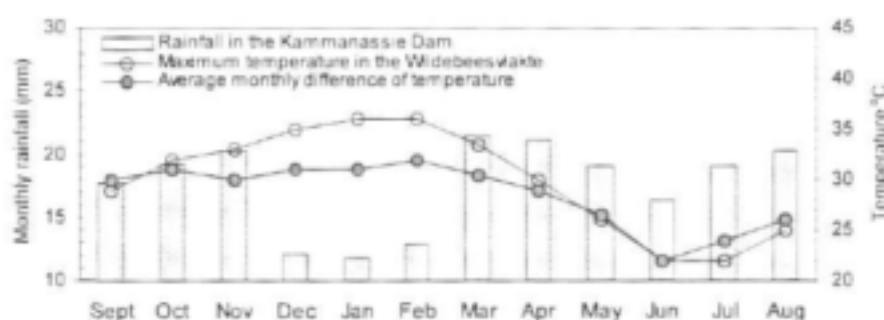


Figure 4.14 Relationship between rainfall and temperature

4.2 Hydrogeology of the Kammanassie Area

An understanding of the characteristics of the fractured rocks of the TMG in the Kammanassie area leads to the understanding of the recharge processes and hence the recharge estimation. The storativity and the connectivity of the TMG aquifer system are major factors responsible for controlling the recharge. The secondary porosity of the Kouga and Peninsula quartzites is estimated to be in the order of 0.1 to 1%; while the secondary porosity of the Vermaaks Keystone Block is thought to be in the order of 5%, and as high as 7% along the faults based on geological conjecture (Hälbich and Greef, 1995, Kotze, 2002). The width of open joints in the Kouga and Peninsula Formations is estimated to range between 1 mm and 10 mm per metre (0.1% to 1.0 % of porosity) and the storativity of the Peninsula Aquifer are from 0.1% to 0.22% (Kotze, 2002). From the west to the east of the Kammanassie Anticline, four different structural domains (Chevallier, 1999) relating to different types of fracturing processes show different fracture patterns, which could imply different recharge processes.

The groundwater regime in the Vermaaks River area is controlled by both rainfall recharge and wellfield development. The water level has declined as deep as 27.54 m - 50.60 m below the initial water level because of a high annual abstraction of 137,000 to 281,000 m³. A number of cold springs which used to occur along the lower western slopes of the Kammanassie Range have dried up since the early 1970's. These still occasionally flow slightly after rainfall events

according to Mr Haasbroek of Overberg water. Three small separate hot springs emanated in the valley floor below the Bokkraal Wellfield until the early 1970's (Smart, 2000).

4.2.1 Physiography

The Western Kammanassie area including Vermaaks River catchment of 21.6 km² and Marnewicks catchment of 25.3 km² within the southern part of the Quaternary catchment (code: J33E) in the Klein Karoo region, is between latitude 33°34' and 33°38' south and longitude 22°30' and 22°42' east (Figure 4.15). The Vermaaks River Well field consists of four production boreholes (VR6, VR7, VR8 and VR11) at depths ranging from 177 to 251.3 m in the upper Vermaaks catchment (10.85 km²) within the Peninsula Formation window (40.6 km²). It is separated by the Cedarberg Formation.

One of the climatic features of the area is the very large diurnal and seasonal fluctuation in temperature. Daily average minimum and maximum temperatures for summer and winter vary between 15 to 42 °C and -3 to 18.5 °C, respectively. The annual precipitation range is between 299 mm and 714 mm in the Parshall rainfall station from 1994 to 2002. Average annual evaporation varies between 1760 mm a⁻¹ and 2050 mm a⁻¹. The study area comprises a narrow valley with an elevation approximately between 500m and 1340 m above mean sea level (a.m.s.l.) surrounded by the Kammanassie Mountains, which have a maximum altitude of 1950 m a.m.s.l. at the south. The main drainage in this region is westward, where the perennial Olifants River drains the area to the north of the Kammanassie Mountain Range. The Vermaaks River and Marnewicks River, which have a relatively short tributary, are ephemeral in the steep upper reaches, with more sustained flow in the lower reaches, drain northward into the Olifants River.

The slope gradients are measured at less than 10°, 10-20°, 20-30° and greater than 30° (Figure 4.15). The slope gradient less than 10° distributes along the ridge of the mountain. The 10-20° slopes extensively spread in the Kammanassie area. The slopes with gradient 20-30° stretch along the lower Range. The slopes with gradient greater than 30° extend along the banks of the Vermaaks and Marnewicks valleys. The extent of these slopes makes them suitable for rainfall infiltration except for the banks of the steep valley.

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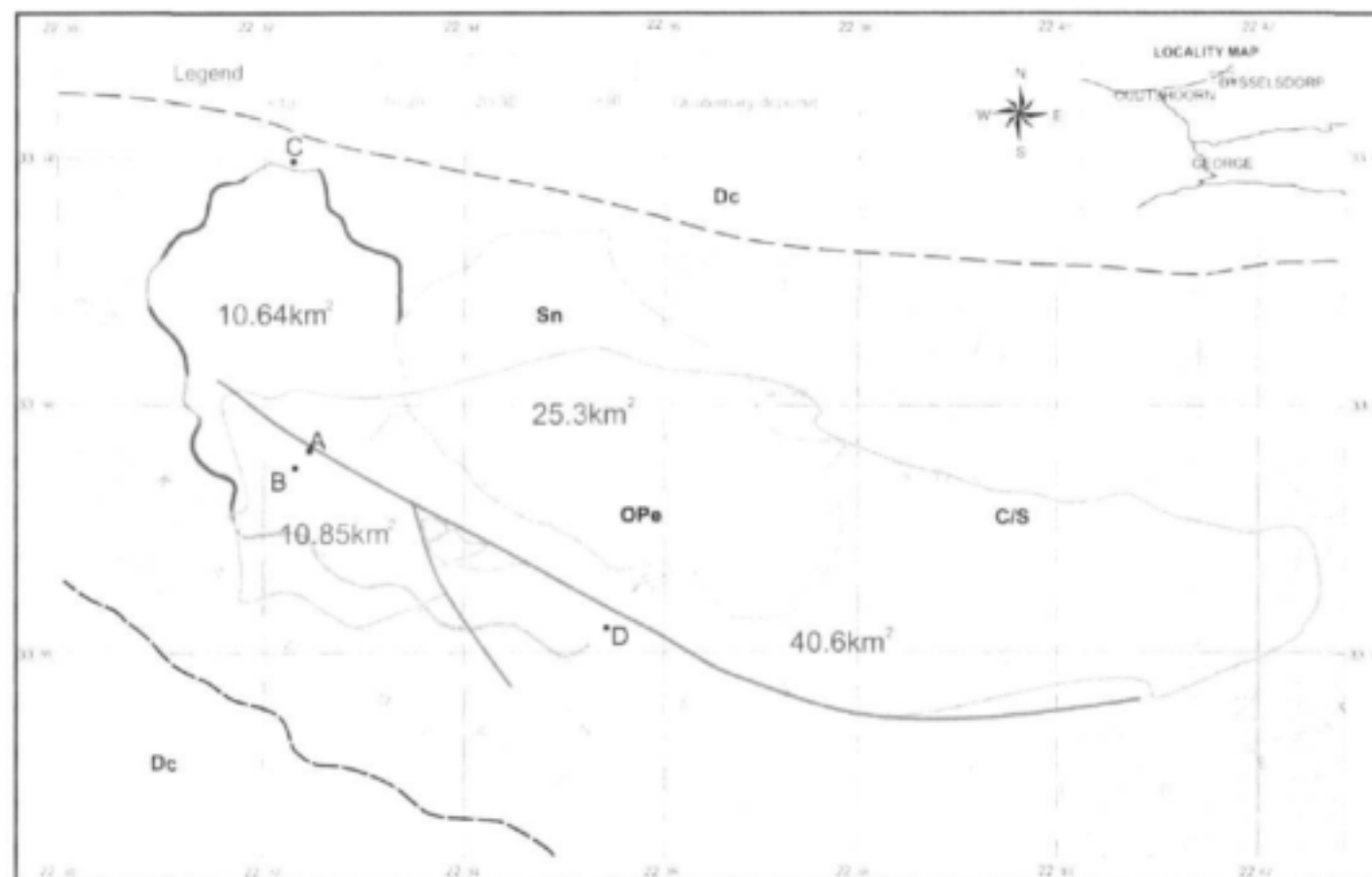


Figure 4.15 Physiography in the Western Kammanassie area

4.2.2 Aquifers

The aquifers consisting of Peninsula and Nardouw Formations, strike EW, coinciding with the axial plane trace of the Kammanassie mega anticline. The outcrops of the aquifer form the main recharge area of the Kammanassie aquifer (Figure 4.16).

The Peninsula aquifer distributes along the mountain crests and comprises two thirds of the total thickness of the TMG (1 800 m to 2 150 m, Kotze, 2002). The recharge areas of the Peninsula aquifer are the outcrops of the Peninsula Formation, which are composed of a uniform succession of medium to coarse grained, thickly bedded, grey sandstones, characterised by cross bedding (Figure 4.17). The recharge areas are approximately 40.6 km².

The Nardouw aquifer distributes around the Peninsula aquifer, which outcrops in the centre of the Kammanassie mega anticline. The rocks of the Nardouw Subgroup are, in general, more brownish on weathered surfaces and thin intercalations of shale are more plentiful than in the Peninsula Formation (Figure 4.18).

The Nardouw formation becomes more feldspathic towards the top. Due to this higher shale content they are less competent than the Peninsula Formation and deformation tends to be more ductile, giving rise to the spectacular fold geometries. Much more complicated characteristics of the aquifer and inhomogeneous recharge occur in the Nardouw aquifer system. The thickness of the Nardouw aquifer cannot be estimated due to the complex folding of the formation.

The Cedarberg Formation comprises a 50 m to 120 m thick shale layer separating the Peninsula and Nardouw (Kotze, 2002). It is a clear marker horizon because it weathers deeply to a smooth outcrop, which contrasts with the grey crags of the sandstones above and below (Figure 4.19). The shale is greenish and extremely fine grained where fresh. It forms a clear aquitard between the Peninsula and Nardouw aquifers.

Alluvial and slope deposits, consisting of sand, gravel and other unconsolidated materials, are distributed at the foot of the mountain and along the valley floor. The thickness of these deposits is up to 15m (Kotze, 2002). The TMG aquifer is likely to receive infiltration from both open fracture networks of the outcrop rocks and the loose deposits.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

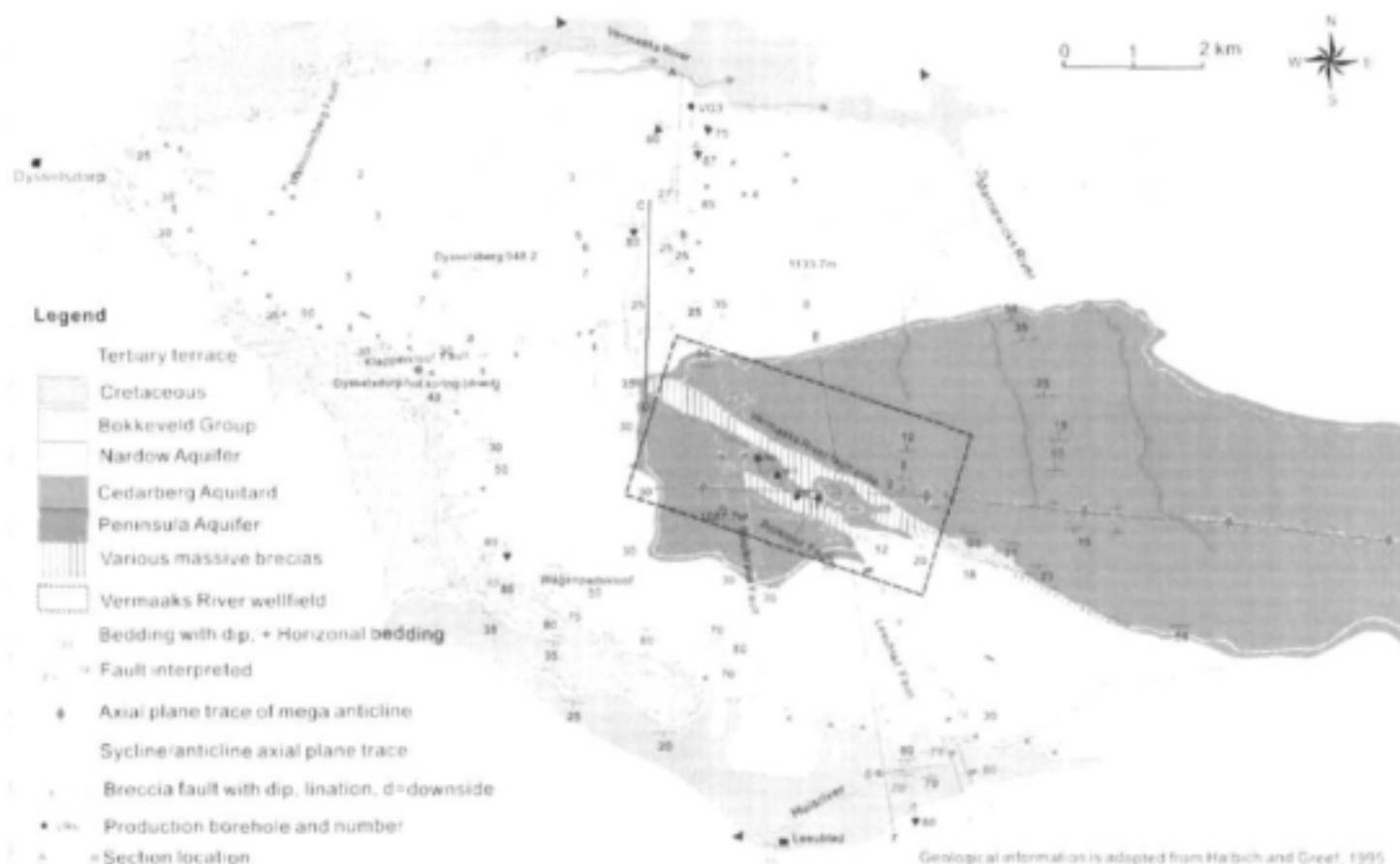


Figure 4.16 Hydrogeological map in the Kammanassie area

The Kammanassie Mountain is an eroded remnant of resistant TMG quartz arenites, also referred to as the Kammanassie mega-anticline, which was reconstructed (Hälbich and Greef, 1995). In the Nardouw subgroup, that is the Kouga, Baviaanskloof and most of the lower part of the Bokkeveld group (Gydo Formation), the folds are overturned. Due to the repetition of small-scale folding, the outcrop length of the Kouga Formation is particularly long (Woodford and Chevallier, 1998). The Vermaaks fault cut the mega anticline along NW-SE strike (see to Figure 4.16). The Vermaaks fault comprises two parts: the centre zone is cemented breccias (silicification), where it's impermeable; the zones beside the centre zone may have high permeability due to branch faulting. The high mountainous areas form the recharge area and the hydrodynamic head. The contact between the Cedarberg shale and Peninsula Formation is a natural discharge zone at valley floor. The regional flow seems to occur at high depth. The faults linking the Peninsula and Nardouw aquifers may be leakage zones.

4.2.3 Well fields

There are 9 production boreholes (Appendix I), 27 unsuccessful boreholes (Appendix II) and 19 boreholes with no detailed data in the study area. Their location is shown in Figure 4.20. Boreholes VR6, VR7, VR8 and VR11 are the only boreholes drilled in the Peninsula Formation.



Figure 4.17 Peninsula sandstone in the Vermaaks River valley

The production boreholes, namely DP10, DP12 and DP29 in Varkieskloof Wellfield, DP28, DP15 and DP25 in Bokkraal Wellfield and VG3 in Voorzorg, were drilled in the Nardouw Subgroup. From the information on boreholes, two or three water strikes with thicknesses of 1.5-15m were found along the 250 m depths of the boreholes. The first water strike is usually drilled below 100 m depths that is from 107m to 228m. The total thickness of the entire water strike is from 7m to 23m.



Figure 4.18 Nardouw Subgroup in the Vermaak's River valley

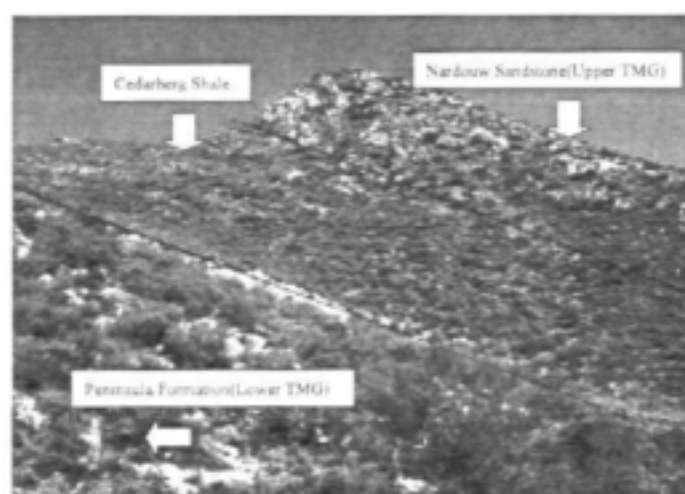


Figure 4.19 Cedarberg shale near the G40171 borehole

4.2.4 Monitoring networks

The unsuccessful production boreholes of the KKRWSS, are currently being used as monitoring boreholes (Appendix II). The boreholes, namely DG, are drilled near the Droekloof production borehole DG110. VR5 and WN101 are situated in the Nardouw Aquifer (Baviaanskloof Formation, just south of the Marnewicks v-notch). VG12 is an important monitoring borehole, midway between the VG3 and the Cedarberg shale layer in the Vermaak's River Valley. DP27 is situated near the Varkieskloof production boreholes, and DP13, 14 and 20 near the Bokkraal production boreholes (Figure 4.20).

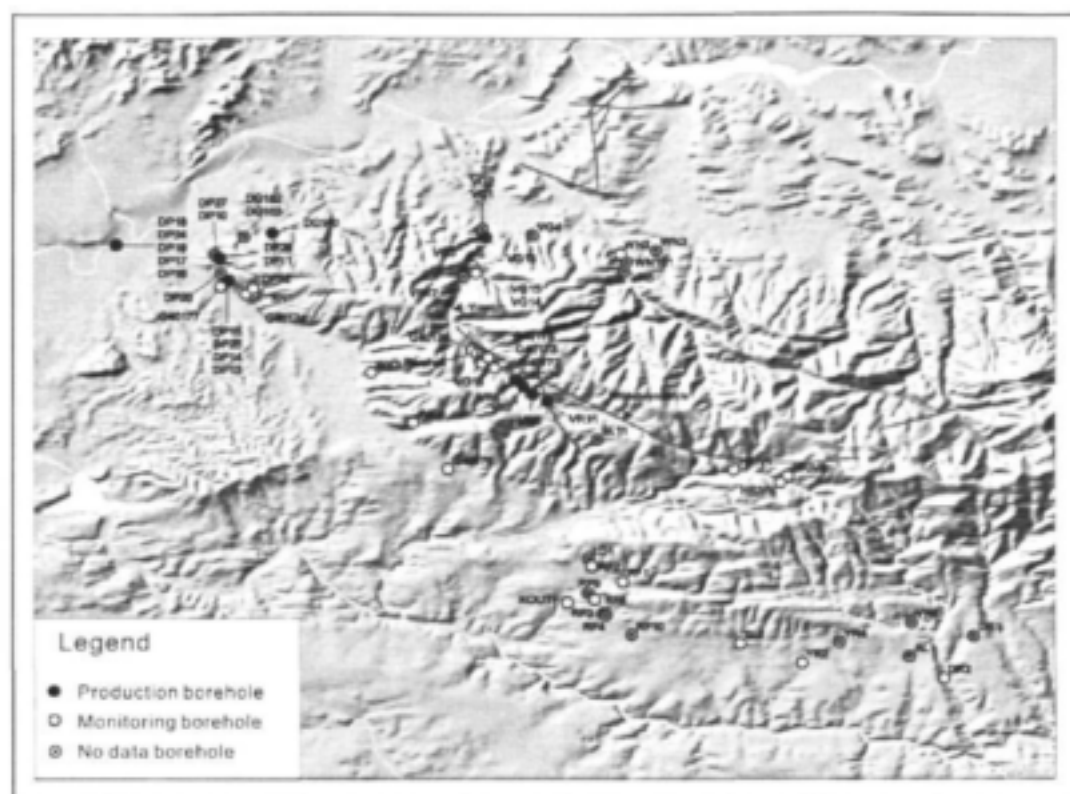


Figure 4.20 Borehole site in the Kammanassie area

The monitoring boreholes (Appendix III) have their individual purpose and specific monitoring requirements. G40171, G40172 and G40173 are shallow boreholes drilled on the farm Voorzorg 124, upstream of the Cedarberg shale layer to monitor the inter-relationships between alluvial and TMG groundwater. G40174 was drilled on the farm Rietfontein 142 to monitor the water level responses to abstraction from private abstraction borehole RN1. G40175 was drilled to a depth of 126 m on the farm Voorzorg 124, it subsequently collapsed and was grouted. It was replaced by the nearby G40175A, next to the abandoned G40175. The purpose of this borehole was to investigate the effectiveness of the Cedarberg shale layer as an aquitard. G40176 was drilled on the farm Voorzorg downstream of the Cedarberg shale layer. G40177 was drilled south of the Bokkraal Well field to monitor water level responses to abstraction from the Well field. G40178 was drilled on the southern flanks of the Kammanassie Mountain to monitor long-term water level fluctuations arising from private and commercial abstractions from the Vermaak's River Well field. The use of these boreholes seems not to be effective to monitor the water level fluctuation in responding to the abstraction in the Kammanassie area at different scale.

4.2.5 Water level fluctuation in the Vermaaks River Wellfield

The groundwater levels of the production boreholes have been declining since 1984 in the Vermaaks catchment (Figure 4.21). Only the water level of VG16 responds to that of the production borehole. The other monitoring boreholes seem to have independent fluctuations. There are about 26 private boreholes with a total annual abstraction of $1.85 \times 10^6 \text{ m}^3 \text{ a}^{-1}$ in the Kammanassie Mountain area (Kotze, 2002). This makes the water level fluctuations in response to rainfall more complex and uncertain.

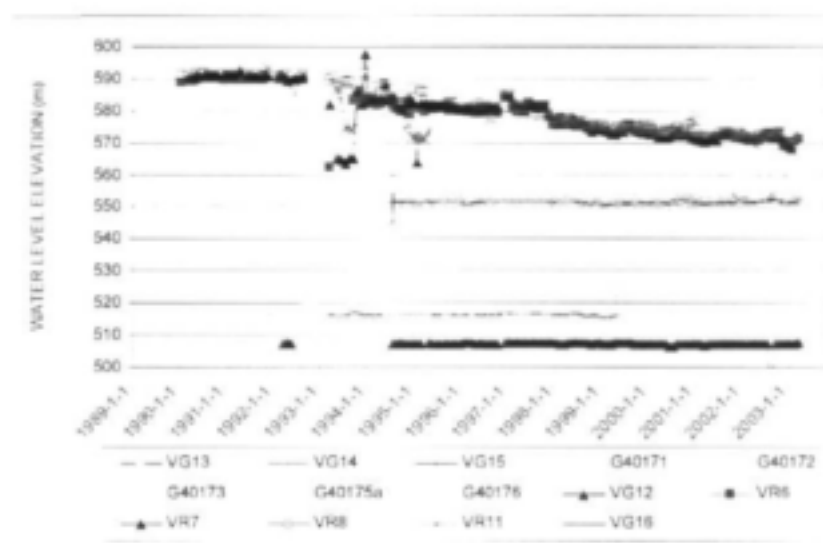


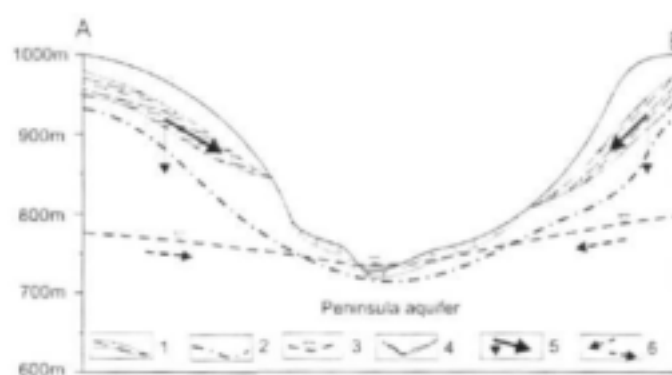
Figure 4.21 Water level fluctuation in the Vermaaks area

4.2.6 Conceptual Hydrogeological Model

A weir for spring monitoring was built on the Vermaaks River (Figure 4.22). The perennial pond suggests beds of relatively low permeability at the bottom. There are no perennial springs in the catchment except the spring in the monitoring station. The relationship between weathered zones and fractured rock aquifer is illustrated in Figure 4.23 and Figure 4.24. The shallow zone is extensively weathered with fractures partly filled in by clayey materials. Groundwater outflow from this zone is in the form of gravitational springs and seepage zones. The pond is maintained in part by overflow from the Peninsula aquifer and a localized shallow aquifer. The artesian borehole G40175A, which is drilled through the Cedarberg shale, may confirm this. Due to the abstraction from the Vermaaks River Well filed, the spring (code: 051) dried up, and the water level dropped lower than the first water strike (Figure 4.24).



Figure 4.22 Monitoring station in the Vermaaks catchment



Box 1: weathered zone with high hydraulic connectivity;
Box 2: weathering limit; Box 3: water level; Box 4: new deposit;
Box 5: infiltrating direction; Box 6: flow direction

Figure 4.23 Hydrogeological section along transect A-B
(See Figure 4.15 for location)

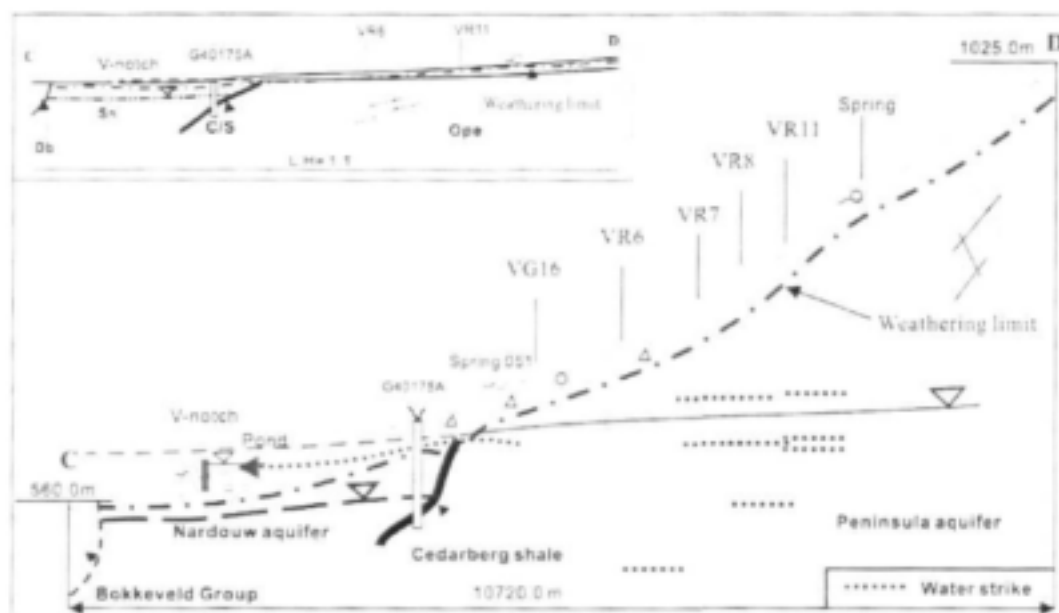


Figure 4.24 Hydrogeological section along the Vermaak's River

4.2.7 Fracture System and Recharge

4.2.7.1 Fracture feature and connectivity

The E-W and N-S fracture systems have different characteristics, in terms of fracture geometries, fracture length and connection. The fracture network consists of (a) E-W and (b) N-S trending fracture systems. The fracture density seems to be high, especially along major faults.

The NNW-SSE fractures and lineaments are well developed in the TMG and are likely to allow groundwater flow. The fractures trending NNE-SSW and fractures reactivated in E-W, and N-S orientations are the preferred orientations for groundwater recharge and flow (Kotze, 2002), such as the Rooikrans and Leeublad Faults. Higher recharge rates may occur along these faults.

4.2.7.2 Porosity and storativity

Porosity or storativity of an aquifer is a major factor in many methods for recharge estimation. There are large irregular spatial differences of the porosity or storativity in an aquifer. The accuracy and variation of the porosity or storativity is related to the estimation methods. This would affect the accuracy of the recharge estimation.

The Peninsula aquifer is a pure quartz arenite with a very low primary porosity due to the cementation of individual sand grains and recrystallisation. However, increased rock induration

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led to a higher potential for brittle fracturing, as well as higher fracture frequencies. The Nardouw Subgroup contains more silty and shaley interbeds and is associated with high feldspar content. Shale layers have a greater impact on the fracturing and folding style of Nardouw Aquifers, which gives rise to large variations in the storativity.

For beddings, joints, fractures and massive fault block in the Vermaaks area, the secondary porosity was estimated up to 5% by Kotze (2002), even in fault zone as high as 7% (Hälbich and Greef, 1995). However, it is observed these figures seem to be overestimated. For discussion purposes, the fracture state of 5% porosity of rock mass is theoretically equal to one fracture of 1000 mm × 1000 mm × 50 mm or one pipe of 252.3 mm diameter in 1 m³ rock mass (Figure 4.25).

The theoretical results based on the fracture model are listed in Table 4.8. The width of a single fissure in a 1m³ rock mass with 0.1%, 1% and 5% porosity and 1, 2, 3, 6 and 9 fractures are from 1-0.111, 10-1.11 and 50-5.556 mm under a parallel condition, respectively; the width of a fracture should be wider under cross condition. These theoretical results show that 1%-5% porosity may not be correct in the TMG aquifer, especially at depth

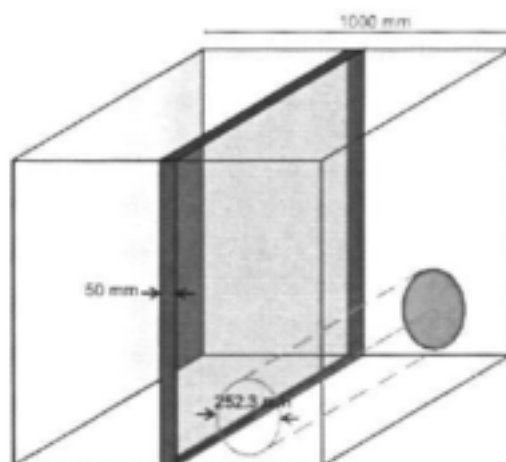


Figure 4.25 Fracture model with porosity of 5% in 1 m³ rock mass

Table 4.8 Theoretic aperture of fracture in 1m³ cubic rock mass with gross section of 1m²

Items of fracture	Condition	Pipe Diameter (mm)	Parallel (mm)					Cross (mm)			
								1×1	1×1×1	2×2×2	3×3×3
Porosity %	Volume of fracture m ³	1	1	2	3	6	9	2	3	6	9
0.1	0.001	3.568	1.00	0.50	0.333	0.167	0.111	0.500	0.333	0.167	0.111
1	0.01	11.284	10.00	5.00	3.333	1.667	1.111	5.013	3.345	1.672	1.115
5	0.05	252.313	50.00	25.00	16.667	8.333	5.556	25.321	16.954	8.477	5.651

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The ranges of transmissivity (T) and storativity (S) in the Vermaaks River Wellfield are listed in Table 4.9. These also illustrate low storativity does not match the supposed high porosity. The accredited storativity value should be from 0.01% to 0.1% based on the discussion above.

Table 4.9 Ranges of transmissivity and storativity in the Vermaaks River Wellfield (After Kotze, 2002)

Conditions	$T_{min}(m^2 \cdot d^{-1})$	$T_{max}(m^2 \cdot d^{-1})$	$T_{aver.}(m^2 \cdot d^{-1})$	S_{min}	S_{max}	$S_{aver.}$
Extreme	7	424	103.9	1.0E-3	2.2E-3	1.35E-3
Condition A	5	144	61.25	1.0E-3	2.2E-3	1.08E-3
Condition B	29	424	191.72	1.0E-3	2.2E-3	1.08E-3
Condition C	17	276	178.92	1.0E-3	2.2E-3	1.08E-3
Condition D	7	161	90.86	1.0E-3	2.2E-3	1.08E-3
Condition E	7	161	90.86	1.0E-3	2.2E-3	1.96E-3

4.2.7.3 Geological origin of the hot springs

The hot spring at Calitzdorp is located at the intersection between faults in the TMG and the basal shale layer of the Gydo Formation (Bokkeveld Group), occurring where the ENE-WSW trending faults connect with the NE-SW trending faults near the base of the Bokkeveld Group (Figure 4.26). The site of the hot spring should be the regional discharge area. Figure 4.27 shows the site of the Dysseldorp hot spring (dry), which is also located at the intersection between faults in the TMG and Bokkeveld Group. Three small separate hot springs emanated within an area of approximately 100 m² in the valley floor below the Bokkraal Well field until the early 1970's. The water temperature was indicative of deep circulating groundwater, which is up-welling along a geological fault/fracture (Smart, 2000). Andreoli *et al.* (1996) showed the positions of epicenters of a post Karoo rock age in South Africa. The positions of epicentres seem to correspond well with the locations of the hot springs. The Dysseldorp hot spring dried and this could not only be associated with the earthquake, but also with a large abstraction near the site and climatic changes (Earthquake: A, Tulbagh in 1969, Magnitude: 6.5. B, Oudtshoorn: Date: 28 October 2001, Latitude: -33.542N, Longitude: 22.228E, Magnitude: 3.6, Depth: 5.0 km).



Figure 4.26 Calitzdorp hot spring
(The painting from Calitzdorp Spa)



Figure 4.27 Site of dried Dysseldorp hot spring

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

A model for relating interflow, local flow and regional flow is conceptualised in figure 4.28. The interflow occurs in the mountainous areas at elevations of 220 to 1950 m. The local flow may occur at depths of 220-1700 m, but regional flow can occur at depths related to the distribution of the TMG aquifer. The regional discharge area is located at Calitzdorp Spa based on the following:

- Two hot spring discharge points were found near the Kammanassie area, one is the dried Dysselsdorp hot spring which used to occur at an elevation of 500 m at the western foothill of the Kammanassie Mountain, this still indicates that the regional flow direction is from east to west in the Kammanassie area. The other one is the Calitzdorp hot spring located at an elevation of 220 m in the western border of the Oudtshoorn Basin.
- The north and northeast boundaries are the Congo Fault and the Baviaanskloof Fault, which are regarded as the hydrogeological boundaries. There are no hot spring discharge zones or points along the faults.
- From a stratigraphical point of view, the Congo Group is the bottom strata of the southern boundary in the Outeniqua Mountain. No discharge zones or points of hot springs are found.
- From a topographical point of view, the highest peak (1950 m a.m.s.l) is located at Kammanassie Mountain. The main drainage (Olifants River) in this region is westward. The Calitzdorp hot spring discharge is at an elevation of 220 m located on the left bank of the Olifants River. This point is almost the lowest point of the basin.

It is important to note that the Calitzdorp hot spring can be fed not only by the recharge from the Kammanassie area, but also by the southern mountainous area.

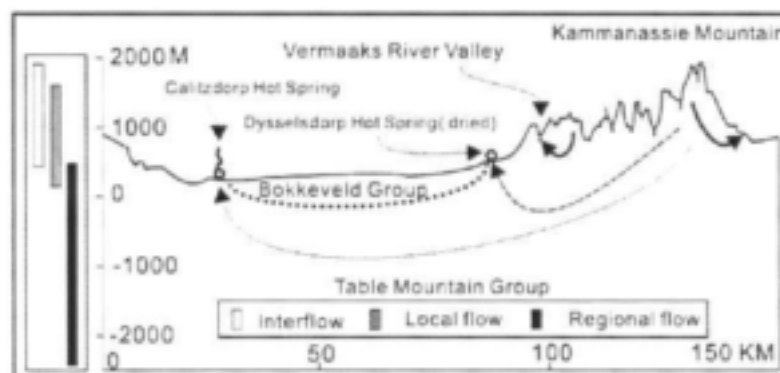


Figure 4.28 Relation between interflow, local flow and regional flow

4.2.7.4 Recharge area

For the Vermaak's River Well field, analysis of satellite imagery and hydrogeological settings suggests that recharge probably takes place within the outcrops of the Peninsula Formation

window of 40.6 km². The Cedarberg shale and the brecciated fault form the recharge boundary of the Vermaaks River Well field. High density of fracture intersections and frequencies appear to facilitate recharge to the deep Peninsula aquifer system.

4.3 Hydrogeochemical Characteristics in the Kammanassie Area

Hydrogeochemical data can be potentially used to aid recharge estimation in the study area as is demonstrated previously in other regions (Plummer *et al.*, 1990; Pitkänen *et al.*, 1999; Epstein and Mayeda, 1953; Kennedy *et al.*, 1986; Maule *et al.*, 1994; Mayo and Loucks, 1995; Liu *et al.*, 1995; Simpkins, 1995; Wood, 1999; Fritz *et al.*, 1976). Isotopes and chlorofluorocarbons (CFC) can also be used as tracers for interpreting the groundwater dating and recharge estimation (Mazor, 1991; Cook and Solomon, 1997).

4.3.1 Outline of Hydrogeochemical Characteristics

At the start of KKRWSS in 1984, the chemical samples were irregularly obtained. Some chemical data were collected from the National Groundwater Data Base (NGBD) of South Africa in 1992 and 1994. Concentrations of some constituents are given in monthly reports of the Overberg Board Water Management since 1994. Apart from the existing data from Kotze (2002) and NGDB (from the Department of Water Affairs and Forestry (DWAF), Bellville), the data used in this study include new measurements of rain water, and in addition, borehole samples and soil samples from the Vermaaks valley were collected for this project between 2002 and 2004. More than 40 sets of water samples were collected from boreholes and springs across the area. The sampling sites in the study area are shown in Figure 4.29. The constituents of interest include pH, TDS, EC, anions (F⁻, Cl⁻, HCO₃⁻, SO₄²⁻, PO₄³⁻, NO₃⁻), cations (Na⁺, K⁺, Ca²⁺, Mg²⁺), nitrogen components, Fe, Mn, Sr, Ba, SiO₂ and isotopes of O-18, H-2, C-14, C-13.

The results of the analysis of water samples in the Vermaaks area are presented in Appendix IV. The main hydrogeochemical characteristics is summarised as follows:

- The average TDS in production boreholes of the Peninsula aquifer ranges from 48 to 60.4 and that of the Nardouw is 86.8 mg l⁻¹. The TDS of the monitoring boreholes varies from 52 to 71.4 mg l⁻¹ when pH<7 and from 96 to 110 mg l⁻¹ when pH>7.
- The TDS values of spring water in the V-notch, rainfall and interflow (VS1, VS2, VS3) are 79, 136.5 and 144-266.8 mg l⁻¹, respectively. Na⁺ and Cl⁻ are predominant; their concentration ranges are 11.90-56.56 mg l⁻¹ and 16.9-83.7 mg l⁻¹, respectively.
- The concentration ranges of HCO₃⁻, SO₄²⁻ and NO₃⁻ are 6.0-152.2 mg l⁻¹, 0.01-12.3 and 0.01-2.07 mg l⁻¹, respectively.
- The concentrations of K⁺, Mg²⁺ and Ca²⁺ range 0.8-6.51 mg l⁻¹, 1.6-13.46 mg l⁻¹ and 1.6-20.88 mg l⁻¹, respectively.
- Water samples with pH>7 are very few with the maximum being 7.43.
- Water quality in the V-notch is similar to that in production boreholes.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

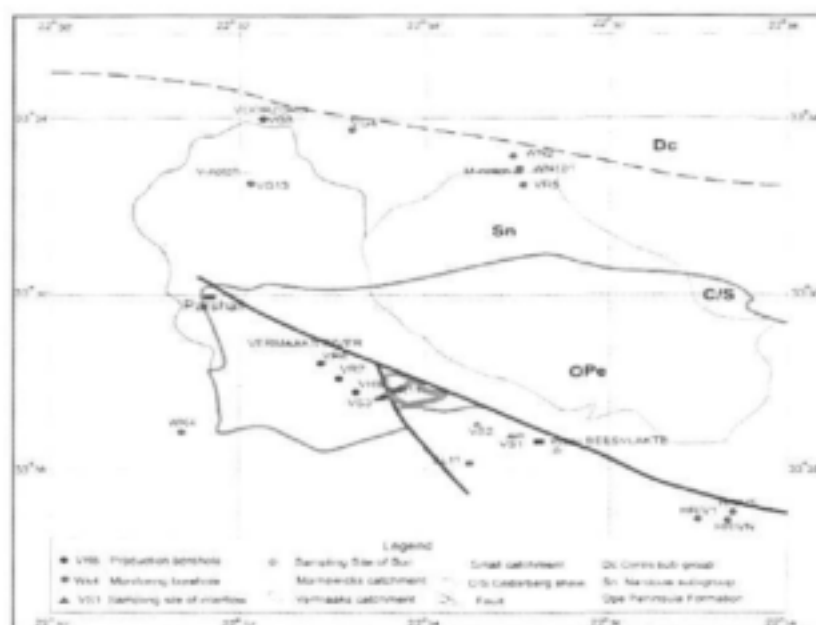


Figure 4.29 Sampling sites in the Kammanassie area

4.3.2 Chemical Characteristics of Chloride and Sodium

The chloride and sodium are very important ions in the study area. The chloride concentration was discussed in detail because it was used as a tracer in the recharge estimation.

4.3.2.1 Chloride

Some chemical characteristics of the chloride in the Vermaaks area is presented in Table 4.10 and shown in Figure 4.30 and Figure 4.31. The observations are obtained as follows:

- 1) In terms of concentration, the sequence is $\text{Cl}^- > \text{Na}^+ > \text{Mg}^{2+} \approx \text{Ca}^{2+}$.
- 2) The chloride concentration in the boreholes is half of that of the interflow.
- 3) The ranges of chloride concentrations in VR6, VR7, VR8, VR11 and VG3 are 19.0- 60.0, 18.0- 46.0, 17.0-56.0, 18.0-50.0 and 20-86 $\text{mg}\cdot\text{l}^{-1}$, respectively; their average values are 34.97, 33.26, 30.78, 34.41 and 56.48 $\text{mg}\cdot\text{l}^{-1}$, respectively.
- 4) The average value of chloride in production boreholes of VR6, VR7, VR8 and VR11 is 32.51 $\text{mg}\cdot\text{l}^{-1}$. The concentration in the Nardouw aquifer is almost twice that of the Peninsula aquifer.
- 5) The chloride concentrations are less than 50 $\text{mg}\cdot\text{l}^{-1}$ in most boreholes, but the concentrations in SL2, G40178 and G40177 are much higher than that of the others in the Nardouw aquifer (Figure 4.31).

Table 4.10 Statistical results of chloride in the Vermaak's area

Statistical result	VR6	VR7	VR8	VR11	VG3	VR6, VR7, VR8 and VR11
Count	88	92	81	65	86	326
Min	19	18	17	18	20	17
Median	34	32.5	30	30	56	32
Max	60	46	56	50	86	60
Average	34.97	33.26	30.78	31.41	56.48	32.5
Geomean	33.96	32.27	29.90	30.62	53.38	-
Harmean	33.27	31.60	29.21	29.87	50.73	31.03
Standard deviation	7.09	6.38	6.95	7.20	13.62	7.03
Confidence	1.24	1.09	1.27	1.47	2.42	0.64

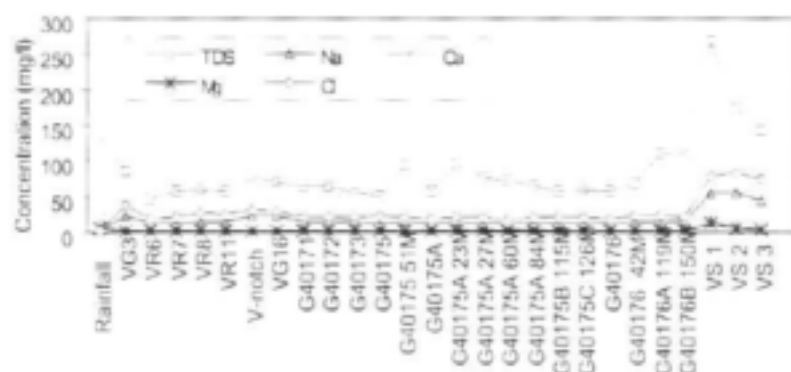


Figure 4.30 Fingerprint of the major ions in the Vermaak's area

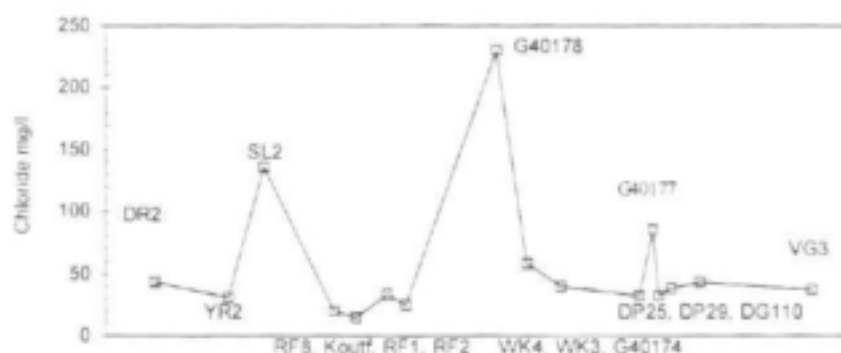


Figure 4.31 Distributions of chloride in the Nardouw Subgroup

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

4.3.2.2 Sodium

The sodium concentrations are smaller than chloride concentrations (Figure 4.32). Considerable differences in range of ion ratios in groundwater, rainwater and soils are listed in Table 4.11. The $\gamma\text{Na}/\gamma\text{Cl}$ ratios of groundwater are close to 1, but the ratios in soil are very low due to ion exchange. Large range of the ratios of other ions occur in the Vermaaks area, such as $(\gamma\text{Na}-\gamma\text{Cl})/\gamma\text{SO}_4$, $(\gamma\text{Cl}-\gamma\text{Na})/\gamma\text{Mg}$. Most of the SAR values in groundwater are less than 1.5. The SAR values greater than 1.5 occur in VG3, VG16 and interflow. Large ranges in SAR and $\gamma\text{Mg}/\gamma\text{Ca}$, but small variation in $\gamma\text{Na}/\gamma\text{Cl}$ can be seen in Figure 4.33.

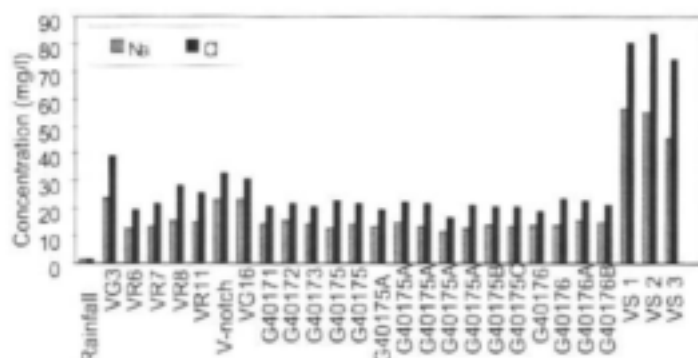


Figure 4.32 Column diagrams of chloride and sodium

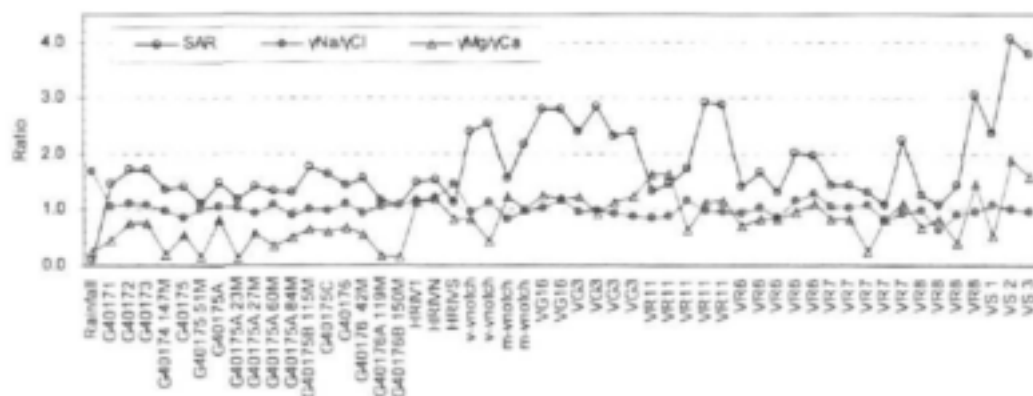


Figure 4.33 Diagram of ion ratios in the Vermaaks area

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

Table 4.11 Ratio of ions in the Vermaak's area

Item	Groundwater					Rain water	Soil
	SAR ¹⁾	$\frac{\gamma Na}{\gamma Cl}$	$\frac{\gamma Na - \gamma Cl}{\gamma SO_4}$	$\frac{\gamma Cl - \gamma Na}{\gamma Mg}$	$\frac{\gamma Mg}{\gamma Ca}$	$\frac{\gamma Na}{\gamma Cl}$	$\frac{\gamma Na}{\gamma Cl}$
Min	0.097	0.631	-2.751	-1.958	0.136	0.071	0.006
Median	1.560	0.996	-0.033	0.033	0.825	0.271	0.016
Max	5.334	1.702	4.643	2.784	1.896	1.703	0.018
Average	1.905	1.011	-0.019	0.059	0.808	0.271	0.014

$$^1) SAR = Na / [(Ca + Mg) / 2]^{0.5}$$

4.3.2.3 Correlations between chloride, EC and TDS

Significant differences of chloride and sodium occur between groundwater and rain water (Table 4.12). This illustrates the components have been changed due to reaction or additional constituents added during recharge processes. Some relationships seem to occur among the chloride concentration, EC and TDS values (Figure 4.34 and Figure 4.35), but the F-Test probabilities are 3.55×10^{-10} and 3.65×10^{-53} , respectively. For purposes of quick estimation of water quality, the correlation equations between chloride, EC and TDS are constructed as follows:

$$TDS = 7.0306EC - 35.503 \quad R^2 = 0.9911 \quad (4.1)$$

$$Cl = 1.9238EC - 1.3913 \quad R^2 = 0.966 \quad (4.2)$$

Table 4.12 Statistical results of chemical data in the Kammanassie area (mg l⁻¹)

Item	Groundwater												Rain water	
	pH	TDS	*EC	Na	K	Mg	Ca	F	Cl	NO ₃	SO ₄	PO ₄	Na	Cl
Min	3.7	12	2.8	2	0.2	0.6	0.4	0.05	4	0	1.9	0.001	0.38	1.74
Max	8.4	1310	182.0	381	22.5	63.0	87.0	1.73	367	11.75	199.0	0.203	1.88	28.9
Median	6.6	89	17.8	20	1.4	2.9	3.5	0.13	32	0.10	12.2	0.011	7.19	14.5
Mean	6.5	187.2	31.9	38.7	3.9	6.2	10.8	0.20	60	0.30	24.5	0.000	1.88	14.5
Standev.	0.8	259.4	37.6	58.7	5.1	9.5	16.8	0.20	74	1.30	35.2	0.000	1.75	5.76
Confidence	0.1	43.8	6.3	9.9	0.9	1.6	2.8	0.00	12.4	0.20	6.0	0.000	0.03	0.4

*EC in mS/m

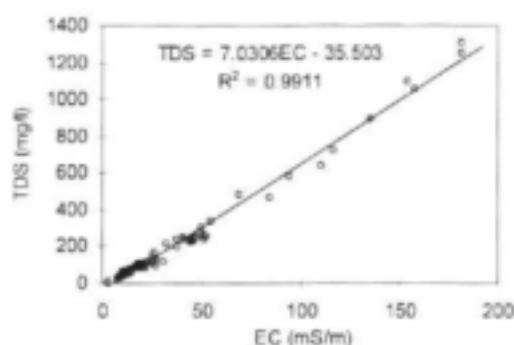


Figure 4.34 EC versus TDS

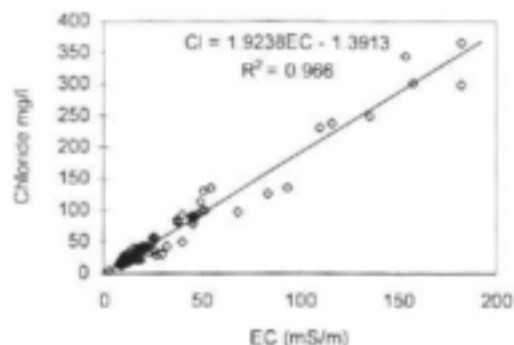


Figure 4.35 EC versus chloride

4.3.3 Isotopes

4.3.3.1 Deuterium and oxygen-18

Isotopic characteristics can be used to discriminate water origins. Values of ^2H or ^{18}O are more negative if the source of the water vapour at the origin of the precipitation is further away (distance effect). If a large part of the water vapour is precipitated from the cloud (reservoir effect) or if the temperature is lower. It can be ascribed to impoverishment in heavy isotopes during successive precipitations, due to depletion of the water vapour from the cloud or to a higher value of the equilibrium fractionation coefficients for oxygen 18 and deuterium with lower temperature. Deuterium excess is the independent term of the equation $D = 8^{18}\text{O} + d$ corresponding to what is referred to as the Meteoric Water Line (Vandenschrick *et al.*, 2002). If evaporation rates are high, because of high temperature and low relative humidity in the atmosphere during the formation of the water vapour, there is a strong kinetic isotopic effect, and d becomes higher (Jouzel and Merlivat, 1984). Deuterium excess is usually considered as a more or less conservative property in the part of the atmospheric water cycle beginning with water vapour formation by evaporation to rainfall just below cloud level (Ciais and Jouzel, 1994).

The Kammanassie MWL ($D = 7.4^{18}\text{O} + 9.1$, $R^2 = 0.75$) d -value is 9.1. Slopes close to 8 are displayed for the Kammanassie area (Figure 4.35), indicating a common origin for all events or at least similar kinetic conditions at the water vapour source (relative air humidity and temperature).

Interpretation of the isotopic signature of the different water types in the Kammanassie area is facilitated by a discussion of the regional precipitation. Stable isotopes data are adopted from NGDB and Kotze (2001) as listed in Appendix V. A comparison of the environmental isotope data for various surveys showed more depleted $\delta^{18}\text{O}$ signatures for July 1997 than those for the same borehole during the survey of January 1996. These differences are postulated to be the result of the effect of the rainfall amount. The ranges ^{18}O and ^2H are shown in Figure 4.36. The observations are as follows:

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

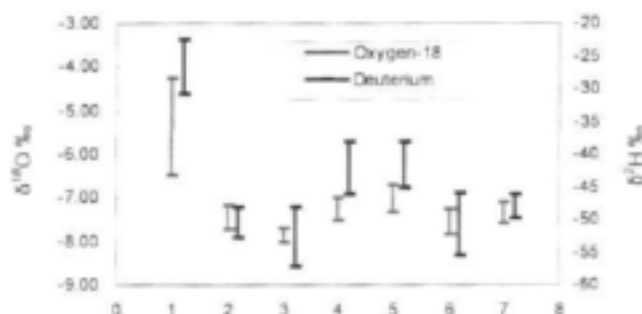


Figure 4.36 Ranges of the ^{18}O and ^2H in the Kammanassie area

1 rainfall, 2 production boreholes of Peninsula Aquifer, 3 production boreholes of Nardouw Aquifer, 4 monitoring boreholes of Peninsula Aquifer, 5 monitoring boreholes of Nardouw aquifer, 6 V-notch and M-notch, 7 hot springs near Kammanassie area

- The isotope signature in production boreholes is different from that of monitoring boreholes. ^2H in production boreholes is more affected by evaporation or altitude than those in monitoring boreholes. It illustrates that different sources or infiltration paths occur.
- Isotopes of the streams in the Vermaaks and Manewicks are clearly concentrated on evaporation.
- Isotope patterns of hot springs differentiate from those of boreholes. The water in the borehole recharges from low elevation areas (more negative value), but the hot springs recharge from high elevation areas or low elevation during a colder period. This would imply the flow paths are different. The flow of the boreholes is local, while the hot spring flow is regional.

The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ relationships for groundwater and rainfall in the Kammanassie area are presented in Figure 4.37. The following can be observed:

- Three groups are identified, which are A: rainfall, B: alluvial aquifer and C: fractured rock aquifer.
- For group A, the isotopic pattern is similar to that of the rainfall. It illustrates that direct infiltration occurs before evaporation.
- For group C, four subgroups are identified, namely monitoring boreholes (C1) and the hot springs (C2). C3 are the boreholes in the Baviaanskloof aquifer. C4 is VG3 in the Skurweberg aquifer and the V-notch water. VG3 is clearly related to water in the V-notch (Figure 4.38). It is worth noting that VG3 is located in the lower part of the Vermaaks notch. Part of the notch water may recharge the Nardouw aquifer again.
- All points of the boreholes in the diagram clearly show an enrichment of heavy isotopes with respect to the South Africa MWL (IAEA, 1981), but only VR7, VR8, VR, VG3, V-notch, M-notch, the Toorwater and Calitzdorp hot springs possess the characteristics with respect to the Global MWL (Dansgaard, 1964). All rainfall points show an enrichment of heavy isotopes with the Cape MWL (Diamond, 1997). This is due to the

amount effect and the heavy rainfall event, whereby a greater amount of precipitation will give a more negative $\delta^{18}\text{O}$ signature. The Kammanassie MWL is suggested as equation (4.3). The slope of 7.4, less than 8, in the Kammanassie is in fact a mixing line, joining a pole of Mediterranean water vapour to a pole of Atlantic water vapour.

$$D=7.4\delta^{18}\text{O}+9.1 \quad R^2=0.75 \quad (4.3)$$

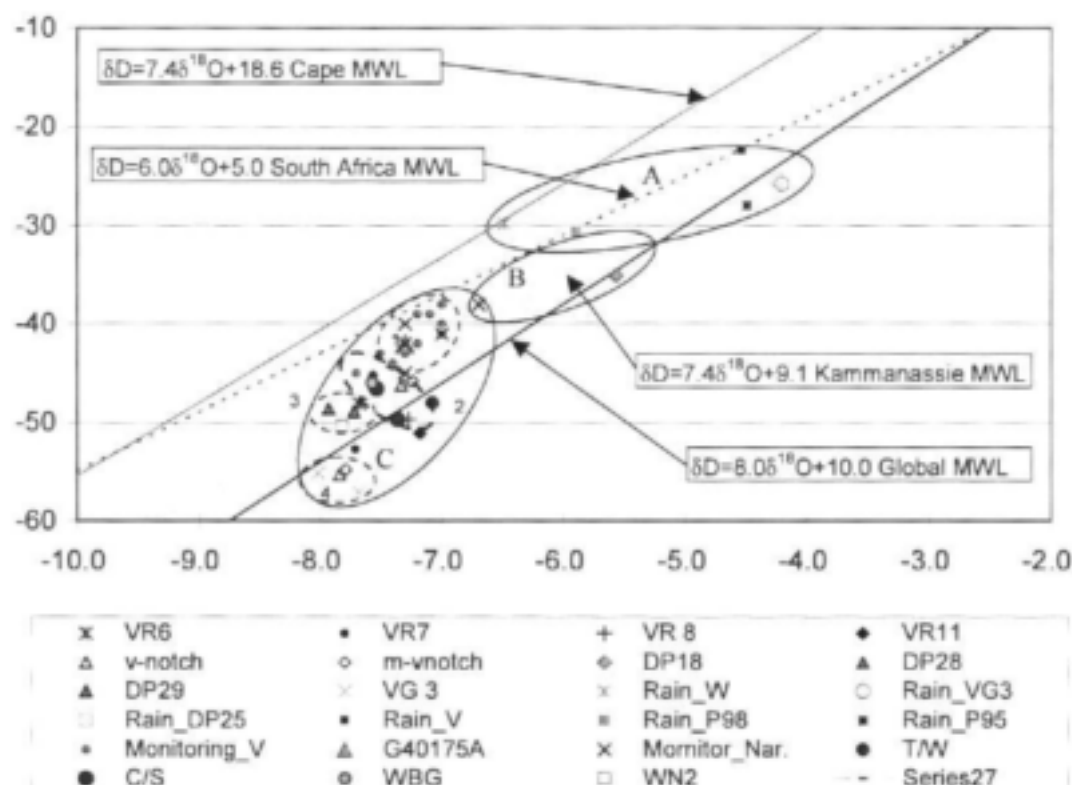


Figure 4.37 $\delta^{18}\text{O}$ and $\delta^2\text{H}$ diagram in the Kammanassie area

Rainfall and mountain seepage (interflow) samples become progressively lighter (more negative) with increasing altitude (lower temperature). This is particularly true for the lighter signatures of rain samples taken at the Wildebeesvlakte and Parshall rain gauges as opposed to the rain samples collected at lower altitudes (Kotze, 2002). The $\delta^{18}\text{O}$ versus altitude diagram seems to identify the sources of water and the precipitation altitude as shown in Figure 4.39. This regional isotopic gradient is approximately 0.44‰. It is higher than the average universal gradient (0.3‰). The equation of the gradient used is

$$\delta^{18}\text{O} = -2.1 - 0.0044\text{Alt.} \quad R^2=0.7191 \quad (4.4)$$

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

The ranges of recharge altitude of the borehole water are according to the equation (4.4) shown in Figure 4.41. The range of altitude of mountainous area is from 900- 1500 m except DP18. It seems to correspond to the mean altitude of the recharge area.

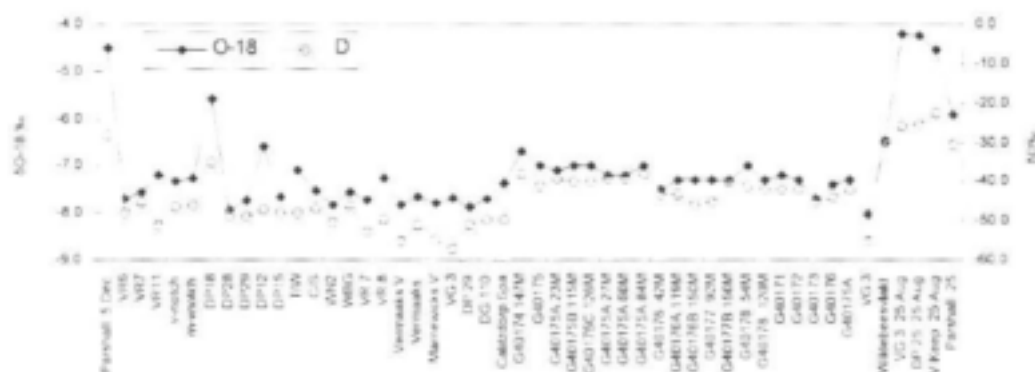


Figure 4.38 Fingerprint of isotopes in the study area from 1995 to 1998

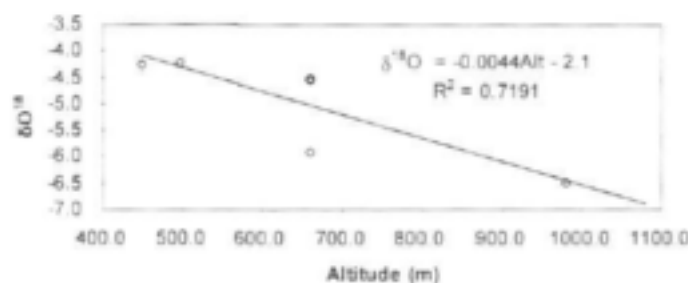


Figure 4.39 Altitude effect of the isotope

Isotopes of ^{18}O are used as indicators of the groundwater component of stream flow during periods of storm runoff and the relation is obtained as (Fritz et al. 1976).

$$Q_g = Q_s \left(\frac{\delta^{18}\text{O}_s - \delta^{18}\text{O}_g}{\delta^{18}\text{O}_s - \delta^{18}\text{O}_{\text{atm}}} \right) \quad (4.5)$$

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

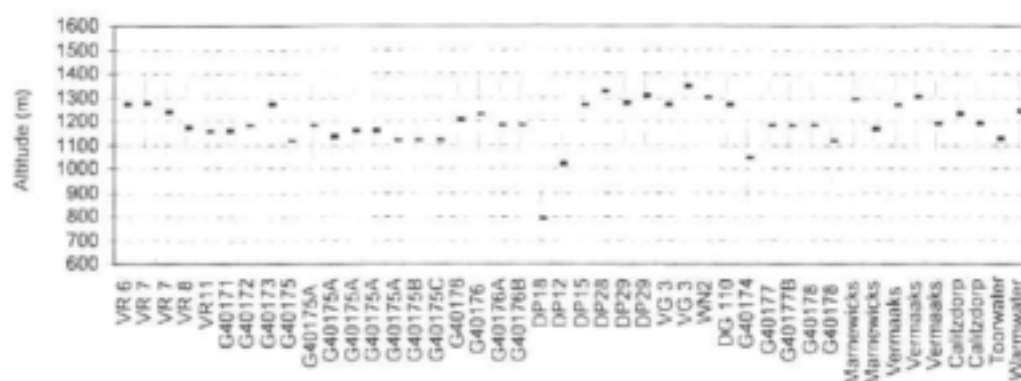


Figure 4.40 Recharge altitude of groundwater

where $\delta^{18}\text{O}$ denotes the ^{18}O content per mille relative to the SMOW standard and the subscripts w , g , and Run indicate stream water, shallow groundwater, and runoff water derived from the rainfall ($Q_w = Q_g + Q_{Run}$). This relation provides for separation of the rain-derived component from the component of the streamflow represented by water that was in storage in the groundwater zone prior to the rainfall event. From the mass-balance point of view, the spring flow (Q_s) recorded at the V-notch consists of the upstream groundwater (Q_g) of Peninsula Formation and the perched water flow (Q_p) in weathered zone as seen in Figure 4.24, i.e.

$$Q_s = Q_g + Q_p$$

where the term Q_p is apparently influenced by rainfall infiltration. The equation (4.5) changes to

$$\frac{Q_s}{Q_g} = \left(\frac{\delta^{18}\text{O}_s - \delta^{18}\text{O}_g}{\delta^{18}\text{O}_s - \delta^{18}\text{O}_p} \right) \quad (4.6)$$

where, average $\delta^{18}\text{O}$ of rain, spring, production borehole and monitoring borehole are -4.86‰ , -7.45‰ , $\delta^{18}\text{O} -7.33\text{‰}$ and -7.25‰ , respectively (refer to original isotope data presented in Appendix V). About 8.37% and 4.86% of spring flow respectively comes from the monitoring borehole (shallow aquifer) and the production borehole (Peninsula aquifer). It confirms the conceptual model shown in Figure 4.24.

4.3.3.2 Tritium

Tritium is the heavy isotope of hydrogen. Tritium atoms are unstable and disintegrate radioactively, forming stable Helium atoms. The half-life of the tritium is 12.3 years. The concentration of tritium in water is expressed by the ratio of tritium (T) atoms to hydrogen (H) atoms:

A ratio of $T/H=10^{-18}$ is defined as 1 Tritium unit (1TU)

Tritium concentrations in groundwater may be used as indicators of groundwater age based on a comparison with historical records of elevated tritium levels in precipitation (maximum 3000TU in the northern hemisphere and 3 – 4 TU in the southern hemisphere), attributed to atmospheric thermonuclear testing between the 1950s and the late 1996 (Schlosser *et al.*, 1988; Busenberg and Plummer, 1993; Dindane *et al.*, 2003). The tritium level in young groundwater is about the same level as in precipitation, but ^3H is significantly diluted by water content of the unsaturated zone. Since the mid-eighties the tritium values in rain water are 10 TU in the northern hemisphere (Mook, 2001). The semi-quantitative dating is possible and very informative (Mazor, 1991):

- Water with zero (in practice <0.5TU) has a pre-1952 age.
- Water with significant tritium concentrations (in practice >10TU) is of a post-1952 age.
- Water with little, but measurable, tritium (between 0.5 and 10TU) seems to be a mixture of pre-1952 and post-1952 water.

The variation of the tritium contents (Appendix V) seems to show two types of water: (i) deep circulation old water including all fractured rock aquifers that is not influenced by the modern recharge (lower than 1.0 TU). (ii) Mixing water near the surface including alluvial aquifers, which have modern water contributions (3.2TU).

The resident time of groundwater samples in the summer season seems to have a short to long sequence as follows (Figure 4.41): DP18 (3.2TU) to the V-notch in the Vermaaks and Manewicks river (1.6TU) to VG3 (1.3TU) to VR7 (1.0TU) to VR6, VR8, DP28 (0.4TU) to VR11, DP12, WN2 (0.3TU) to DP29, DG110 (0.1TU). For winter samples, the tritium (0.5TU) in the V-notch of the Vermaaks and Manewicks rivers corresponds to VR6 and VR7.

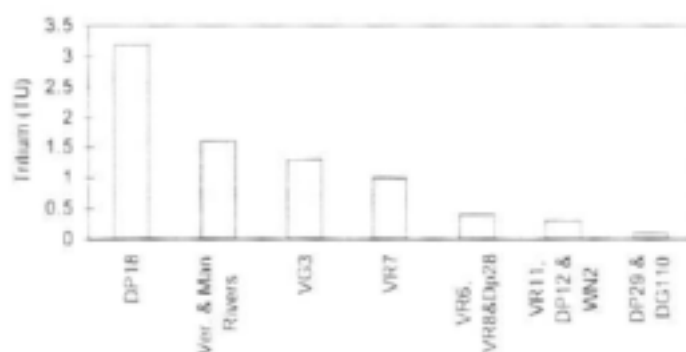


Figure 4.41 Fingerprint of tritium

It seems to demonstrate that the water in the V-notch is partially fed both by local flow and by rainfall in summer season, but totally fed by groundwater in winter season. The sequence

seems to illustrate the circulation is deeper or slower in VR6, VR8 and VR11 than that in VR7 and VG3, or preferential recharge paths occur in VR7 and VG3.

4.3.3.3 Carbon-13 and Carbon-14

Stable carbon isotope ^{13}C of water samples were analysed to aid in determining the flow paths of water entering the boreholes and to distinguish between source waters. Based on the piston flow model, new water should have a more negative ^{13}C signature relative to the 'old' water in the system, since new water is heavily influenced by contact with soil zone CO_2 . The ^{13}C in water of boreholes and hot springs varies from -18.8 to -21.3‰ PDB and -14.5 to -17.3‰ PDB, respectively. These seem to indicate more modern water components in boreholes than those in the hot springs. For ^{13}C in the alluvial waters it's as heavy as, or heavier than, ^{13}C measured in the hot spring and in the TMG aquifer. This borehole water, in theory, is relatively old.

The pattern of distribution of ^{13}C and ^{14}C tends to be opposite to each other except for sample DP18 (Figure 4.43). The ^{14}C values for the boreholes in the Peninsula aquifer are from 74.6 to 83pmc. Boreholes in the Nardouw aquifer give a value of 41.2 - 62.3pmc, which is much lower than those in the Peninsula aquifer. The carbon isotopes in the Calitzdorp, Toorwater and Warmwaterberg hot springs are completely different from those of boreholes. The lower ^{14}C and higher ^{13}C demonstrate that hot spring water is relatively older than that in the boreholes. The results further suggest that the groundwater flow in boreholes in the Peninsula aquifer is faster than that in the Nardouw aquifer and the hot springs.

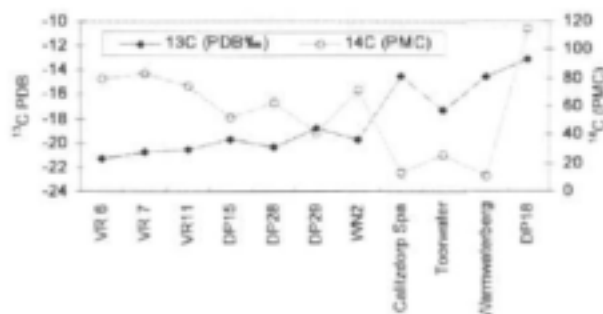


Figure 4.42 Diagram of carbon isotopes in the study area

Different tracers may also give different ages where mixing of very young and very old groundwater occur (Cook and Solomon, 1997). In comparison to DP18, the relevant residence time from carbon-14 and tritium shows large differences (Table 4.13).

The hot spring water is thought to be old water and its ^{14}C ages are from 12364.2 to 19348.5 years. For the borehole water, the residence time of ^{14}C is from 2628.6 to 8968.7 years; but the residence time of tritium is from 12.3 to 61.5 years. For cold springs in the Notches, the tritium gives ages of 10.2 - 36.9 years. It is important to note that the residence time estimated here would be higher than actual ages because A_0 of carbon-14 must be larger than that in DP18. The measured $^3\text{H}/^3\text{He}$ age will reflect the residence time of the younger water component, because the $^3\text{H}/^3\text{He}$ ratio is unaffected by dilution with old (greater than 50 years) water

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(Kamensky *et al.*, 1991). CFC ages of 30 - 45 years (Weaver and Talma, 1999) may also reflect the younger water component. Therefore, the abstractions are likely to be the mixture of the old water (old storage) and newly recharged water. Based on data of the temperature, tritium, CFC age and ^{14}C (Weaver and Talma, 1999), the VR6, VR7 and VR8 in Peninsula aquifer seem to form a group, and VG3, DP15, DP25 and DP28 in the Nardouw aquifer another group.

Table 4.13 Dating results from carbon-14 and tritium (Weaver and Talma, 1999)

Site	DATE	^{14}C (pmc)	T (TU)	Residence time in comparison to DP18 (years)			
				^{14}C		^3H	
				Minimum	Maximum	Minimum	Maximum
DP18	1996-1-31	114.2	3.2	0.0		0.0	
VR6	1996-1-31	79.4	0.4	3005.2		36.9	
VR7	1997-7-1	80.6 $\pm$ 2.5	1.0 $\pm$ 0.2	2628.6	3141.7	17.4	24.6
VR7	1996-1-31	83	0	2638.5		12.3	
VR8	1997-7-1	79.3 $\pm$ 2.5	0.4 $\pm$ 0.1	2758.9	3280.5	32.9	42.0
VR11	1996-1-31	74.6	0.3	3520.8		42.0	
DP12	1996-1-31		0.3			42.0	
DP15	1996-1-31	52		6504.8			
DP28	1996-1-31	62.3	0.4	5010.6		36.9	
DP29	1996-1-31	41.2	0.1	8429.7		61.5	
DP29	1997-7-1	40.6 $\pm$ 2.0	0.0 $\pm$ 0.2	8153.4	8968.7	49.2	
VG3	1997-7-1	78.1 $\pm$ 2.5	1.3 $\pm$ 0.2	2881.1	3410.7	13.4	19.0
VG3	1998-6-1	65.7 $\pm$ 2.3		4286.7	4865.8		
WN2	1996-1-31	72	0.3	3814.1		42.0	
DG 110	1997-7-1	41.0 $\pm$ 2.0	0.0 $\pm$ 0.2	8076.2	8883.5	49.2	
Marnewicks Notch	1997-7-1		1.6 $\pm$ 0.2			10.2	14.7
Marnewicks Notch	1996-1-31		0.5			32.9	
Vermaaks (Partial)	1997-7-1		0.5 $\pm$ 0.1			29.7	36.9
Vermaaks Notch	1997-7-1		1.6 $\pm$ 0.2			10.2	14.7
Vermaaks Notch	1996-1-31		0.5			32.9	
Calitzdorp hot spring	1996-1-31	13		17967.2			
Toorwater hot spring	1996-1-31	25.6		12364.2			
Warmwaterberg hot spring	1996-1-31	11		19348.5			

According to piston model, flow velocity of the water for production boreholes are estimated based on the distance between the borehole and the watershed and the maximum eastern lithological boundary (Table 4.14). Large ranges of flow velocity and recharge rate occur in the Nardouw aquifer because the recharge can occur in the whole outcrops of the Nardouw Subgroups. If the recharge area is close to the borehole, the recharge should be less than these values. From the watershed point of view, the recharge rate in the Peninsula Formation is larger than that in the Nardouw Subgroup; but the water resident time of VR11 is older than those of other boreholes of the Peninsula aquifer, although VR11 is closer to the watershed than the other boreholes. Perhaps the flow paths are different in the fracture networks.

4.3.4 Surface water quality

The chemical data of the surface water in the study area recorded monthly in the NGDB are discontinuous. The chemical features of surface water, including Stompdrift, Kammanassie, Calitzdorp, Gamkapoort and Koos Rauben dams in the surrounding area are illustrated in Table 29. The concentrations of constituents, especially major ions Cl^- and Na^+ in the dams are higher than those in the boreholes except for the Koos Raubenheimer dam (refer to Appendix IV). Values of pH are usually greater than 7.0. Low chloride concentrations near the Kammanassie dam were accompanied with extremely high rainfall events from 1985 to 2003 (Figure 4.43). The chloride concentration appeared to be increasing until a new extremely high rainfall event occurred in 1996 to 2001. Perhaps the high chloride concentrations are attributed to high evapotranspiration. The chloride concentrations of the dam water flowing in during flood periods are lower than that stored in the dam resulting in the dilution of chloride concentrations; but the chloride concentrations are still higher than those of the boreholes in the Kammanassie area.

Table 4.14 Flow velocity and recharge based on residence time of ^{14}C

Borehole	^{14}C Ages (years)	Distance (m)		Flow velocity ($\text{m}\cdot\text{a}^{-1}$)		Recharge, if S=1‰ ($\text{mm}\cdot\text{a}^{-1}$)	
		To watershed	To eastern boundary of Nardouw/Peninsula	Min	Max	Min	Max
WN2	3,814.1	6,100	44,900	1.60	11.77	1.60	11.77
VG3	4,865.8	8,150	46,950	1.67	9.65	1.67	9.65
DG110	8,883.5	12,000	50,800	1.35	5.72	1.35	5.72
DP29	8,429.7	14,750	53,550	1.75	6.35	1.75	14.04
VR6	3,005.2	7,500	15,500	2.50	5.16	2.50	5.16
VR7	2,628.6	7,250	15,250	2.76	5.80	2.76	5.80
VR7	2,638.5	7,250	15,000	2.75	5.68	2.75	5.68
VR8	2,758.9	7,000	14,750	2.54	5.35	2.54	5.35
VR11	3,520.8	6,750	14,500	1.92	4.12	1.92	4.12

Table 4.15 Chemical characteristics of surface water in the surrounding areas of the Kammanassie in 1992-1998 ($\text{mg}\cdot\text{l}^{-1}$)

Station	EC $\text{mS}\cdot\text{m}^{-1}$	TDS	pH	Na	K	Mg	Ca	F	Cl	NO_3^- as N	SO_4	P	Si	NH_4^+ as N	Alk
Stompdrift Dam	96.02	604.39	8.3	112.8	6.0	21.9	44.0	0.4	164.8	0.3	82.5	0.0	1.7	0.1	138.4
Kammanassie Dam	43.12	264.51	7.8	50.8	7.1	10.3	15.3	0.2	80.2	0.3	25.6	0.1	2.3	0.6	63.5
Calitzdorp Dam	26.16	182.05	8.0	16.4	6.2	7.9	22.9	1.0	15.4	0.1	16.9	1.1	2.2	0.1	99.1
Gamkapoort Dam	62.79	422.09	8.2	65.5	7.3	12.5	38.7	0.3	78.8	0.5	55.9	0.1	4.0	0.1	131.5
Koos Rauben Dam	7.20	49.18	7.3	6.1	0.5	1.9	4.1	0.2	8.5	0.1	5.8	0.0	1.9	0.0	17.9

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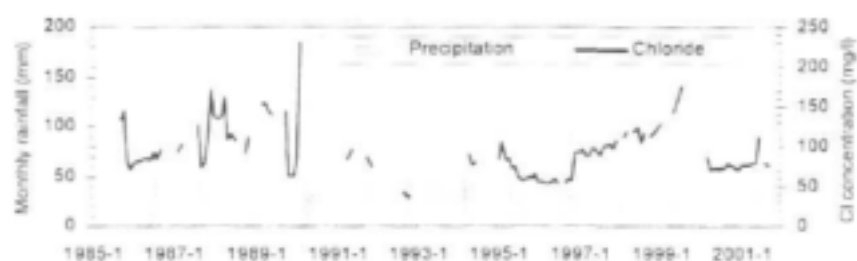


Figure 4.43 Cl concentration versus rainfall in the Kammanassie Dam

5. RECHARGE ESTIMATION IN THE KAMMANASSIE AREA

In the last 10 years, several recharge studies of the TMG aquifer using CMB, SVF, CRD and EARTH methods and hydrograph separation technique were carried out for the Kammanassie area. The recharge rates estimated using these methods were very different for the same area. The estimated recharge rate was as high as 17% of mean annual precipitation (Bredenkamp, 1995). The recommended abstractions were adjusted many times in an attempt to stop the continued decline of groundwater levels in the Vermaak's River wellfield since 1984. In order to estimate realistic recharge rates in the Kammanassie area, conceptual recharge models are developed. Several new methods were used to cross validate the results of the other methods.

5.1 *Recharge Estimation with Mixing Model of CMB*

From a water balance point of view, groundwater abstraction from an aquifer can be considered a mixture of recent recharge and old storage. A mixing model of CMB assumes that extent of the mixing can be indicated by a ratio of the recent recharge to the old storage. The recent recharge rates are obtained through the recharge volume of the annual average precipitation over the study area.

A conceptual model for the abstraction from recharge and aquifer storage is shown in Figure 5.1. The following information is illustrated:

- Direct recharge, which is recharge through preferential fracture networks without additional chloride input. The chloride concentration of the water thus remains the same as that of the rainfall infiltration through the fracture.
- Indirect recharge through the unsaturated zone dissolves additional chloride. From the dissolution, the chloride concentration of the recharge water will increase during the infiltration process.
- Interflow, which occurs along the shallow weathered zone and discharges to local drainage systems within a short time. The chloride concentration of the indirect recharge is similar to that of the interflow.
- Local flow (stagnate flow or old aquifer storage) may contribute a portion of discharges into wells.

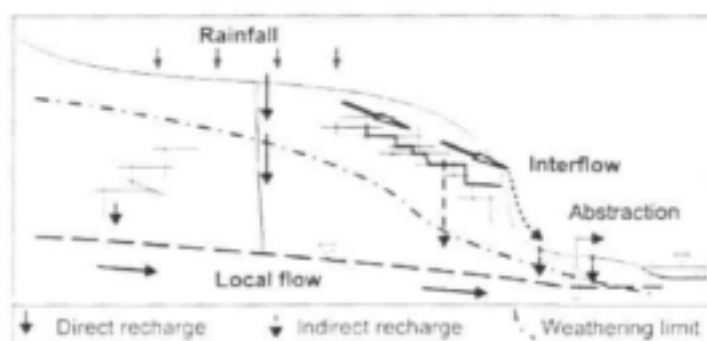


Figure 5.1 Conceptual mixing model (Wu and Xu, 2004)

5.1.1 Mixing model

A proposed two-component mixing model assumes that the abstraction consists of old aquifer storage and recent recharge. In the three-component mixing model, the rainfall water component is further divided into direct and indirect recharge components. The three-component mass balance equations were obtained by sequentially expanding the two-component mass balance equations.

5.1.1.1 Two-component mixing model

Based on the conceptual model in Figure 5.1, the two-component chloride mass balance equations are written as:

$$Q_p = Q_r + Q_s \quad (5.1)$$

$$Q_p C_p = Q_r C_r + Q_s C_s \quad (5.2)$$

where Q_p is the volumes of abstraction from production boreholes, Q_r is volumes of recharge, Q_s is volumes of aquifer storage and the chloride concentrations of the abstraction water, recharge water and aquifer storage water represented by C_p , C_r and C_s , respectively. For direct recharge, an amount of recharge water with low chloride concentration would dilute that of storage. In this case, the equation for the recharge can be derived from equation (5.1) and equation (5.2) as follows:

$$Q_r = Q_p \frac{C_s - C_p}{C_r - C_p} \quad \text{where } C_r < C_s \quad (5.3)$$

For indirect recharge, an amount of recharge water with high chloride concentration would increase that of storage. In this case, the equation for the recharge can be derived from equation (5.1) and equation (5.2) as follows:

$$Q_v = Q_r \frac{C_p - C_s}{C_r - C_s} \quad \text{where } C_p > C_s \quad (5.4)$$

where Q_v is the volumes of indirect recharge and C_v is the chloride concentration of indirect recharge.

5.1.1.2 Three-component mixing model

The limitation of the two-component model is that direct and indirect recharge cannot be considered together at the same time; one of them must be ignored. Sequel to this and the need of further recharge estimation, the three-component mass balance equations based on the abstraction of the production boreholes can be written as

$$Q_p = Q_r + Q_v + Q_s \quad (5.5)$$

$$Q_p \times C_p = Q_r \times C_r + Q_v \times C_v + Q_s \times C_s \quad (5.6)$$

where Q and C represent abstraction (L^3T^{-1}) and concentration of chloride (ML^{-3}) in direct recharge (r), indirect recharge (v) and aquifer storage (s), respectively. If equation (5.5) and equation (5.6) are divided by Q_p , the following equations are obtained:

$$1 = Q'_r + Q'_v + Q'_s \quad (5.7)$$

$$C_p = Q'_r C_r + Q'_v C_v + Q'_s C_s \quad (5.8)$$

where $Q'_r = Q_r / Q_p$, $Q'_v = Q_v / Q_p$ and $Q'_s = Q_s / Q_p$ are the relative contributions to the abstraction in the Vermaak's River Well field at any given time from the direct and indirect recharge and aquifer storage, respectively. Rearranging equation (5.7) and equation (5.8), the following equations are derived:

$$Q'_v = \frac{Q'_r (C_r - C_s) + (C_p - C_s)}{C_v - C_s} \quad (5.9)$$

$$Q'_s = 1 - (Q'_r + Q'_v) \quad (5.10)$$

The solution of the set of the above equations can refer to the solution of implicit functions (Geary *et al.*, 1963). For achieving the approach to the solution of the set of the equation (5.9) and equation (5.10), one of the ways is the utilization of the rainfall component (Q_r) defined by equation (5.3) or equation (5.4). From equation (5.9) and equation (5.10), the relative contributions of each water component to the production borehole in the Vermaaks River Wellfield at any instant could be determined, although errors occur in the solutions.

The three-component mixing model over a given time interval i (month) was constructed as follows:

$$\sum_{i=1}^n C_i Q_i = \sum_{i=1}^n [C_{si} Q_{si} + C_{ri} Q_{ri} + C_{ui} Q_{ui}] \quad i = 1, 2, 3 \dots n \quad (5.11)$$

$$\sum_{i=1}^n Q_i = \sum_{i=1}^n [Q_{si} + Q_{ri} + Q_{ui}] \quad i = 1, 2, 3 \dots n \quad (5.12)$$

where, C_i is monthly chloride concentration in pumping borehole, C_{si} , C_{ri} and C_{ui} are concentrations of chloride in aquifer storage, rainfall and unsaturated zone, respectively. Q_i is monthly abstraction. Q_{si} , Q_{ri} and Q_{ui} are fluxes from aquifer storage, direct and indirect recharge, respectively.

5.1.2 Chloride concentration of recharge components

Based on the above models, the chloride concentrations of the abstraction are the result of the mixing of the different components including storage, direct and indirect recharge. Therefore, the chloride concentration of the different components should be identified first.

5.1.2.1 Chloride concentration of the old storage

Based on hydrogeochemical evolution, the initial concentration of the storage water in an aquifer can be represented by the background chloride concentration. The background concentration may change due mixing by recharge water. Therefore, it is very important to determine the background chloride concentration and its evolution in an aquifer. To identify the background value and the evolution, a statistical analysis method is used:

A classical time series of chloride concentration may be written as

$$C = C_{si}(t) \cdot C_{ry}(t) \cdot C_{re}(t) \cdot C_{ra}(t) \quad (5.13)$$

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where C is the chloride concentration, and $C_{tr}(t)$, $C_{cy}(t)$, $C_{se}(t)$ and $C_{ir}(t)$ are the trend, cyclical, seasonal and irregular, respectively; t is the time.

The trend describes the net influence of long-term factors whose effects on the chloride concentration tend to change gradually. Generally, these factors include changes in size as well as geological and hydrogeochemical evolution. This value would represent the chloride background value. The cyclical component describes the net influence of global and regional climatic change on annual rainfall distribution. It illustrates the relationship between the changes of the concentration of chloride with a change of rainfall. The seasonal component describes effects that occur regularly over a period of a year, quarter, month, day or rainfall event. The irregular component describes residual changes that remain after the other components have been taken into account. Irregular changes reflect effects of unique and nonrecurring factors, such as sampling and analyses.

The relationship between the mean annual precipitation and chloride concentration seems to be negative in the Peninsula aquifer (VR6, VR7, VR8 and VR11) but positive in the Nardouw aquifer (VG3) as illustrated in Figure 5.2. The results show that rainfall results in more direct recharge in the Peninsula aquifer, but creates more indirect recharge in the Nardouw aquifer. The chloride concentrations related to the abstractions in the Vermaaks River Well field from 1994 to 2003 are used to perform modelling (Figure 5.3). It is important to note that the data is only available up to March 2003.

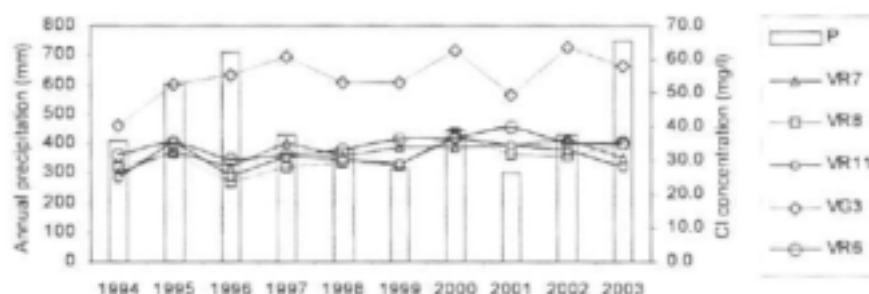


Figure 5.2 Cl concentration versus precipitation in the Vermaaks area

To evaluate the hydrogeochemical evolution trend in the Vermaaks area, a time series analysis was performed. The data of chloride concentration were adopted from a monthly report of the Overberg Board Water Management. The chloride concentration evolution trend from 1994 to 2003 in the Vermaaks River Well field is demonstrated by regression analysis although the correlation coefficient is low (Figure 5.4). The initial background values of chloride varied from 25.1 to 28.9 mg l^{-1} in the Peninsula aquifer and is 48.7 mg l^{-1} in the Nardouw aquifer (Table 5.1). The background values of chloride concentration from 1994 to 2003 are listed in Table 31, which forms the basis of calculations for the mixing model.

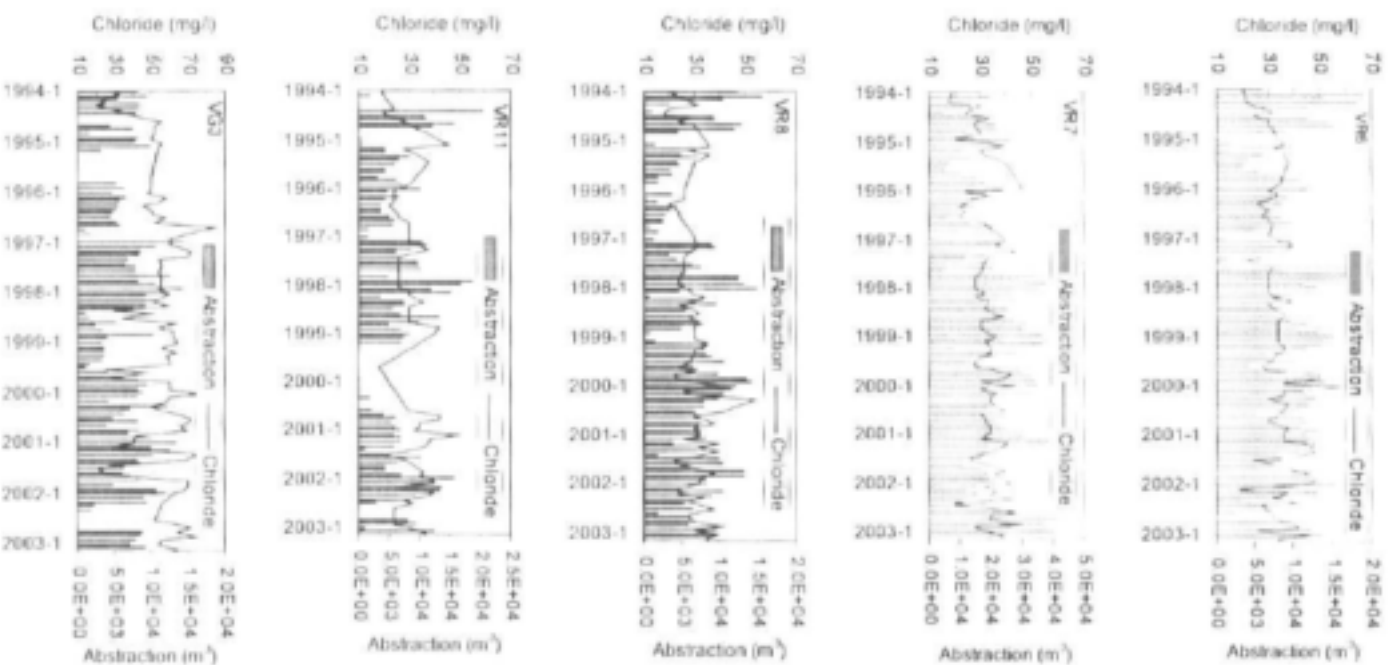


Figure 5.3 Cl conics response to the abstraction in the Vermaaks River Well field

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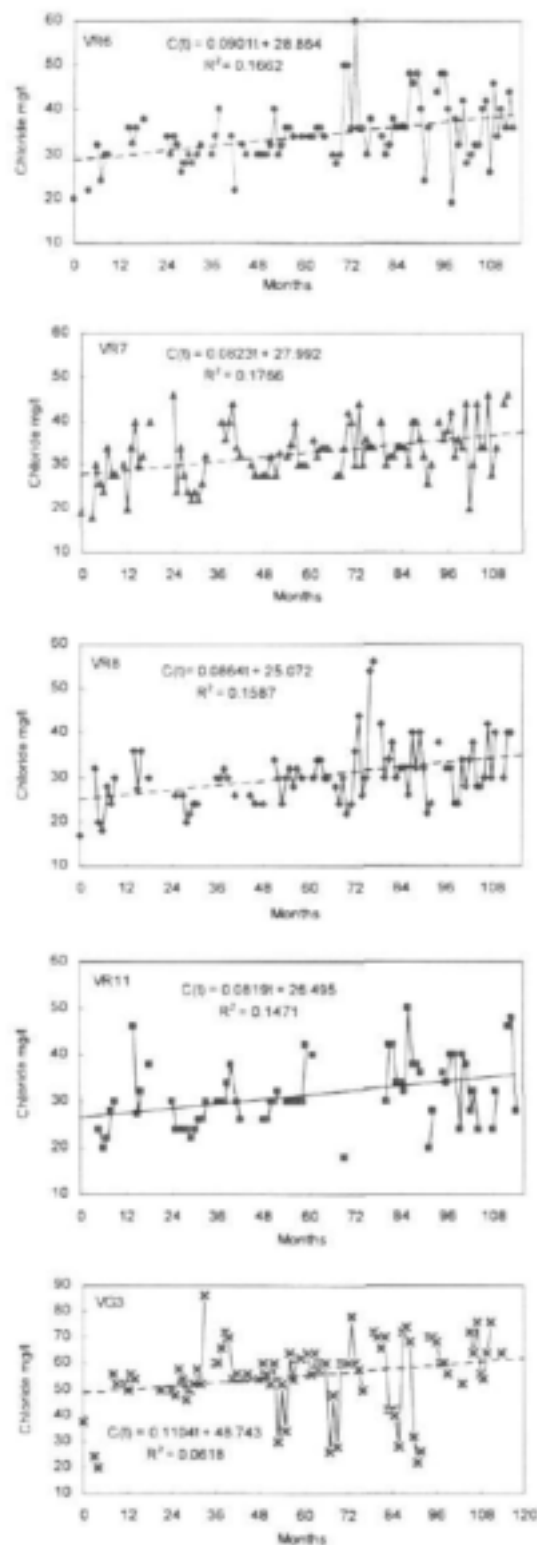


Figure 5.4 Chemograph of boreholes in the Vermaaks River Well field

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Table 5.1 Background values of the Cl conc

Aquifer	Borehole	Regression equation	Coefficient of correlation	Initial background concentration ($\text{mg}\cdot\text{l}^{-1}$)
Peninsula	VR6	$C(t) = 0.0901t + 28.86$	$R^2 = 0.1662$	28.9
	VR7	$C(t) = 0.0823t + 27.99$	$R^2 = 0.1766$	28.0
	VR8	$C(t) = 0.0864t + 25.07$	$R^2 = 0.1587$	25.1
	VR11	$C(t) = 0.0819t + 26.50$	$R^2 = 0.1471$	26.5
Nardouw	VG3	$C(t) = 0.1104t + 48.74$	$R^2 = 0.0618$	48.7

Table 5.2 Basis of background values of the Cl conc ($\text{mg}\cdot\text{l}^{-1}$)

Borehole	VR6	VR7	VR8	VR11	VG3
1994	28.9	28.0	28.2	24.8	48.7
1995	29.0	28.1	32.4	35.9	48.8
1996	29.0	28.2	23.7	25.4	49.0
1997	29.1	28.2	28.3	31.1	49.1
1998	29.2	28.3	29.3	30.6	49.2
1999	29.3	28.4	28.6	29.0	49.3
2000	29.4	28.5	38.2	37.0	49.4
2001	29.5	28.6	31.8	34.7	49.5
2002	29.9	28.7	31.2	33.3	49.6
2003	29.7	28.7	35.0	28.0	49.7

5.1.2.2 Chloride concentration of the other components

The chloride concentration of direct recharge is obtained from the rainfall stations in Wildebeesvlakte as well as three new monitoring rainfall stations in the Purification Works at Dysselsdorp. The chloride concentrations of the direct recharge (rain water) range from $1.33\text{ mg}\cdot\text{l}^{-1}$ to $1.74\text{ mg}\cdot\text{l}^{-1}$. The concentrations of indirect recharge (interflow) measured range between 74.1 and $83.7\text{ mg}\cdot\text{l}^{-1}$ as captured after a rainfall event on May 13, 2003.

5.1.3 Scenarios for modelling

In order to identify the relative contributions of different components to the abstraction, five scenarios are considered. The relative contributions of different components in scenario 1 are computed based on the results of the analysis. Scenarios 2 to 5 are designed according

to the distribution pattern of the chloride concentration and possibility of the extreme event to evaluate the changes of the contributions of the different components with the variation of the chloride concentration input.

Scenario 1 (S1): Based on the information of the results of the analysis: chloride concentration of rainfall (C_r): $1.74 \text{ mg}\cdot\text{L}^{-1}$, chloride concentration of indirect recharge (C_v): $74.1 \text{ mg}\cdot\text{L}^{-1}$ and chloride background concentration of storage water in aquifer (C_a): $26.5 \text{ mg}\cdot\text{L}^{-1}$.

Scenario 2 (S2): C_r : $1.74 \text{ mg}\cdot\text{L}^{-1}$, C_v : $74.1 \text{ mg}\cdot\text{L}^{-1}$ and initial background concentration in considering time series (C_{so}).

Scenario 3 (S3): C_r : $1.0 \text{ mg}\cdot\text{L}^{-1}$, C_v : $60.0 \text{ mg}\cdot\text{L}^{-1}$ and chloride background concentration.

Scenario 4 (S4): C_r : $10.0 \text{ mg}\cdot\text{L}^{-1}$, C_v : $60.0 \text{ mg}\cdot\text{L}^{-1}$ and chloride background concentration.

Scenario 5 (S5): C_r : $1.00 \text{ mg}\cdot\text{L}^{-1}$, C_v : $85.0 \text{ mg}\cdot\text{L}^{-1}$ and chloride background concentration.

5.1.4 Results

5.1.4.1 Recharge volume from two-component model

The proportions of components based on the two-component model are presented in Appendix VI (a, b). The following patterns are recognised:

- The ratio of recharge to the abstractions in VR6, VR7, VR8 and VR11 are 0.009 - 0.685, 0.0002 - 0.388, 0.002 - 0.586, 0.010 - 0.488, respectively.
- The ranges of direct recharge ratio to the abstraction in VR6, VR7, VR8 and VR11 are 0.038 - 0.380, 0.006 - 0.381, 0.002 - 0.346, 0.010 - 0.354, respectively.
- The ranges of indirect recharge ratio to the abstraction in VR6, VR7, VR8 and VR11 are 0.009 - 0.685, 0.0002 - 0.388, 0.007 - 0.586, 0.018 - 0.488, respectively.
- The direct and indirect recharge ratios to the abstraction in VG3 vary from 0.020 to 0.612 and from 0.024 to 1.473, respectively. A ratio value larger than 1.0 indicates that recharge exceeds abstraction.

The frequency of recharge events from 1994 to 2003 is listed in Table 5.3. The frequency of the direct and indirect recharge events in Peninsula and Nardouw aquifers was 14.3% - 26.4% and 73.6% - 85.7%, respectively.

The results of recharge volumes of the scenarios are presented in Appendix VII. The following were observed:

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- The volumes of direct recharge in VR6, VR7, VR8 and VR11 and VG3 from 1994 to 2003 were up to $4874 \text{ m}^3 \cdot \text{a}^{-1}$, $11220 \text{ m}^3 \cdot \text{a}^{-1}$, $10963 \text{ m}^3 \cdot \text{a}^{-1}$, $3840 \text{ m}^3 \cdot \text{a}^{-1}$ and $19345 \text{ m}^3 \cdot \text{a}^{-1}$, respectively.
- The volumes of indirect recharge were constantly larger than that from direct recharge in different scenarios.
- Direct recharge events are often associated with continuous high rainfall events, such as in September, October and November of 2001; the direct recharge occurred over a long period although annual precipitation was small.
- The largest recharge volume of 255413 m^3 associated with abstraction occurred in VR7 while the least recharge volume of 65088 m^3 occurred in VR11.

Table 5.3 Frequency of the direct and indirect recharge (1994 to 2003)

Aquifer	Peninsula								Nardouw	
	VR6		VR7		VR8		VR11		VG3	
Frequency	Direct	Indirect	Direct	Indirect	Direct	Indirect	Direct	Indirect	Direct	Indirect
Count	12	72	21	69	18	64	19	53	17	70
%	14.3	85.7	23.3	76.7	22.0	78.0	26.4	73.6	19.5	80.5

The relationship between recharge and precipitation is shown in Figure 5.5. The recharge volumes seemed to increase from 1996 to 2002. It may imply that the impact of the flood in 1996 which lasted for 6 years resulted in favourable conditions for recharge in the production boreholes.

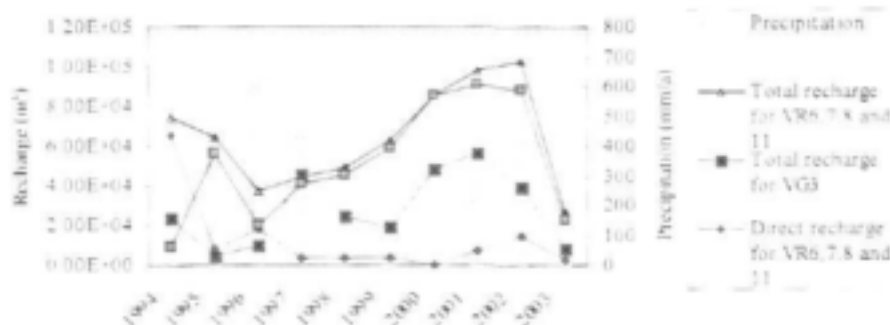


Figure 5.5 Recharge volumes versus precipitation

5.1.4.2 Recharge volume from three-component model

Based on chloride concentrations, volumes of the direct and indirect recharge and aquifer storage water were obtained (presented in Appendix VIII). The results of scenario 2 are not included in the following because the background chloride value is the same as the initial background value. The following observations were made:

1. The proportions of total recharge volume ($Q_r + Q_v$) of VR6, VR7, VR8 and VR11 in scenario 1 were 0.066 - 0.262, 0.073 - 0.204, 0.058 - 0.278 and 0.067 - 0.259, respectively.
2. The minimum and maximum recharge proportions of the boreholes occurred in different years. These demonstrate that there was a different recharge path or recharge area for the boreholes.
3. The minimum proportions related to the maximum precipitation occurred in 1996, but the maximum proportions were related to minimum precipitation for VR6 in 2001.
4. The direct recharge was always small. The proportions of the direct recharge were smaller than those of the indirect recharge except in 1994.
5. The results from different scenarios were close to those from field measurements

The variation of total recharge proportions ($Q_r + Q_v$) with time is shown in Figure 5.6. The high recharge proportions occurred during rainfall events with high initial chloride concentrations and indirect recharge with low concentrations, such as scenario 3. The lowest recharge proportions occurred in the lowest chloride concentration of direct recharge and in the highest chloride concentration of indirect recharge (scenario 4). The highest recharge volume occurs in the scenario 4, and lowest one is scenario 5 (Figure 5.7 and Appendix IX). The different patterns of the proportions and volumes demonstrate that different recharge processes occur in the region. In other words, the fracture networks may not be hydraulically connected in the Vermaak's River Well field.

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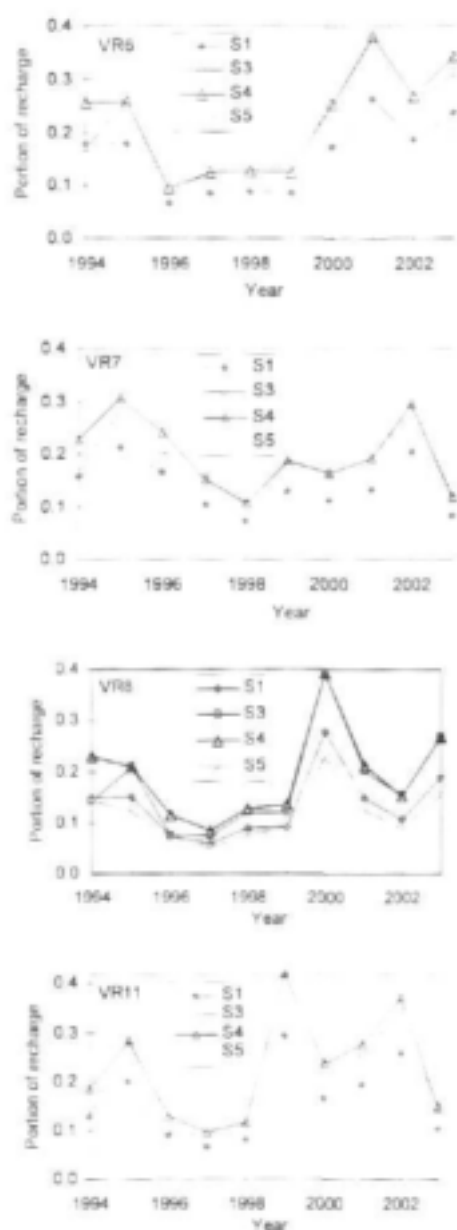


Figure 5.6 Total recharge ratios of the different scenarios in the Vermaak's River Well field

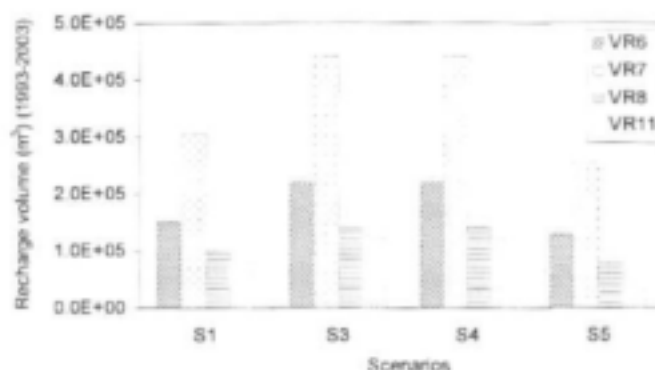


Figure 5.7 Recharge volumes of the different scenarios

5.1.5 Recharge volume and recharge rate

To convert a recharge volume to a recharge rate, the recharge area must be known. In other words, recharge volumes are recharge-area dependent. Variations of the contributing areas directly caused by the pumping regime result in a number of recharge rates although the recharge volume is supposed to be unique. In this study, the recharge rates are calculated based on the contributing areas identified from the hydrogeological setting and boundaries.

For the four scenarios in the two-component model, both the upper catchment and the outcrop window of the Peninsula Formation (see Figure 4.15) are considered over two separate periods. The recharge rates based on the two-component model are listed in Appendix X in Wu (2005). The salient features of the estimated recharge rates are summarized as follows:

1. The recharge rate on an annual and cumulative annual basis, ranged between 1.54% to 2.09% and 1.41% to 2.04% respectively of the rainfall based on the upper catchment area. The values based on the Peninsula Window area ranged from 0.41% to 0.60% and 0.38% to 0.55% respectively of the rainfall.
2. The recharge rate is affected by the concentration of rainfall input. The higher the chloride concentration of rainfall, the higher the recharge rate.
3. The indirect recharge rates seem to be associated with chloride deposition and dissolution in the unsaturated zone.
4. The recharge rates become smaller if chloride concentrations of old storage water increase.
5. The annual average recharge rates are slightly higher than those obtained from the cumulative recharge method.

For scenario 1, the recharge rates against the catchment area in VR6, VR7, VR8 and VR11 were 0.332%, 0.666%, 0.216% and 0.187% of the precipitation, respectively (presented in Appendix XI (a) (Wu, 2005)). The highest recharge rate was obtained from VR7 and the smallest recharge rate occurred in VR11. From the calculated recharge rates, the infiltration condition in the recharge area of VR7 could be said to be the most favourable.

The results of the recharge rates based on the three-component mixing model are presented in Appendix XI (b) (Wu, 2005). The observations are as follows:

- The highest recharge rate was related to the catchment area, and the lowest one was related to the area of equivalence of *Theis's* effective radius (see Xu et al. 2002). The rates against the catchment and the equivalent area of *Theis's* effective radius were 1.182% - 2.027% and 0.006% - 0.011% of the precipitation respectively. The recharge rates would increase significantly if the size of the recharge area was reduced, for example, in the form of an elongated belt zone.
- The recharge rates from the two methods relate to each other in the same scenario. Recharge rates from the three-component model were persistently smaller than that of the two-component model. This would be attributed to the critical values of the two-component model representing the minimum recharge scenario.
- The differences of recharge rates between the different scenarios with different initial chloride concentrations appear to be small.

5.1.6 Influence factor of recharge rate

For recharge rates based on the mixing model, the influence factors include the chloride concentration, the time lag and the recharge area. If the time lag were ignored, it would reduce the accuracy of the estimated recharge rate. The uncertainty of the recharge area would lead to unreliable recharge estimates.

The chloride concentration data was collected from the monthly report of the Overberg Board Water Management. Some missing data were estimated by an interpolation method. The relationship between the chemograph of chloride and rainfall in the Peninsula aquifer is shown in Figure 5.8. The figure indicates that the chloride concentrations of the boreholes vary with the variation of rainfall. Chloride concentration seemed to rise after 3 - 9 month lag accompanying a high rainfall event. The longest time lag appeared to be about 39 - 40 months after the flood in November of 1996. The high chloride concentration after a high rainfall event may be a sign of indirect recharge from rainfall. About 23 higher chloride concentrations seemed to be related to rainfall with precipitation more than 50 mm per month at the Parshall rainfall station, which mainly contributed to the groundwater recharge.

The relationship between the chemograph of chloride in the Nardouw aquifer and the rainfall at the Parshall rainfall station is shown in Figure 5.9. Chloride concentration rises from a high

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rainfall event with no time lag. The low chloride concentration with a time lag of 20-34 months seems to be associated with a high rainfall event. These should be attributed to the direct recharge via fracture networks originating in mountainous areas.

However, the time lag observed in the Figure 5.8 and Figure 5.9 cannot be accepted without scientific explanation, hence auto-regression and cross-correlation had to be done for the determination of time lags. An auto-regression and F-Test was performed to confirm the time lag.

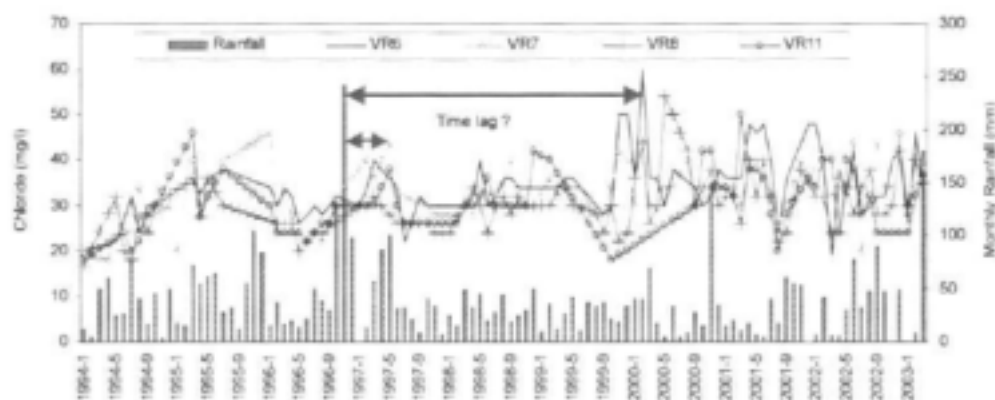


Figure 5.8 Cl concs versus rainfall in the Peninsula aquifer

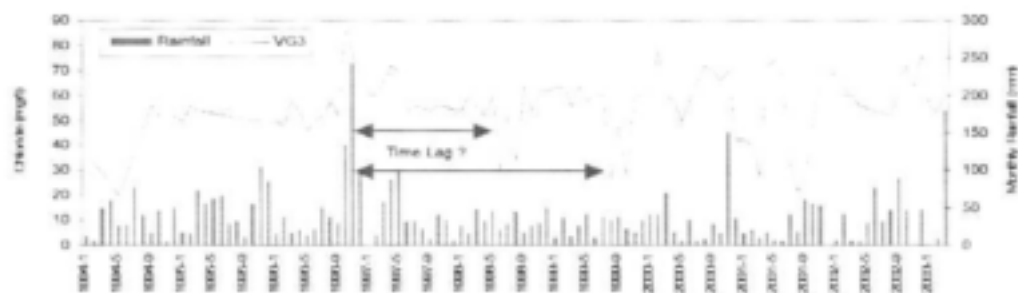


Figure 5.9 Cl concs versus rainfall in the Nardouw aquifer

The quickest response to the rainfall is approximately 1-4 months; the longest one is about 54 months (Table 5.4). From a continuously high one-tailed probability, the chloride concentration has a complex response to the rainfall. There are similar response patterns in the Peninsula aquifer, but largely different from the Nardouw aquifer.

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Table 5.4 Time lag and its probability

Peninsular Aquifer													Nardouw Aquifer			
Borehole	VR6			VR7			VR8		VR11				VG3			
Time lag (month)	1	20	37	2	11	25	2	33	2	4	15	32	2	5	34	54
Probability	0.84	0.97	0.56	0.85	0.61	0.45	0.9	0.98	0.94	0.77	0.79	0.98	0.96	0.94	0.75	0.87

Although the auto-regression can work out the time lag, the cross-correlation is not significant. It is therefore difficult to construct a one-to-one relationship between the rainfall and the chloride concentration under these conditions, perhaps the time series is not long enough. For the effects of time lag on recharge estimation, the cumulative rainfall from 1994 to 2003 is considered instead of monthly rainfall at present.

5.2 Recharge Estimation with Water Level Fluctuation

The type of the groundwater regime, including climatic rainfall-infiltration, evapotranspiration, abstraction, runoff, hydraulic, irrigation, frozen and leakage types, can be used to analyse the cause of a particular groundwater level fluctuation (Fang *et al.*, 1996, refer to Appendix XIII in Wu, 2005), but some of them, such as the climatic rainfall-infiltration type, can be used to estimate recharge because of the cause effect relationship between rainfall and groundwater level rise. Thus, identification of the type of groundwater regime is important if one wants to estimate recharge by groundwater fluctuation method.

5.2.1 Model of cumulative water level departure

To keep the dimension of Cumulative Water level Departure (CWD) consistent with the Cumulative Rainfall Departure (CRD) in respect of data analysis, the CWD is defined as follows:

A time series of water level fluctuation [L]: $X_1, X_2, \dots, X_n$

The average value [L]: $\bar{X} = \frac{\sum_{i=1}^n X_i}{n}$

The departure [L]: $D_i = X_i - \bar{X}$

The cumulative water level departure

[L]: $CWD = \sum_{i=1}^n D_i$

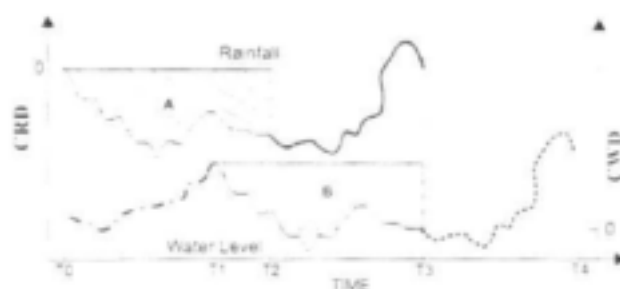


Figure 5.10 Model of CRD-CWD method

The following information can be seen in Figure 5.10:

- The statistical period is from T0 to T3.
- There is a resemblance between A (CRD) and B (CWD). The period of A is from T0 to T2 and B is from T1 to T3. Lag time is assumed as T1-T0.
- The dash curve extrapolated from T3 to T4 period might be related to the black bold solid line from T2 to T3 period.

A simple water balance equation based on the water level response is written as:

$$CWD \times S \times B + Q = rCRD \times A \quad (5.14)$$

where CWD is the cumulative water-level departure [L]; CRD is the cumulative rainfall departure [L]; S is the aquifer storativity; A is the catchment area [L²]; B is the groundwater catchment area [L²]; r is the recharge rate. Q is the abstraction [L³] is defined as:

$$Q = \Delta h \times S \times A_R, \text{ with } A_R = \pi R^2$$

Δh is the average drawdown related to effective radius; R is the effective radius, $R=f(t)$, t is pumping time.

$$\text{Let } CWD \times S \times B + Q = y \text{ and } CRD \times A = x$$

then, the equation (5.14) changes to

$$y = rx \quad (5.15)$$

This is similar to the linear equation $y=kx+c_0$ with $c_0=0$. Therefore, r is the slope of the CWD-CRD. Equation (5.15) is rewritten as

$$r = \frac{CWD \times S \times B + Q}{CRD \times A} \quad (5.16)$$

Suppose A=B and Q=0 in a monitoring borehole, thus

$$r = \frac{CWD \times S}{CRD} \quad (5.17)$$

The resemblance can be determined by correlation analysis. If the plotting series is long enough, the maximum and minimum points in CWD and CRD curves could be found. We call these points peak points (similar to the maximum amplitude of a sine wave). If the minimum error and maximum correlation coefficient between A and B are found in a certain period, the resemblance is identified. Another method is to use the CRD and CWD to perform a correlation, trend regression analysis and F-Test with time lag.

5.2.2 CRD and CWD patterns in the Kammanassie area

In the Varkieskloof area, the CRD curve patterns of the Purification Works East (PWE) rainfall station do not match with their CWD curve patterns in DP12, DP15 and DP27, but they match well with that of DP18 (Figure 5.11 and Figure 5.12). The matched patterns of the CRD and CWD in DP18 seem to have the quickest response between rainfall and the water level. It may be due to the fact that the borehole is located in the alluvial deposits near the banks of the Olifants River.

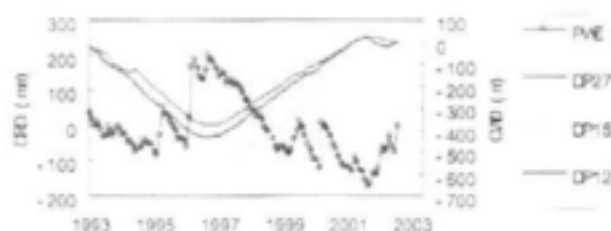


Figure 5.11 CWD patterns of boreholes versus CRD in the Varkieskloof area

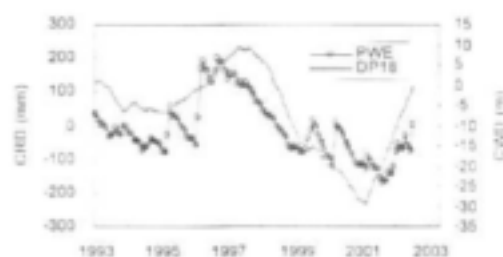


Figure 5.12 CWD patterns of borehole DP18 versus CRD in the PWE

The records of four rainfall stations were used for CRD analysis in the Kammanassie area from September 1993 to March 2003 (Figure 5.13).

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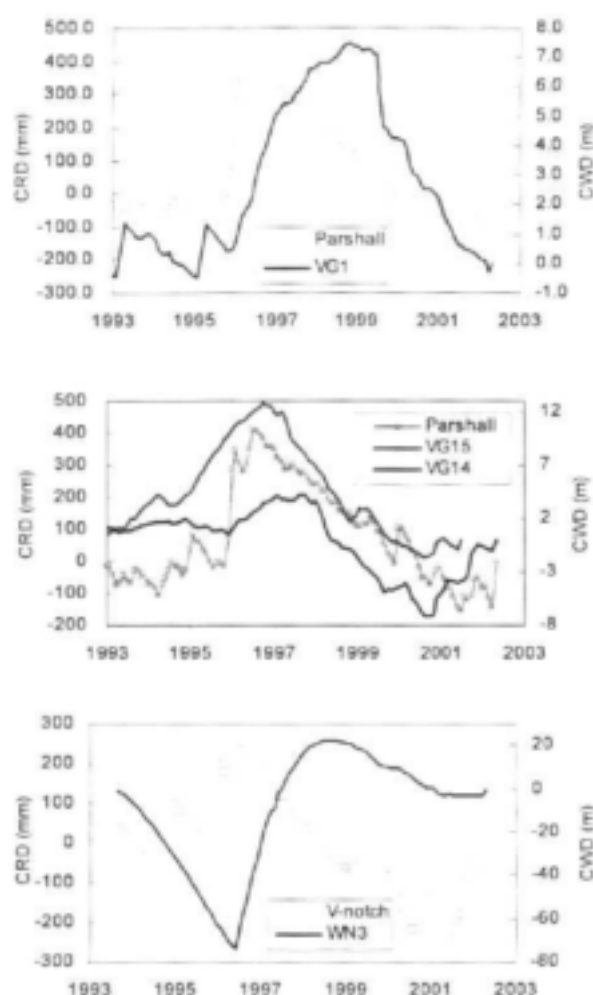


Figure 5.13 CWD patterns of boreholes versus CRD in the Kammanassie area

In the Voorzorg area, the CWD of VG 1 and the CRD of the Parshall rainfall station show a matching pattern. The CWD of VG 14 and VG15 show a very good match with the CRD in the Parshall rainfall station as well. The clear peaks of these CWD patterns occur due to the high rainfall season of September to November 1996, although a different time lag exists. Borehole WN3 in the Marnewicks valley shows no match pattern between CWD and CRD, but the peak of the CWD seems to have similar peak to the boreholes in the Voorzorg area. These phenomena would imply that high rainfall events are very important to recharge.

5.2.3 CRD and CWD patterns in the Vermaaks River Wellfield

The groundwater levels of the production boreholes have been declining in the Vermaaks River Well field from 1994 to 2003 as can be seen in Figure 5.14. The water level recovered

a little after a few heavy rainfall events, such as in October 1996, December 2000 and May 2002. This is attributed to not only more recharge but also a reduction in abstraction (see to Figure 5.4).

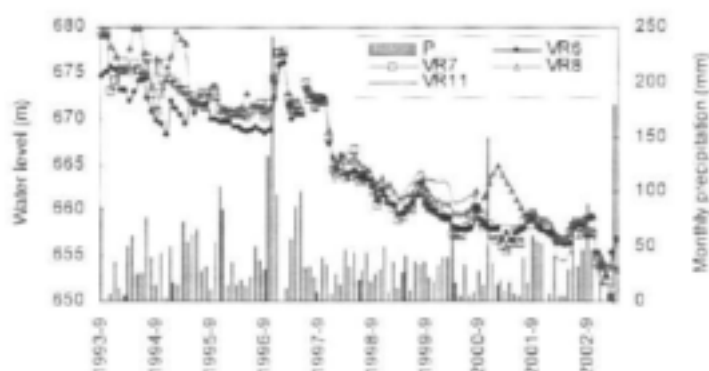


Figure 5.14 Groundwater levels in the Vermaaks River Well field

A similar CWD patterns occurred in VR6, VR7 and VR11 although there was a large difference in abstraction (Figure 5.14). A positive trend occurred in VR6, VR7 and VR11 from September 1994 until May 1998 and a negative trend from May 1998 onwards. There were some correlation between CRD and CWD in VR6, VR7 and VR11. A positive trend occurred in VR8 from September 1994 to January 2001 and a negative trend from January 1998 onwards. The peak point on the CWD matched the peak point on the CRD with a 49 month lag for monitoring boreholes of G40171 and G40173 (Figure 5.16).

The variances in CRD of the Wildebeesvlakte rainfall station versus the CWD of the VR6, VR7 and VR8 are not significantly different based on an F-test for the one-tailed probability test (Figure 5.17). The coefficient of correlation (R^2) has the maximum value while the lag time is 52 months. However, the one-tailed probability of the CRD of the Wildebeesvlakte rainfall station versus the CWD of the VR6, VR7 and VR8 are 0.121, 0.020 and 0.0177, respectively. Therefore, there is a significant correlation between CRD of the Wildebeesvlakte rainfall station and the CWD of the VR6, VR7 and VR8. It is important to note that the size of the time series, which is used to perform time lag analysis reduce from 157 to 106 with increase in time lag. It would increase the coefficient of correlation (R^2). That is why there is a higher coefficient of correlation (R^2) with 0.76 after a 45 months lag time (see to Figure 5.8). Although there are no time lag between the CRD and the CWD in VR6 and VR7, cross-correlation analysis can determine the time lag between the curves. The lag time is not clear between P-WL and P-P_{av}-WL-W_{av} pattern (Figure 5.18 and Figure 5.19). These types of curves cannot be used in lag time analysis in the Vermaaks River Well field. It is important to note that the best match of the CRD with CWD would be selected for recharge estimation.

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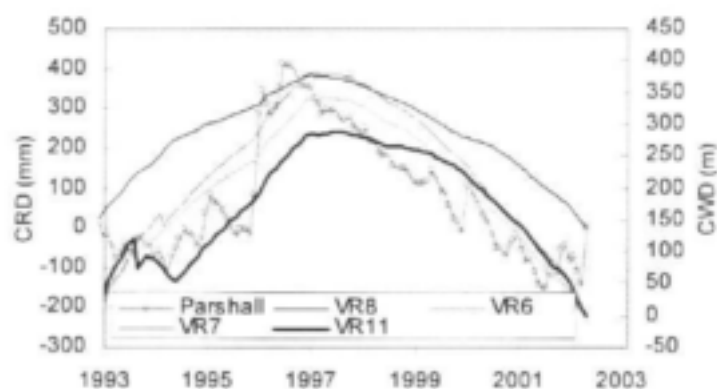


Figure 5.15 CWD patterns of production boreholes versus CRD of the Parshall rainfall station

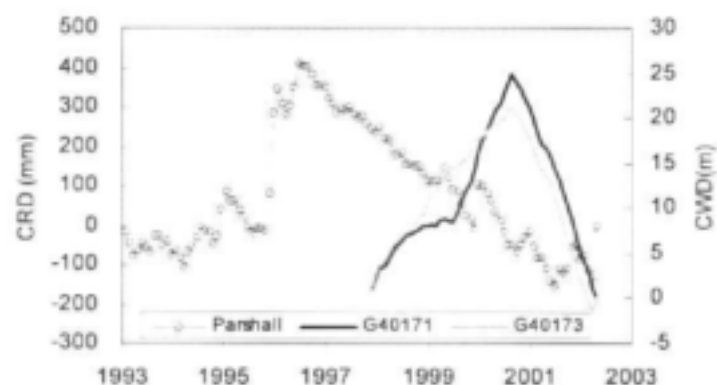


Figure 5.16 CWD patterns of G40171 and G40173 versus CRD of the Parshall rainfall station

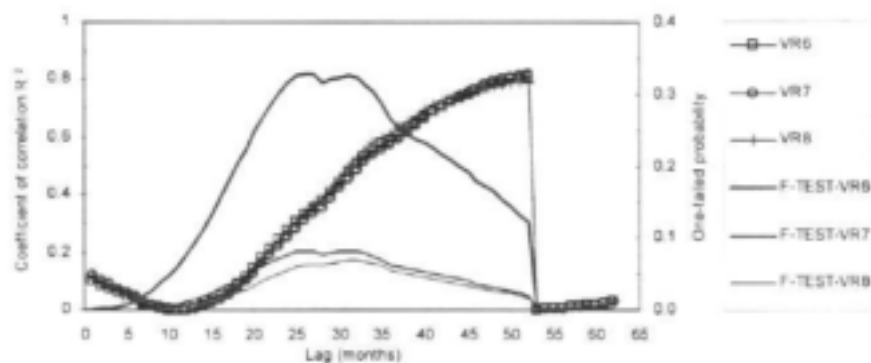


Figure 5.17 Time lag and F-Test of CWD and CRD in the Vermaaks River Well field

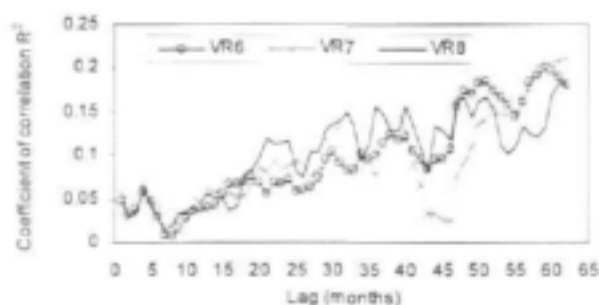


Figure 5.18 Time lag analysis using WL versus P of the Wildebeesvlakte in the Vermaak's River Well field

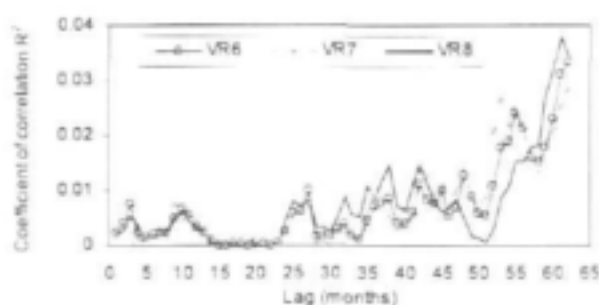


Figure 5.19 Time lag analysis of WL-WL_{av} vs P-P_{av} of Wildebeesvlakte in the Vermaak's River Well field

To confirm the cyclical period of rainfall, the auto-regression of the rainfall time series in the Wildebeesvlakte was performed (Figure 5.20). The cyclical pattern is not clear due to length of the time series. In other words, the time series is not long enough to illustrate the long-term effect of the rainfall on the water levels.

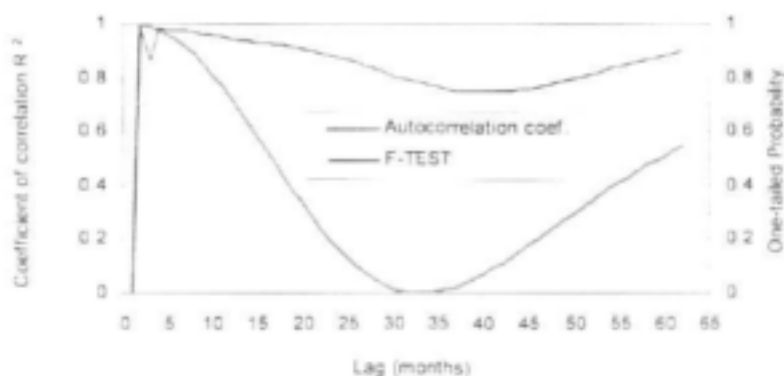


Figure 5.20 Auto-regression lag analysis and F-Test in the Wildebeesvlakte rainfall station

5.2.4 Recharge estimation with CRD and CWD

Figure 5.21 is an example of the regression analysis for recharge estimation. The recharge rates vary from 0.24% to 24.34% of the MAP in VG15 using the CWD-CRD regression analysis based on different storativity (Table 5.5). This suggests that the accuracy of estimates is dependent on the storativity of the aquifer. An example of the correlation coefficients of the CRD of different rainfall stations and the CWD of G40171 is listed in Table 5.6. The observations are as follows:

1. The coefficients of correlation and regression constant are positive values.
2. The recharge rates are positively related to storativity (S): high recharge rates are associated with high storativity.
3. High recharge rates range from 27.4% to 75.64% of the precipitation if storativity is 1%. These values are obviously overestimated.
4. Each CRD-CWD match could be used to estimate recharge rate due to relevant high coefficient of correlation. This suggests that one match with the highest coefficient of correlation should be adapted preferentially. In this study, the best one is CRD of the V-notch versus CWD of G40171.

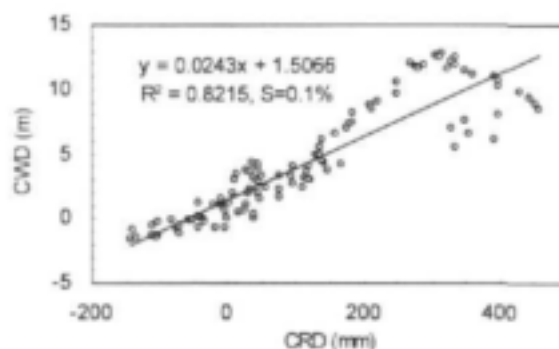


Figure 5.21 Regression of CWD- CRD of VG15

Table 5.5 Regression analysis results of VG15 using CRD of the Parshall rainfall station

S (%)	Regression equation	RE (%)	C (mm)	RE (mm)
0.1	$CWD = 0.0024CRD + 0.1507$	0.24	0.15	1.07
0.5	$CWD = 0.0122CRD + 0.7533$	1.22	0.75	5.46
1	$CWD = 0.0243CRD + 1.5066$	2.43	1.51	10.88
5	$CWD = 0.1217CRD + 7.5328$	12.17	7.53	54.48
10	$CWD = 0.2434CRD + 15.066$	24.34	15.07	108.96

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Table 5.6 Results of regression analysis of G40171

Coefficient of correlation	S (%)	Regression equation	RE (%)	C (mm)	RE _m (mm)
V-notch 0.83	0.1	$CWD = 0.0076CRD + 0.51$	0.76	0.51	2.88
	0.5	$CWD = 0.0378CRD + 2.55$	3.78	2.55	14.31
	1	$CWD = 0.0756CRD + 5.1$	7.56	5.10	28.62
	5	$CWD = 0.3781CRD + 25.49$	37.81	25.49	143.15
	10	$CWD = 0.7562CRD + 50.98$	75.62	50.98	286.31
Purification Works East 0.76	0.1	$CWD = 0.0061CRD + 0.901$	0.61	0.90	1.94
	0.5	$CWD = 0.0305CRD + 4.50$	3.05	4.50	9.69
	1	$CWD = 0.061CRD + 9.01$	6.1	9.01	19.37
	5	$CWD = 0.3048CRD + 45.03$	30.48	45.03	96.8
	10	$CWD = 0.6095CRD + 90.06$	60.95	90.06	193.56
Wildebeesvlake 0.70	0.1	$CWD = 0.003CRD + 0.53$	0.304	0.53	1.45
	0.5	$CWD = 0.0152CRD + 2.66$	1.52	2.66	7.27
	1	$CWD = 0.0304CRD + 5.32$	3.04	5.32	14.54
	5	$CWD = 0.1518CRD + 26.61$	15.18	26.61	72.58
	10	$CWD = 0.3035CRD + 53.21$	30.35	53.21	145.11
Parshall 0.67	0.1	$CWD = 0.0027CRD + 0.762$	0.27	0.76	1.21
	0.5	$CWD = 0.0137CRD + 3.81$	1.37	3.81	6.13
	1	$CWD = 0.0274CRD + 7.62$	2.74	7.62	12.27
	5	$CWD = 0.137CRD + 38.11$	13.7	38.11	61.33
	10	$CWD = 0.274CRD + 76.22$	27.40	76.22	122.66

The recharge estimates of borehole VR6 using the CWD-CRD regression analysis is listed in Table 5.7. The recharge rates vary from 5.98% to 299.12% based on storativity from 0.01% to 0.5%. The results could not be used in practice although the correlation coefficient is 0.86 ($R^2=0.74$) because abstractions were ignored. It also illustrates that the CRD-CWD method is unsuitable for estimating the recharge under conditions where the storativity is unreliable, even though the correlation coefficient is very high. The CRD-CWD method cannot work adequately under most pumping conditions because the effective radius could not be calculated precisely, that is, more errors would be introduced due to uncertainty factors, especially under unsteady state and discontinuous pumping conditions.

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Table 5.7 Regression analysis results of VR6 using CRD in the Parshall rainfall

S (%)	Regression equation	RE (%)	C (mm)	RE (mm)
0.1	$CWD = 0.0598CRD + 10.685$	5.98	10.69	22.64
0.5	$CWD = 0.2991CRD + 53.424$	29.91	53.42	113.24
1	$CWD = 0.5982CRD + 106.85$	59.82	106.85	226.49
5	$CWD = 2.9912CRD + 534.24$	299.12	534.24	1132.50

5.2.5 Comparison of the results with RIB

The water level series is simulated using a computer program named RIB (Xu and Beekman, 2003). The average pumping rate is 820 m³/d for the production boreholes. Simulated water levels are compared to observed levels in Figure 5.22, where dh (crd) refers to change in water level calculated by CRD method (Bredenkamp *et al.*, 1995), while dh (rib) represent water level change calculated using the RIB equation (Xu and Beekman, 2003). For storativity of 0.1% and recharge area of 11 km², the average recharge rates modelled for VR6, VR7, VR8 and VR11 are 2.25%, 2.55%, 2.14% and 2.17% of the rainfall, respectively, although there are disparities in curve matches. These results are consistent with the results by CRD-CWD as previously indicated.

5.2.6 Discussion

The value of the regression constant (C_0) will always be a positive value because the aquifer has some storage before the water level fluctuation occurs. The C_0 value could be related to ranges of water level under a no recharge condition, and then the C_0 value is the basic water level change. For instance, there is a 5.1 m basic water level change in G40171 under conditions of $C_0 = 5.1$ mm and $s = 0.1\%$ if no recharge occurs. If the water level range is 4.05 m (C_0) this means 4.05 mm recharge under condition of $s = 0.1\%$ in G40171.

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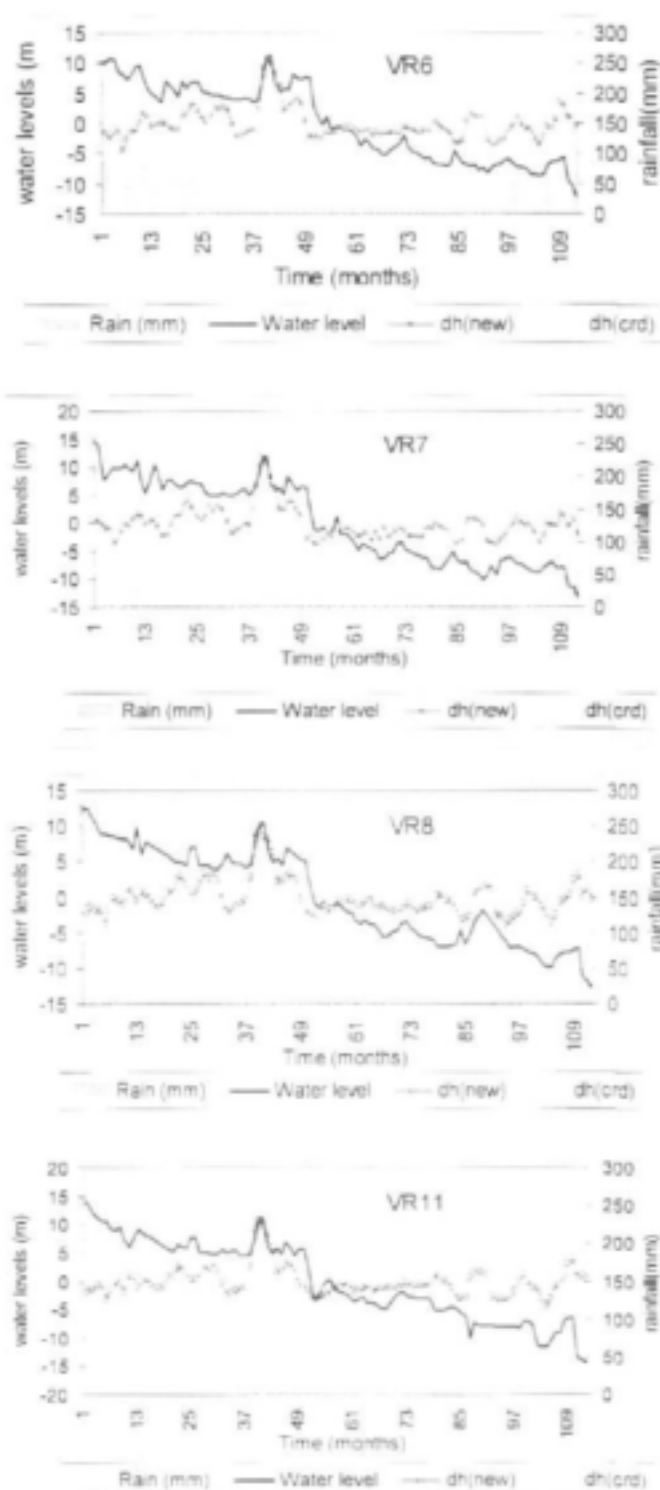


Figure 5.22 Results of RIB analysis in the Vermaak's River Well field

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Storativity is a very sensitive parameter in estimating recharge with CWD-CRD method. Based on the storativity range of 0.1% - 0.22% (the storativity data was adopted from Kotze, 2002), the recharge would be 7.56% - 16.63% and 2.43% - 5.35% of the rainfall in G40171 and VG15, respectively. The large difference of recharge occurs because of the uncertainty with storativity.

In order to confirm the storativity, different water level fluctuation and storativity are assumed to simulate the recharge area of the hot springs. If the area simulated is close to the natural recharge area, the storativity is assumed to be reliable. The recharge areas simulated are listed in Table 5.8 and Table 5.9. The simulated recharge area with 24,600 km² and 24,156.5 km² are close to the area of 24,650 km² based on the geological boundaries if the storativity is 0.01%-0.05%. It confirms that the storativity of the TMG should be from 0.01% to 0.1%. Based on the above discussion, the recharge rate of 0.24%-7.56% of the rainfall seems to be creditable in G40171 under storativity of 0.01%-0.1%.

Table 5.8 Recharge area simulation based on the storativity and flow rates of hot springs
(1.0m water fluctuation)

Storativity (%)			0.1	0.5	1	5	10
Specific recharge (m ³ ·yr ⁻¹ ·km ⁻²)			100	500	1000	5000	10000
Hot Spring	l·s ⁻¹	m ³ ·yr ⁻¹	Recharge area required for change in water level by 1 m				
The Baths (Citrusdal)	29	914,544	9,145.4	1,829.1	914.5	182.9	91.5
Goudini	11	346,896	3,469	693.8	346.9	69.4	34.7
Brandvlei	127	4,005,072	40050.7	8,010.1	4,005.1	801	400.5
Caledon	9	283,824	2,838.2	567.6	283.8	56.8	28.4
Avalon(Montagu)	38	1,198,368	11,983.7	2,396.7	1,198.4	239.7	119.8
Baden	37	1,166,832	11,668.3	2,333.7	1,166.8	233.4	116.7
Warmwaterberg	37	1,166,832	11,668.3	2,333.7	1,166.8	233.4	116.7
Calitzdorp	8	252,288	2,522.9	504.6	252.3	50.5	25.2
Tooverwater	11	346,896	3,469	693.8	346.9	69.4	34.7
Stuutis	31	977,616	9,776.2	1,955.2	977.6	195.5	97.8
Uitenhage	45	1,419,120	14,191.2	2,838.2	1,419.1	283.8	141.9
Total	383	12,078,288	120,782.9	2,4156.5	1,2078.2	2,415.8	1,207.9

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Table 5.9 Recharge area simulation based on the storativity and flow rates of hot springs
(3.0 m water fluctuation)

Storativity (%)			0.1	0.5	1	5	10
Special recharge ($\text{m}^3 \cdot \text{yr}^{-1} \cdot \text{km}^{-2}$)			300	1,500	3,000	1,5000	30,000
Hot Spring	$\text{l} \cdot \text{s}^{-1}$	$\text{m}^3 \cdot \text{yr}^{-1}$	Recharge area required for change in water level by 3 m (km^2)				
The Baths (Citrusdal)	29	914,544	3,048.5	609.7	304.8	61	30.5
Goudini	11	346,896	1,156.3	231.3	115.6	23.1	11.6
Brandvlei	127	4,005,072	13,350.2	2670	1335	267	133.5
Caledon	9	283,824	946.1	189.2	94.6	18.9	9.5
Avalon (Montagu)	38	1,198,368	3,994.6	798.9	399.5	79.9	39.9
Baden	37	1,166,832	3,889.4	777.9	388.9	77.8	38.9
Warmwaterberg	37	1,166,832	3,889.4	777.9	388.9	77.8	38.9
Calitzdorp	8	252,288	841	168.2	84.1	16.8	8.4
Tooverwater	11	346,896	1,156.3	231.3	115.6	23.1	11.6
Studtis	31	977,616	3,258.7	651.7	325.9	65.2	32.6
Uitenhage	45	1,419,120	4,730.4	946.1	473	94.6	47.3
Total	383	12,078,288	2.46E+04	1.00E+01	2.04E-03	8.31E-08	1.69E-12

5.3 Recharge Estimation with Spring Flow

Spring flow fluctuation is a direct indicator of groundwater recharge. Two perennial ponds occur at the geological contact zone between the Peninsula and Cedarburg Formations in the Vermaak and Marnewicks valleys. The springs are maintained in part by overflow from the Peninsula aquifer and a local shallow aquifer. Flows have been recorded since 1994 at the V-notch rainfall station, the monthly average fluxes, in litres, are available (Kotze, 2002). A good spring flux curve seems to respond to rainfall (Figure 5.23). Large flows appeared after a time lag of a few months after the rainfall. The curve based on rainfall-spring is therefore used to estimate recharge by regression.

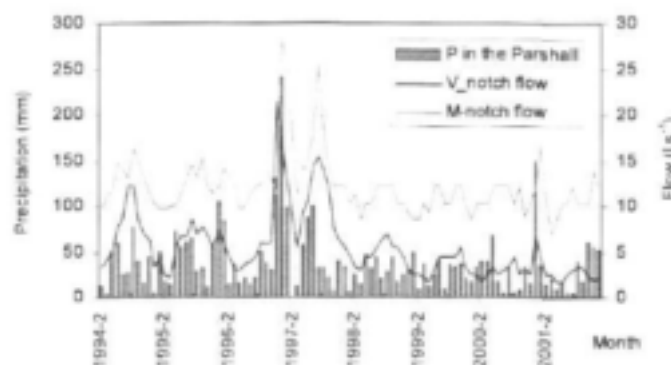


Figure 5.23 Hydrographs of the Vermaaks and Marnewicks Rivers

5.3.1 Theory

According to the hydrogeological model, the spring flux is a fraction of the total rainfall, assuming the recharge area is equal to the catchment area, a simple water balance over a given time interval i can be formulated as follows:

$$\sum_{i=1}^n Q_{si} = r \sum_{i=1}^n Q_{Ri} + c \quad (i = 1, 2, 3 \dots n) \quad \text{with} \quad (5.25)$$

$$Q_{si} = q_s \times A \times t_i$$

where Q_{si} is spring flux, Q_{Ri} is rainfall, A is the recharge area; q_s is the special flow of the spring and t_i is the time. c is a constant of regression.

5.3.2 Regression

5.3.2.1 Vermaaks catchment

A statistical analysis was carried out to examine relationships between cumulative flow (CF) and rainfall (CR), cumulative flow departure (CFD) and rainfall departure (CRD). The results are given in Table 5.10. A high correlation coefficient 0.9963 between CF of the V-notch and CR of the Parshall rainfall station exists. The one-tailed probabilities from the F-Test between CF in the Vermaaks and Marnewicks and CR in the Parshall are 0.190 and 0.183, respectively. The best correlation coefficient of 0.9966 between CF and rainfall exists in the preceding month. This suggests that rainfall from both the current and preceding months is important to estimate recharge. Therefore, the CF and CR are used to perform the regression analysis for recharge estimation.

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The results of recharge rate by regression analysis with V-notch spring data and rainfall in different rainfall stations from 1994 to 2001 are listed in Table 5.11. An example of a regression, based on flow in the V-notch and rainfall in the Parshall against the catchment area, is shown in Figure 5.24. The recharge of 3.7% of the precipitation is adopted because it has the highest correlation coefficient between the CR of the Parshall rainfall station and the CF. Considering about 11.32% of the spring water is derived from the local aquifer (8.37% from the shallow aquifer as monitoring borehole and 4.86% from Peninsula aquifer as production borehole), (see Chapter 4), the actual recharge rate should be 3.21% of the precipitation.

Table 5.10 Statistical results between different components

Pair of Factors		CF Versus CR ¹⁾	CF Versus CR	CFD Versus CRD	Flow Versus Rainfall	CRD Versus CTF	CR Versus CRD	Rainfall Versus CRD
Vermaaks	r	0.9966	0.9963	0.9032	0.5756	0.4038	0.3770	0.1154
	F-Test	1.61E-1	1.90E-1	2.3E-8	1.1E-4	2.3E-8	1.1E-5	8.4E-6
Marnewicks	r	0.9962	0.9959	0.7302	0.1454	0.2984	0.3770	0.1154
	F-Test	1.21E-1	1.83E-1	3.1E-8	2.9E-4	3.1E-8	5.93E-5	1.6E-5

$$^1) CR = \sum_{i=1}^n (0.2P_i + 0.8P_{i-1}) \quad i = 1, 2, \dots, n \text{ month}$$

Table 5.11 Recharge estimated with V-notch spring related to different rainfall stations
(from 1994 to 2001 by the CF method against recharge area of 10.85 km²)

Station	% of Precipitation	C	R ²
Parshall	3.7	3.11	0.9926
Wildebeesvlakte	3.54	1.74	0.991
V-notch	4.77	1.23	0.9874
Voorzorg	4.88	4.31	0.9842
Dorekloof	5.59	1.6	0.9882

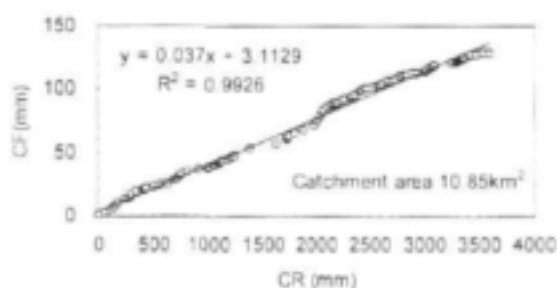


Figure 5.24 Regression based on flow in the V-notch and rainfall in the Parshall against the catchment area

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5.3.2.2 Marnewicks catchment

The same method of regression analysis is used in the analysis of the Marnewicks Spring. The recharge rates range from 3.65% to 5.76% of the precipitation from different rainfall stations for 1994 to 2001 using the CF method against the area of Marnewicks catchment (Table 5.12). A recharge rate of 3.65% of the precipitation is related to the maximum correlation coefficient between CR in the Wildebeesvlakte rainfall station and CF in the Marnewicks flow gauge (Figure 5.25).

Table 5.12 Recharge rates related to different rainfall station from 1994 to 2001 against the Marnewicks catchment area

Station	% of precipitation	C	R ²
Parshall	3.82	-2.8	0.9954
Wildebeesvlakte	3.65	-4.3	0.9961
V-notch	4.93	-4.83	0.9926
Voorzorg	5.05	-1.85	0.9947
Dorekloof	5.76	-7.58	0.9989

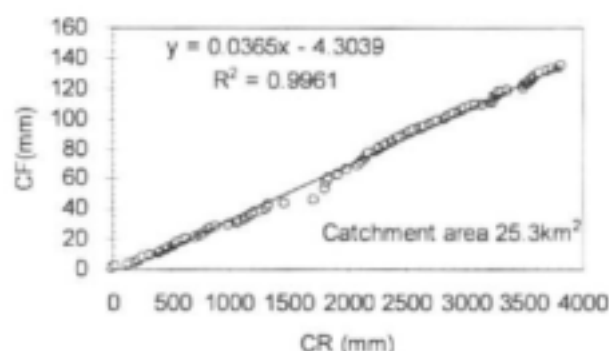


Figure 5.25 Regression based on flow in the M-notch and rainfall in the Wildebeesvlakte against the catchment area

5.3.3 Discussion

5.3.3.1 Impact of flood event on recharge

The recharge rate before the flood is slightly higher than that after the flood in all rainfall stations (Table 5.13). If the floods are removed, the recharge rate ranges from 3.93% to 4.03%, based on different rainfall station data from 1994 to 2001 (Table 5.14).

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Table 5.13 Recharge rate against recharge area of 10.85 km² from 1994 to 2001

Station	Before flood			After flood		
	% of P	C	R ²	% of P	C	R ²
Parshall	3.55	3.37	0.9873	3.36	10.87	0.9622
Wildebeesvlakte	3.53	0.8	0.9919	3.31	11.24	0.9575
V-notch	4.27	2.53	0.9906	4.07	12.25	0.9489
Voorzorg	5.38	-0.33	0.9923	3.87	11.34	0.956
Dorekloof	4.74	1.72	0.988	5.05	14.43	0.9484

Table 5.14 Regression results with runoff removed from 1994 to 2001

Catchment	Station	% of the P	C	R ²	Flood removed (mm)
Vermaaks	Parshall	4.03	5.56	0.991	250.9
Marnewicks	Wildebeesvlakte	3.97	-4.51	0.995	270.2
Total	Wildebeesvlakte	3.93	-2.68	0.994	270.2

5.3.3.2 Recharge area

Equation (5.25) has shown that the recharge area is a principal factor in regression analysis when special flow is converted to a volume. The recharge related to the recharge area of the Vermaaks and Marnewicks catchments are shown in Figure 5.26. It illustrates that the recharge rate is recharge-area dependent. This is the limitation of the method. The recharge rate ranges from 4.01% to 3.29% based on a $\pm 10\%$ difference between the natural catchment and recharge area. This means there is a recharge rate increase of 0.09% of the precipitation if the recharge area is reduced by 10%, and recharge rate decrease of 0.33% of the precipitation if the area increases by 10%. The recharge rate is underestimated because the impact of abstraction is ignored.

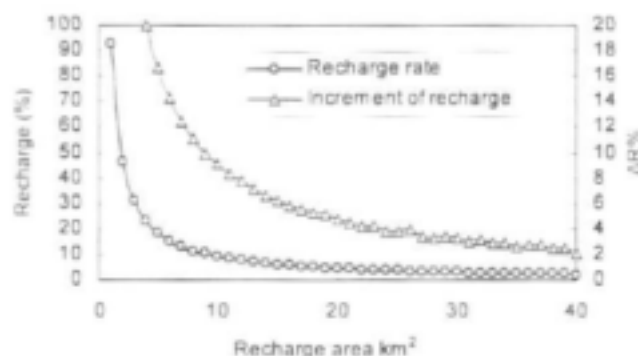


Figure 5.26 Relationship between recharge rates and recharge area

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Due to the hydrogeological contact of the Marnewicks and Vermaaks catchments, the recharge rate of 3.2% of the precipitation is computed against two catchment areas as shown in Figure 5.27 (13.23% of the aquifer leakage was taken out). This result is only slightly different from the result of the separate catchment. The result removes the impact of the watershed of groundwater catchment between the two catchments.

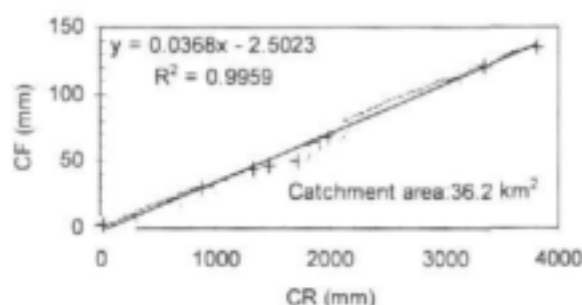


Figure 5.27 Regression based on total flow in the two catchments and rainfall in the Wildebeesvlakte rainfall station

5.3.3.3 Regression constant

The slope of the regression equation or the recharge rate is given in Table 5.11 to Table 5.13. The source of base flow could be deduced by the constant in the regression equation. A positive constant c occurs in the Vermaaks catchment. The influence of abstraction leads to the dewatering of the aquifer in the upper Vermaaks catchment. Spring flow is not only fed by recharge, but also by aquifer storage. There is a very small or no aquifer storage contribution to the spring in the Marnewicks catchment because a negative c value occurs.

5.3.3.4 Impact of preceding rainfall on spring flux

Impact of preceding rainfall on spring flux can be computed as shown in section 5.3.7. The recharge rate is calculated as follows:

$$RE = \frac{\sum \Delta H}{\sum x} 100\% \quad (5.26)$$

where $\sum \Delta H$ is cumulative increment of flux and $\sum x$ is cumulative rainfall which influences recharge in time series.

Three periods with an increasing spring flux were observed from 1994 to 2003. The recharge rates are calculated using the Parshall rainfall data over an upper catchment area of 10.85 km² as presented in Appendix XII(b). The maximum recharge rate is 5.21% of the

precipitation, which is associated with a preceding rainfall of 257 mm from February to June 1997. The minimum recharge rate is 1.73% of precipitation, which is related to preceding rainfall of 97 mm from April to July 1998. The recharge rate is linked to the rainfall of the study period, such as period 3; the 1.38% and 1.92% of the precipitation are related to 42 mm and 37 mm, respectively. The recharge rate is 2.42% of the precipitation of 867 mm from February 1997 to March 1999. The recharge rate is 0.81% of the precipitation of 2,593 mm from January 1994 to March 1999 if two or more years lag. This may be attributed to the exclusion of the recharge contribution to the spring flow.

5.4 Numeric Simulation

Assuming the abstraction of the Vermaaks River Wellfield from 1993 to 2002 is equal to the recharge rate for the same period, it gives crude recharge estimates in percentage of precipitation as 11.09% and 2.96% over the areas of the upper Vermaaks catchment (10.85 km²) and Peninsula Formation Window (40.6 km²), respectively. As the water levels in the wellfield experienced a continued decline over the period, the recharge rate estimated above appears exaggerated and larger than realistic recharge rate. In other words, the realistic recharge rate must be smaller than the crude estimates.

For confirming the water budget, two 3D models were constructed (Wu, 2005). The aquifer thickness is 750 m deduced from Figure 4.20. In the model, the Cedarberg shale acts as an inactive boundary and the abstraction from the Vermaaks River Wellfield could be coming from the entire Peninsula Formation closed by Cedarberg Formation. The boundary of the Peninsula Formation and Nardouw Subgroup is another inactive boundary in the south. The parameters of the models are listed in Table 5.15. The initial transmissivity can be referred to in Table 4.9. The horizontal transmissivity assigned to each layer is based on weight factors of borehole penetration and water strike of 0.1 and 0.9, respectively. The ratio of horizontal transmissivity to vertical transmissivity (T_h/T_v) is 10. The time step is 10 seconds in the transient model. The period modelled is from 1993 to 2002.

Flows are calculated for each zone in each layer and each time step. The unit of the flows is [m³·s⁻¹]. Flows are considered IN, if they are entering a zone. Flows between sub-regions are given in a Flow Matrix. HORIZ. EXCHANGE gives the flow rate horizontally across the boundary of a zone. EXCHANGE (UPPER) gives the flow rate coming from (IN) or going to (OUT) to the upper adjacent layer. EXCHANGE (LOWER) gives the flow rate coming from (IN) or going to (OUT) to the lower adjacent layer. The percentage discrepancy is calculated by $100 \cdot (IN - OUT) / (IN + OUT) / 2$ (Chiang and Kinzelbach, 2001).

The water budget at steady state using PMWIN is listed in Table 5.16. Calibration is performed using a recharge rate of 0.316% ($4.23 \times 10^{-11} \text{ m s}^{-1}$) from a three-component mixing model in the transient model and the results are listed in Table 5.17. The observations are obtained as follows:

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- For the model A, negative water balance occurs if recharge is $1.22 \times 10^{-10} \text{ m s}^{-1}$ and $2.44 \times 10^{-10} \text{ m s}^{-1}$. The discrepancy is -65.54% and -1.5%, respectively. Positive budget occurs if the recharge rate is $2.5 \times 10^{-10} \text{ m s}^{-1}$ in the model A. It means that the recharge rate cannot be higher than 1.96% of rainfall because the water level had been declining since 1993.
- For the model B, the same scenario of the water budget occurs. A positive budget occurs if the recharge rate is $2.427 \times 10^{-10} \text{ m s}^{-1}$ in the model B.
- The simulated water level is higher than those measured if the recharge rate is $4.23 \times 10^{-11} \text{ m s}^{-1}$. The errors of VR 6, VR7, VR8 and VR11 between water levels from the transient flow model and measurement are 1.8, 2.41, 1.18 and 3.50 m, respectively.

Table 5.15 Parameters of the models

Model	Layer Thickness (m)		T ($\text{m}^2 \cdot \text{s}^{-1}$)	n_e	Pumping ($\text{m}^3 \cdot \text{s}^{-1}$)			
A	1	50	6.55E-5 (Fold limb) 6.42E-4 (Fold axis)	0.5~2‰	VR6	VR7	VR8	VR11
	2	50	6.55E-5 (Fold limb) 6.42E-4 (Fold axis)	1~2‰	-	3.64E-3	8.44E-4	7.08E-4
	3	50	6.55E-5 (Fold limb) 5E-4 (Fold axis)	1~2‰	-	3.64E-3	8.44E-4	1.42E-3
	4	50	6.55E-5 (Fold limb) 6.42E-4 (Fold axis)	1.5~2‰	-	-	8.44E-4	-
	5	50	6.55E-5 (Fold limb) 6.42E-4 (Fold axis)	1~1.5‰	3.50E-3	-	-	-
	6	50	6.55E-05	1‰	-	-	-	-
	7	50	6.55E-05	1‰	-	-	-	-
	8	50	6.55E-05	1‰	-	-	-	-
	9	50	6.55E-05	1‰	-	-	-	-
	10	50	6.55E-05	1‰	-	-	-	-
	11	50	6.55E-05	1‰	-	-	-	-
	12	50	6.55E-05	1‰	-	-	-	-
	13	50	6.55E-05	1‰	-	-	-	-
	14	50	2.65E-05	1‰	-	-	-	-
	15	50	2.65E-05	1‰	-	-	-	-
B	1	250	6.55E-5 (Fold limb) 6.42E-4 (Fold axis)	0.5~2‰	3.50E-3	7.28E-3	2.53E-3	2.128E-3

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Table 5.16 Water budget of the steady state model flow

	Recharge rate (%)	Recharge ($\text{m}\cdot\text{s}^{-1}$)	Discrepancy of the Peninsula Formation Window budget (%)
Model A	0.96	1.22×10^{-10}	-65.54
	1.91	2.44×10^{-10}	-1.50
	1.96	2.5×10^{-10}	0.91
Model B	0.96	1.22×10^{-10}	-63.89
	1.88	2.40×10^{-10}	-0.3
	1.90	2.427×10^{-10}	0.75
	1.91	2.44×10^{-10}	1.32

Table 5.17 10 year- Water level with recharge of $4.23 \times 10^{-11} \text{ m}\cdot\text{s}^{-1}$ at transient state for model A

Boreholes	VR6	VR7	VR8	VR11
Measured water level in Dec. 2002 (m)	654.40	653.87	654.82	652.27
Simulation water level in Dec. 2002(m)	656.20	656.28	656.00	655.77
Error (m)	1.80	2.41	1.18	3.50

5.5 Integrated Recharge Rate in the Kammanassie Area

It is well known that different results are produced from different methods. Each result represents the degree of closeness to the realistic recharge rate based on the understanding of the recharge mechanisms and hydrogeological settings. Recharge estimates could be grouped into manageable and unmanageable ones depending on types of data used. The recharge rates estimated from the abstractions of the production boreholes are of manageable recharge results as these are verified by actual pumpages. The others may represent the theoretical recharge. The objective of this section is to develop a means for integrating various estimates in the TMG area. The uncertainties of factors relating to the methods are analysed, and the reliability of the conversion would be improved by analysing raw and empirical data. The determination of weighting factors is considered as the largest uncertainties for the model in the study area. A method is developed to calculate the weighting factors based on the role of the factors in recharge estimation processes. Four patterns of recharge rate can be estimated, namely a: highly precision and highly bias, b: highly precision and lowly bias, c: lowly precision and lowly bias, d: lowly discrete and highly bias (Figure 5.28). The four patterns would be utilised to formulate a formula for calculating an integrated recharge rate.

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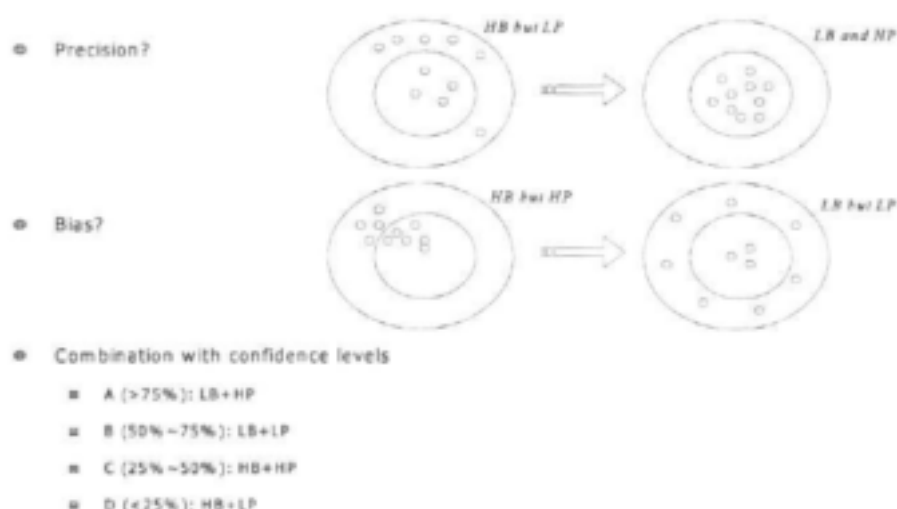


Figure 5.28 Four confidence levels of recharge estimates

The factors of the methods are listed in Table 5.18. The factors, size of the mixing model, CWD-CRD, regression of CF as well as water balance are 4, 3, 3 and 9, respectively. Generally, the results derived from the complication of effect factors may create more discrepancy because of error propagation.

Table 5.18 Factors of the methods in the study

Method	Factor	Size	Size of domination factor
CMB Mixing Model	C_1 , C_2 , C_3 , recharge area A (refer to equation 10.8 and equation 10.9)	4	recharge area A
CWD-CRD	CWD, CRD and storativity of the aquifer (refer to equation 10.16)	3	storativity
Regression of CF	CR, CF, recharge area A (refer to equation 10.25)	3	recharge area A
Water Balance	Precipitation (P), maximum (T_{max}), minimum (T_{min}), soil factors (f_s), vegetation cover (f_v), rainfall last time (f_t), CN, R_A and λ (refer to equation 11.3, equation 11.7 and equation 11.8)	9	Precipitation (P), difference of the temperature, soil factors (f_s) and Vegetation cover

For the purpose of better expression of the different results, the integrated recharge rate is computed as follows:

$$R_{int} = \sum_{i=1}^n R_i W_i \quad (5.27)$$

where R_{int} is integrated recharge rate, R_i is the result of the i^{th} method, W_i is the weight of the i^{th} method.

The total factors of methods considered is defined as

$$W = \sum_{i=1}^n S_i, \quad i=1,2,\dots,n \text{ method} \quad (5.28)$$

where S_i is the factor size of the i^{th} method.

The weight of the i^{th} method is defined as

$$M_i = \left[\frac{S_i}{W} \right]^{-1} \quad (5.29)$$

The normalization weight is defined as

$$F_i = \left[\sum_{i=1}^n M_i \right]^{-1}, \quad i=1,2,\dots,n \quad (5.30)$$

The normalization weight of domination factor can be written as

$$F_d = \left[\sum_{i=1}^n M_d \right]^{-1}, \quad i=1,2,\dots,n \quad (5.31)$$

The integrated weight is written as

$$W_i = \frac{F_i \times F_d}{\sum_{i=1}^n (F_i \times F_d)}, \quad i=1,2,\dots,n \quad (5.32)$$

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The factor size and integrated recharge rate are listed in Table 5.19. The integrated recharge rates relating to total factors, dominating factors and the integrated weight of the four methods are 1.68%-4.72%, 1.65%-4.52% and 1.63%-4.75% of the precipitation, respectively. The recharge rates of 1.63%-4.75% of the precipitation are recommended in the Kammanassie area.

Table 5.19 Factor size and integrated recharge estimates

Recharge Method	Total Factors		Dominating Factors		Integrated Weight
	Size	Weight	Size	Weight	
CMB Mixing Model	4	0.243	1	0.308	0.26
CWD-CRD	3	0.324	1	0.308	0.35
Regression of CF	3	0.324	1	0.308	0.35
Water Balance	9	0.108	4	0.077	0.03
Integrated recharge estimates (%)	Minimum	1.68	1.65		1.63
	Maximum	4.72	4.52		4.75

5.6 Confidence Interval for Recharge Rates

The complexities of the hydrogeological settings make estimating local or regional recharge difficult. The spatial variation of rainfall and topographic conditions result in complex recharge mechanisms. Not all of the components of the mixing model can be measured accurately within a given catchment. Inputs such as precipitation and abstraction are quantifiable; however, the time lag has changed significantly relative to the residence time of the water, and the solutions of three-component model mostly rely on the results from the two-component model, even the mixing model cannot include piston flow, which likely occurs in the preferential fracture system. The recharge area size may not be determined accurately in a fractured rock aquifer. Under these limitations, estimates of recharge from the mixing model are subject to large potential errors.

The sum of the deviations of the X_i values from their mean $\bar{X}$ is zero in a time series. This may lead to incorrect resemblances identified from the CWD-CRD curves. The aquifer area is different from the catchment area, and the storativity of the aquifer is inexact and has high variation. These would produce more discrepancy in the CWD-CRD, the CTF-CTR and the other area-dependent or storativity-dependent methods.

The results of recharge estimation from several methods are summarized in Appendix XIII (Wu, 2005). The range of the recharge estimates is 1.07-28.62 mm·a⁻¹ (0.21%-7.56%) with the coefficient of variation from 46.07% to 68.16% (Table 5.20). The estimates have significant variation. The recharge interval based on the 95% and 90% confidence is 12.18-14.92 mm·a⁻¹ (2.17%-2.93%) and 12.4-14.70 mm·a⁻¹ (2.23%-2.87%), respectively.

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Table 5.20 Outline of the recharge estimates

Statistical analysis	Annual recharge Rate (%)	Annual Recharge (mm·a ⁻¹)
Minimum	0.24	1.07
Maximum	7.56	28.62
Mean	2.55	13.55
Median	2.20	14.86
Standard deviation	1.74	6.24
Coefficient of variation (%)	68.16	46.07
Confidence interval (95%) Equation (2.9)	2.17-2.93	12.18-14.92
Confidence interval (90%) Equation (2.9)	2.23-2.87	12.4-14.70

To reduce the heavy reliance either on the assumption or on the sample sizes, the nonparametric procedures (Neter *et al.*, 1988) are conducted. The nonparametric procedures are concerned with the median of a single population. When the population is highly skewed, the population median (η) is more centralised in the distribution than the population mean and thus may be more representative. When the population is symmetrical, the population mean and the population median coincide and thus are equally meaningful measures of central tendency (Neter *et al.*, 1988). The inference procedures for the population median have the following two assumptions (Neter *et al.*, 1988):

- The population is continuous.
- The sample is a random one.

The estimated recharge rates satisfy the two assumptions obviously; therefore, a confidence interval for the population median of the recharge rate is performed. A random sample of 80-recharge rate has been selected in ascending order. It is desired to construct a 95% and 90% confidence interval for the population median amount of recharge rate. The integer numbers of r selected are 31 and 33 for 95% and 90% confidence interval, respectively.

For 95% confidence interval, the required confidence limits for η are the 31st-smallest (L_{31}) and the 31st-largest (U_{31}) recharge rate in the 80 samples of recharge estimates, they are: $10.88\text{ mm} \leq \eta \leq 16.09\text{ mm}$, or $1.65\% \leq \eta \leq 3.30\%$; for 90% confidence interval, the required confidence limits for η are the 33th-smallest (L_{33}) and the 33th-largest (U_{33}) recharge rate in the 80 results of recharge estimates, they are: $12.27\text{ mm} \leq \eta \leq 15.83\text{ mm}$, or

$1.89\% \leq \eta \leq 3.04\%$ (Figure 5.29). Based on the discussion, the range of annual recharge rate for 95% and 90% confidence interval is 10.88-16.09 mm (1.65%-3.30%) and 12.27-15.83 mm (1.89%-3.04%), respectively.

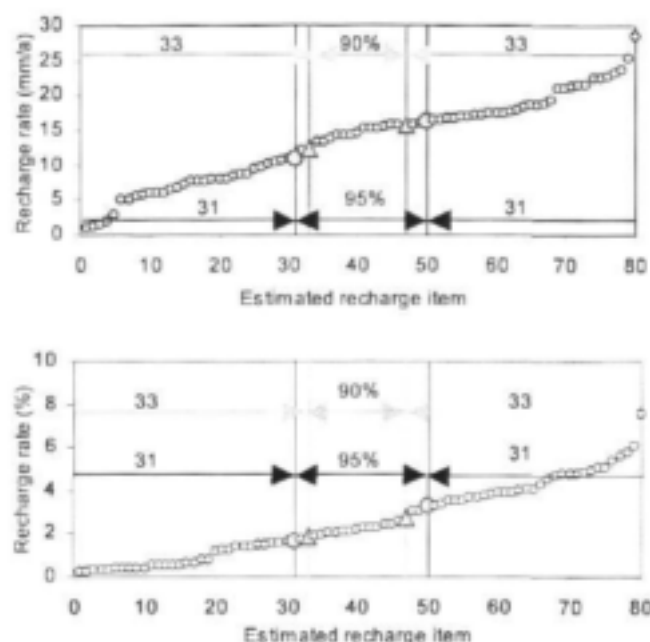


Figure 5.29 A high confidence interval for the population median amount of recharge estimates

5.7 Other Case Studies

In the Montagu Area, recharge estimation was using Water Balance approach is touched on. In the Montagu area of the Western Klein Karoo, the TMG aquifer comprises approximately 4,000 m thick sequence of sandstone with an estimated outcrop area of 3,124 km². The outcrops are characterized by mountainous topography with sparse vegetation, shallow and intermittent diverse soils and mean annual precipitation of 350-450 mm/yr (Sun, 2005). Based on detailed analysis and interpretation of factors influencing recharge, Sun used the water balance method to estimate recharge rates with rainfall, stream flow and temperature data in the area. By employing empirical evapotranspiration and run off models to determine and analyse long-term average recharge she modelled the long-term recharge as a function of the regional interaction of the site conditions of climate soil, geology and topography.

Recharge ranges vary from 0.1 – 38.0 mm/yr in the study area. The values less than 20.0 mm/yr were predominant. Recharge is less than 5.0 mm/yr if MAP is less than 400 mm/yr. Ranges of 10.0-20.0 mm/yr of recharge occur in precipitation ranging from 600-1,200 mm/yr. The recharge rates exceeding 20.0 mm/yr are more related to the precipitation of 800 mm/yr or more. The low recharge rates less than 2.0 mm/yr are related to singly high rainfall events.

Although recharge rates increase with increasing precipitation, the recharge percentage is in a non-linear relationship with the precipitation. Spatial distribution of the recharge is associated with the variations in precipitation, geological and morphological settings in the study area.

In the case of the Uitenhage Spring of the Eastern Cape, a recharge rate of 11% of the average annual rainfall is estimated based on the spring flow simulation, while the deep groundwater flow into the confined section of the Coega Ridge Aquifer is suggested as 3% of the total recharge by the isotope method (Xu and Maclear, 2003).

In another study, recharge estimates are also made using Water Balance approach (Jia, 2006). But results are generally biased towards higher end of estimated range. Water balance based method fails to separate interflow from total rainfall infiltration and its recharge estimates include the component of interflow. Advantage of this approach is simple and rapid.

As a part of the project, case studies carried out by the Council of Scientific and Industrial Research (CSIR) before the start of this project is reviewed and evaluated (Beekman et al., 2004). CSIR has conducted research in the TMG areas of Agter-Witzenberg, Jonkershoek, Klein Karoo, Struisbaai, and Zachariashoek. The Agter-Witzenberg and Struisbaai areas were selected for detailed evaluation and verification of recharge estimation. Groundwater recharge in the Agter-Witzenberg area was quantified by Weaver et al. (1999) using the Chloride Mass Balance method. A review of chloride data revealed a recharge rate of ~30% of annual rainfall (579 mm) over the period May 1995 to May 1996. Weaver (1998) demonstrated that groundwater of the Struisbaai aquifer originates from rainfall, and that the salinity of the groundwater is mostly explained by the input of sea-spray and dry deposition. The groundwater model of Weaver et al. (1999) was also revised and upgraded to accommodate for major deficiencies and the sensitivity of the revised model for recharge estimation was determined. Within the range of transmissivities as derived from pumping tests, recharge varies between 1 and 20% of average annual rainfall. The recharge values as derived from the CMB method fall within this range. At this stage for both Agter-Witzenberg and Struisbaai the groundwater models cannot further refine the recharge estimates. More detailed investigations are needed to apply the method of inverse groundwater modelling for precise recharge estimation.

6. SPATIAL DISTRIBUTION OF RECHARGE IN THE TMG AREA

A water balance analysis is often used to estimate hydrologic fluxes in an atmospheric water balance, soil water balance and surface water balance on different scales. Groundwater recharge is only a surplus from the soil-water budget. It appears negligible in most hydrologic studies because the recharge is only a small component of the process. 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 and Sokolov, 1983; Manabe, 1969; Mather, 1972; Alley, 1993; Willmott *et al.*, 1985; Mintz and Walker, 1993; Mintz and Serafini, 1992). This approach is attractive because of its simplicity, which requires minimal input data: precipitation, potential evapotranspiration and soil-water holding capacity. In its simplest form, the soil-water budget model does not account for situations where the precipitation rate is greater than the infiltration capacity of the soil.

6.1 Water Balance Approach

6.1.1 Principle

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

Rainfall = Runoff - Direct + Actual Evapotranspiration $\pm$ Change of Field Moisture Capacity of Soils/Rocks + Groundwater Recharge

Total runoff comprises direct runoff and base flow. Direct runoff includes surface runoff and interflow as follows:

- Surface runoff: precipitation that cannot be absorbed by the soil because the soil is already saturated with water (saturation excess overland flow); precipitation that exceeds infiltration; the portion of rain, snow melt, irrigation water, or other water that moves across the land surface and enters a wetland, stream, or another body of water (overland flow). Overland flow usually occurs where the soils are very fine textured or heavily compacted (response time about half an hour in a 1 km² basin).
- Interflow that travels laterally or horizontally through the zone of aeration (vadose zone) with high hydraulic connectivity during or immediately after a precipitation event and discharges into a stream or other body of water (response time of several hours in a 1 km² basin, more lag time if the slope is longer. Its hydrogeochemistry with high concentration of chloride appears to have characteristics of leachate.)

- **Field Moisture Capacity:** The total amount of water remaining in a freely drained soil after the excess has flowed into the underlying unsaturated soil.

Based on the above definitions, the water balance equation is written as

$$R_E(t) = P(t) - E(t) - R(t) \pm \Delta\omega(t) \quad (6.1)$$

where $R_E(t)$ is groundwater recharge, $P(t)$ is precipitation, $E(t)$ is actual evapotranspiration, $R(t)$ is direct runoff and $\Delta\omega(t)$ is change in storage. 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. Assuming $\Delta\omega(t)$ is negligible between subsequent years, an annual estimate of $R_E(t)$ can be calculated following the flow diagram shown in Figure 6.1.

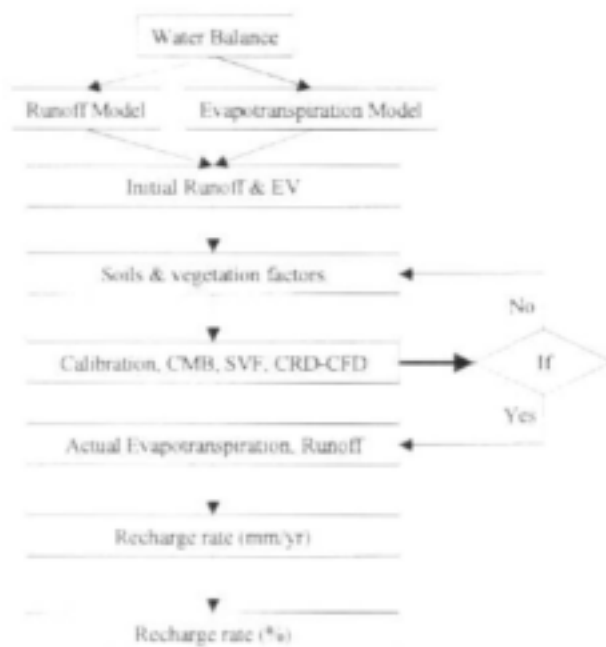


Figure 6.1 A flow diagram of recharge estimation based on water balance

6.1.2 Model selection

A common approach for approximating $E(t)$ involves initial estimates of evapotranspiration for a reference soil, vegetation and coverage of plants, $E_R(t)$, such as grass or Fynbos. Additive $E_A(t)$ is then calculated by multiplying $E_R(t)$ by a group coefficient, $K_c(t)$, an experimentally defined crop-specific and soil moisture variable whose value varies throughout the round year, such that:

$$E_r(t) = K_c(t)E_p(t) \quad (6.2)$$

We initial estimates $E_p(t)$ are as follows (Hargreaves and Samani, 1985):

$$E_p(t) = 0.0023R_A T_D^{1/2} (T + 17.8) \quad (6.3)$$

where T_D is the difference between average monthly maximum and minimum temperatures ($^{\circ}\text{C}$), T is average monthly temperature ($^{\circ}\text{C}$), R_A is extraterrestrial radiation ($\text{MJ m}^{-2} \text{ d}^{-1}$). R_A is represented as follows (Duffie and Beckman, 1980):

$$R_A = G_{sc} d_r \frac{(\omega_s) \sin(\phi) \sin(\delta) + \cos(\phi) \sin(\delta) \sin(\omega_s)}{\pi} \quad (6.4)$$

where ϕ is latitude (radians; positive for north, negative for south), and

$$\begin{aligned} \delta &= 0.4093 \sin[2\pi(284 + j)/365] \\ d_r &= 1 + 0.033 \cos(2\pi J/365) \\ \omega_s &= \arccos[-\tan(\phi) \tan(\delta)] \end{aligned}$$

G_{sc} is the solar constant with a value of $118.1 \text{ MJ m}^{-2} \text{ d}^{-1}$, and J represents the number of days since January 1 of the current year. Values from equation (6.5) were converted to units of md^{-1} by:

$$E_p(t)[\text{md}^{-1}] = \frac{E_p(t)[\text{MJm}^{-2}\text{d}^{-1}]}{\lambda[\text{MJkg}^{-1}]\rho[\text{kgm}^{-3}]} \quad (6.5)$$

where ρ is the density of water; λ is the latent heat of vaporization, given by (Harrison, 1963):

$$\lambda[\text{MJkg}^{-1}] = 2.501 - 2.361 \times 10^{-3} T [T \text{ in } ^{\circ}\text{C}] \quad (6.6)$$

The $K_c(t)$ is expressed as

$$K_c(t) = \frac{1}{200} P(f_s + f_v \times f_c + f_i)(t) \quad (6.7)$$

where P is the precipitation (mm), f_s is the soil factor, f_v is the vegetation factor, f_c is the coverage of vegetation, f_i is the factor of intensity of the rainfall with $f_i = h_r/T_t$, h_r is the rainfall lasting time, T_t is the monthly total time.

A common method to estimate Q is to apply the runoff equation (USDA-SCS, 1985):

$$Q_r = \frac{(P - 0.2S)^2}{P + 0.8S}; \quad P \geq 0.2S \quad \text{with}$$

$$S = \frac{1000}{CN} - 10 \quad (6.8)$$

where CN is the curve number. CN , based on site conditions ranges from 36 to 99 (USDA-SCS, 1985). Considering the features of the landform of the TMG, the ranges of CN are 80 to 98. Measured $P(t)$ was used in equation (6.8) to estimate Q_r .

6.1.3 Calibration of parameters

The parameters relating to the soil and vegetation are empirical values as shown in the equation (6.7). These values should be calibrated scientifically. The calibration of 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 limited within $\pm 2^\circ\text{C}$ because the long-term (1925-2003) change of the average temperature should not exceed 4°C . The range of the temperature is usually controlled within $\pm 0.5^\circ\text{C}$.
- The intensity of rainfall is considered as three classes, which are A: less than 5 hrs per month, B: 5 to 10 hrs per month and C: exceeds 10 hrs per month.
- The coverage percentage of vegetation is estimated based on field investigations, 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 soil factor values are empirical values in initial calculation process; the values given refer to the type, thickness and distribution of soil defined by Midgley *et al.* (1994). The calibrated value for soil factor is based on the trial calculation in the Kammanassie area with the critical value of annual recharge range of 10.88-16.09 mm derived from Chapter 10 (Wu, 2005). The soil factors in other TMG areas are given based on the trial calculation in the Kammanassie area but the error range does not exceed 10% under similar condition of lithology and topography.

6.2 Distribution of Mean Annual Precipitation (MAP) in the TMG

In discussing the distribution of MAP, the entire 478 rainfall stations, with their codes shown in Figure 6.2, were used in making a contour of MAP method in the TMG area. The raw data of the rainfall stations were adopted from WR90 (Midgley *et al.*, 1994). The MAP values higher than 500 mm occurred in the hydrogeological units 4, 5, 6, 15 and 17 (Figure 6.2). The precipitation in most areas of the other hydrogeological units is less than 500 mm·a⁻¹. The precipitation of more than 800 mm·a⁻¹ distributes in the mountainous areas close to Franschhoek, Knysna and Humansdorp. The precipitation in the north of the Outeniqua Mountains and Langeberg Ranges is obviously less than that in the southern side due to the mountainous barrier, which is the southern boundary of Klein Karoo. The precipitation gradually decreases from south close to the Indian Ocean (hydrogeological unit 15) to the north in the Karoo (hydrogeological unit 13). In the hydrogeological unit 2, the precipitation in the high elevation area of the Cedarberg Mountain is from 300 to 500 mm·a⁻¹, but rainfall in the lower elevation area is less than 300 mm·a⁻¹. The precipitation in the Swartberg Mountain (hydrogeological unit 13) ranges from 100 to 400 mm·a⁻¹. It is important to note that there are no rainfall stations in most of the mountainous areas of the TMG.

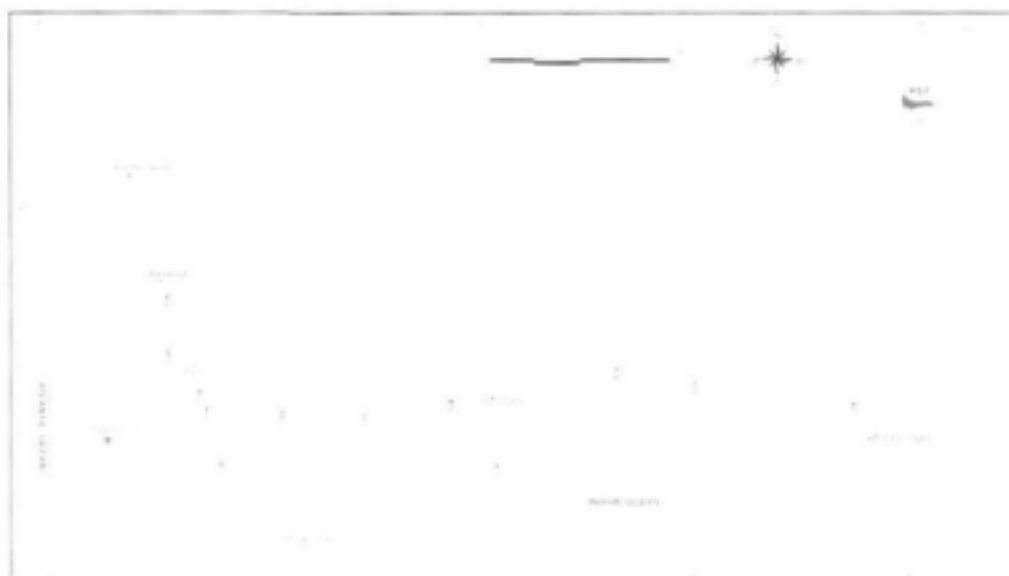


Figure 6.2 Rainfall station codes in the TMG area

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Figure 6.3 Annual average precipitation in the TMG area

6.3 Runoff and Actual Evapotranspiration

The runoff was estimated based on 175 rainfall stations near or on the outcrop of the TMG using equation (6.8). The runoff contour map is shown in Figure 6.4. Runoff is less than 150 mm·a⁻¹ in most areas. Low runoff amounts of less than 25 mm·a⁻¹ occur in the hydrogeological units 1, 11, 14 and 16. High runoff amounts of between 150-500 mm·a⁻¹ in green and blue occur in the hydrogeological units 5 and 8, but the data is more than 500 mm·a⁻¹ from WR90 (Midgley *et al.*, 1994). In general, the runoff values are smaller than that of WR90.

The data was selected before it was used in the calculation by comparing the background conditions, such as homogeneous rainfall districts, catchment size and potential evaporation. Soil factors were adopted according to types, texture and thickness of soils in 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 6.1. The factors of vegetation in the outcrop of the TMG are listed in Table 6.2. The temperatures used in initial evapotranspiration model are presented in Table 6.3.

Table 6.1 Soil factor and coverage percentages of vegetation

Hydrostratigraphic unit No.	Soil factor	Coverage percentage	Hydrostratigraphic unit No.	Soil factor	Coverage percentage
1	0.25-0.30	5-35	10	0.25-0.30	20-40
2	0.25-0.35	10-30	11	0.25-0.30	20-55
3	0.25-0.30	15-65	12	0.03-0.40	30-40
4	0.20-0.25	20-70	13	0.35-0.40	5-45
5	0.20-0.25	25-75	14	0.02-0.35	10-45
6	0.20-0.25	40-60	15	0.25-0.30	15-65
7	0.25-0.30	50-75	16	0.20-0.30	10-35
8	0.20-0.30	25-65	17	0.15-0.25	40-55
9	0.15-0.45	5-50	18	0.25-0.3	20-40

Table 6.2 Factors of vegetation in the outcrop of the TMG (Midgley *et al.*, 1994)

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

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Table 6.3 Model of temperature in the TMG area

(Temperature data obtained from CNC website)

Month	Greyton		Buffelsklip		Wildalsvlakte		Wildebeesvlakte		Elandsvlakte		Paardvlatte1		Paardvlatte2	
Temp. °C	Max	Min	Max	Min	Max	Min	Max	Min	Max	Min	Max	Min	Max	Min
Jan.	24.5	10	39	7.5	37	4	36	5	33	5	33	3	32	2.5
Feb.	26	12.5	40	6	35	4	36	4	34	5	33	5	31	2
Mar.	29	14	38	6	35	3	33.5	3	32	4	30	2	28	1.5
Apr.	30	15	32	2	30	1	30	1	28	2	27	1	25.5	0
May	30	15	29	2	26	0	26	-0.5	24	2	24	-1	21	-0.5
Jun.	28	14	26	0	22	-1	22	0	21	0	20	-4	19	-2.5
Jul.	25.5	10.5	25	-0.5	23	-0.5	22	-2	21	-1	19	-4	19	-2.5
Aug.	22.5	7.5	30	1	27	-1.5	25	-1	25	1	22	-4	21	-3
Sep.	20	4.5	32	2	30	0	29	-1	29	1	28	-4	25	-2
Oct.	19.5	4	35	3	33	0	32	1	31	3	27	1	28	1
Nov.	20	5.5	37	4	33	2	33	3	29	3	30	1	29	-0.5
Dec.	22	8	38	5	36.5	4	35	4	31	5	31	2	31	0
Ref. Alt. (m)	500-1500		<800		800-1200		800-1500		1000-1300		>1500		>1500	
MAP (mm)	400-500		<350		300-450		400-650		500-800		700-1000		>1000	

The actual annual evapotranspiration is calculated based on the annual average precipitation of 175 rainfall stations as can be seen in Figure 6.5. Evapotranspiration with 700-1300 mm per year are associated with greater than 900 mm of MAP in the hydrogeological units 5, 8 and 17. There are 700-900 mm a⁻¹ of evapotranspiration in some mountainous areas of the hydrogeological units 4, 7, 9, 15 and 18. The actual annual evapotranspiration is less than 500 mm a⁻¹ in most areas of the hydrogeological units 1, 2, 3, 9, 13 and 16. The distribution differentiates clearly from potential evapotranspiration of 1200-2600 mm a⁻¹ that is based on an open water surface shown in Figure 6.6. The potential evapotranspiration increases gradually from south to north.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

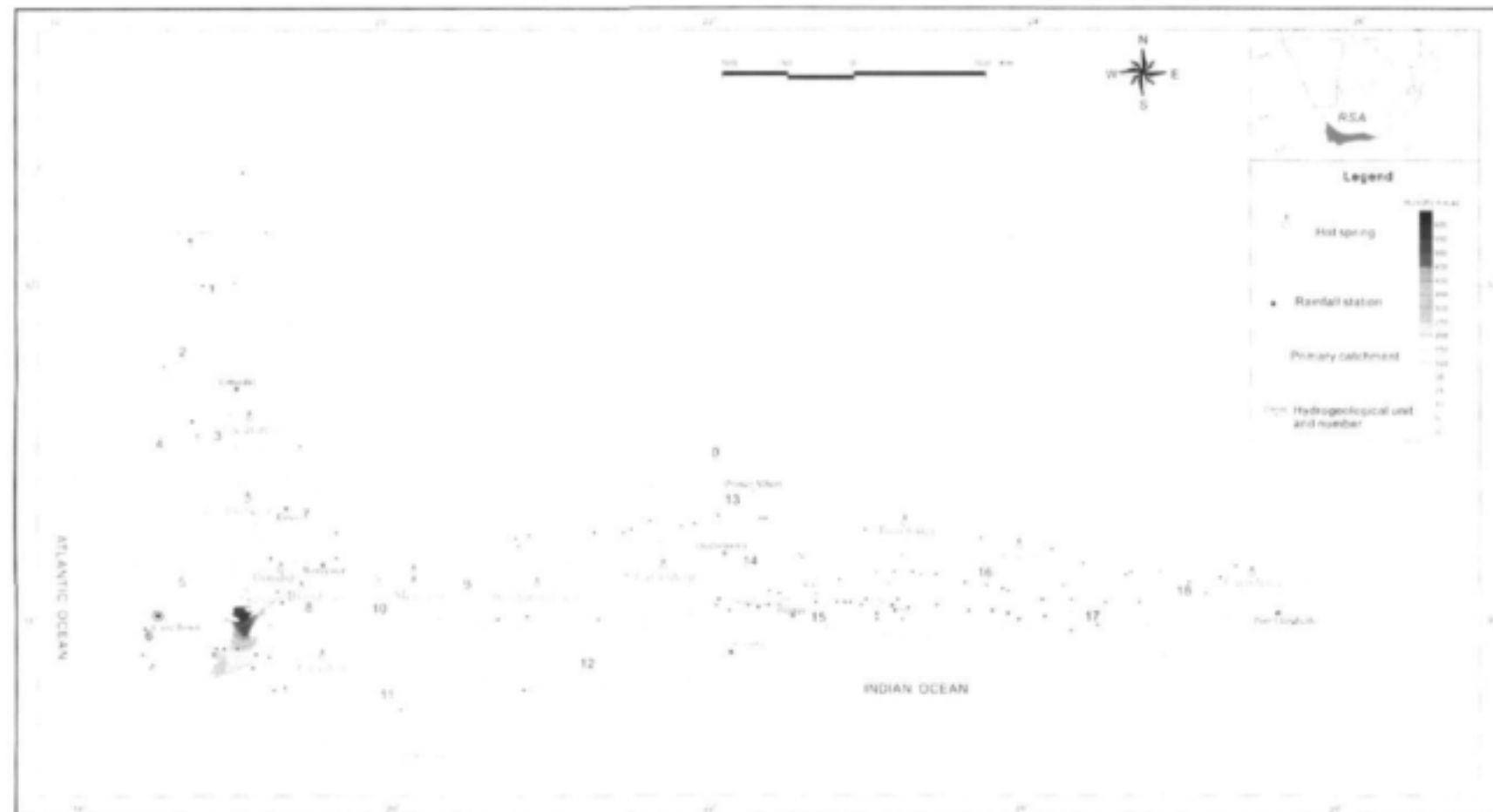


Figure 6.4 Annual average runoff in the TMG area

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES



Figure 6.5 Actual annual average evapotranspiration in the TMG area

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

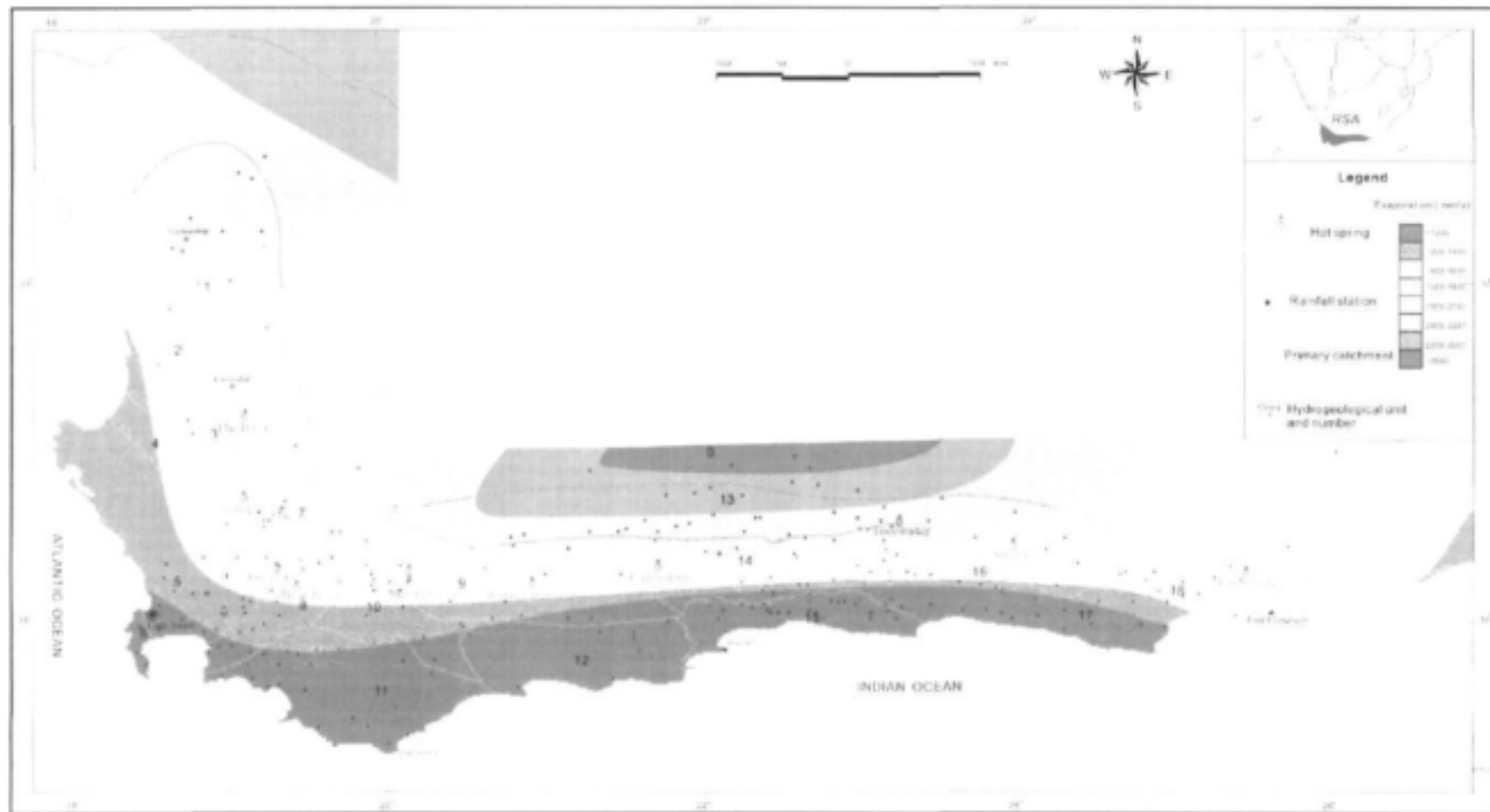


Figure 6.6 Potential annual average evapotranspiration in the TMG area (After WRC, 1990)

6.4 Recharge Rate

6.4.1 Distribution of the recharge rate

The annual average recharge rates given by the water balance method refer to the results from the study in chapter 5 in the modelling process as presented in Appendix XIV. The distribution of recharge rates is illustrated in Figure 6.7. The following observations are made:

- 1) The average recharge rate is $29.74 \text{ mm}\cdot\text{a}^{-1}$. The highest recharge rate is $137.4 \text{ mm}\cdot\text{a}^{-1}$ related to 1842.06 mm MAP at rainfall station No. 0022116 in the hydrogeological unit 5; the lowest recharge rate is $0.7 \text{ mm}\cdot\text{a}^{-1}$ related to 163.51 mm MAP at rainfall station No. 0048043 in the hydrogeological unit 13.
- 2) The regions of the recharge rate more than $35 \text{ mm}\cdot\text{a}^{-1}$ occur in the hydrogeological units 5, 6, 8 and parts of hydrogeological units 7 and 17. The other areas are less than $35 \text{ mm}\cdot\text{a}^{-1}$.
- 3) *The recharge rate is from 0.28% of the rainfall at station No. 0107169 in the hydrogeological unit 1 to 12.6% of the rainfall at station No. 0012220 in the hydrogeological unit 15.*
- 4) The recharge rate in the Agter-Witsenberg Mountain is $35.76\text{--}97.11 \text{ mm}\cdot\text{a}^{-1}$.
- 5) The distribution of the recharge rate in the eastern section from Riversdale to Cape St Francis has a clear boundary consistent with the outline of the mountain, in which a high recharge rate is from 35 to $65 \text{ mm}\cdot\text{a}^{-1}$ in the southern area (green-blue), and less than $35 \text{ mm}\cdot\text{a}^{-1}$ (red-yellow) in the northern area.

The recharge rates are smaller than that from Vegter (1995) as can be seen in Figure 6.8. It is clear that the recharge rates are the same in most areas of the recharge map, which has the same distribution in different climatic region and homogeneous rainfall districts.

The outlines of recharge rates in the TMG area are summarised in Table 6.4 and shown in Figure 6.9. The following observations are summarised:

- The biggest range is from 0.88 to $135.52 \text{ mm}\cdot\text{a}^{-1}$ in the hydrogeological unit 8, and the smallest range is 0.50 to $1.29 \text{ mm}\cdot\text{a}^{-1}$ in the hydrogeological unit 1.
- The average recharge rate is smaller than $10 \text{ mm}\cdot\text{a}^{-1}$ in the hydrogeological units 1, 10 and 13. The average recharge rate varies from $10 \text{ mm}\cdot\text{a}^{-1}$ to $25 \text{ mm}\cdot\text{a}^{-1}$ in the hydrogeological units 2, 9, 14, 16 and 18.
- The average recharge rates are from $25 \text{ mm}\cdot\text{a}^{-1}$ to $50 \text{ mm}\cdot\text{a}^{-1}$ in the hydrogeological units 4, 7, 11, 12, 15 and 17.
- The average recharge rate ranges from $50 \text{ mm}\cdot\text{a}^{-1}$ to $85 \text{ mm}\cdot\text{a}^{-1}$ in the hydrogeological units, 3, 5, 6 and 8.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

Table 6.4 Outline of the annual recharge estimates in the TMG area

Hydro- stratigraphic unit No	Annual average precipitation (mm)			Annual average recharge rate (mm)			Annual average recharge rate (%)		
	Minimum	Maximum	Mean	Minimum	Maximum	Mean	Minimum	Maximum	Mean
Unit 1	179.6	238.2	218.6	0.5	1.3	0.8	0.3	0.5	0.4
Unit 2	240.7	369.6	293.1	2.0	24.0	11.9	0.7	6.5	3.7
Unit 3	439.3	830.6	635.0	48.1	96.5	72.3	10.9	11.6	11.3
Unit 4	461.3	1002.7	718.2	11.9	43.4	26.1	2.6	5.1	3.6
Unit 5	558.3	2063.0	1291.7	35.1	137.4	82.4	4.0	8.7	6.4
Unit 6	584.7	922.7	695.1	43.8	67.7	56.4	4.7	11.4	8.5
Unit 7	286.7	1073.3	585.7	14.3	97.1	35.8	2.8	9.0	5.6
Unit 8	237.3	1604.8	854.6	0.9	135.5	75.1	0.4	12.4	8.4
Unit 9	132.4	999.0	442.2	0.9	52.5	13.3	0.4	8.2	2.4
Unit 10	256.6	482.1	331.9	2.2	8.3	5.8	2.0	3.1	2.4
Unit 11	404.5	928.9	525.5	24.7	64.6	41.5	3.2	12.3	8.3
Unit 12	354.8	454.4	417.2	11.4	41.6	27.8	2.5	9.4	6.7
Unit 13	104.9	553.5	277.6	0.7	17.8	4.9	0.4	5.6	1.8
Unit 14	165.7	1049.3	474.9	1.1	43.4	20.1	0.7	8.1	4.0
Unit 15	490.3	1099.3	837.8	3.1	93.0	49.4	2.4	10.4	6.4
Unit 16	256.0	602.3	401.1	9.1	51.3	16.6	1.6	3.8	3.0
Unit 17	415.9	1208.4	781.2	14.1	90.8	38.1	2.7	9.6	4.9
Unit 18	362.1	833.3	544.4	9.2	49.1	22.9	1.8	6.0	4.1

The recharge rate is shown in Figure 6.10. The recharge rates are higher than 11% of the precipitation (in yellow) occurring in the hydrogeological units 3 and 8, 6-11% of the precipitation (in green) in the hydrogeological units 3, 6, 8, 11, 12, 15 and 17; 2%-6% of the precipitation in blue in the hydrogeological units 1,2,4,7,9,10,13,14,16 and 18 and lower than 2% of the precipitation in the other areas. For management purposes, the recharge rates for the quaternary catchment in the outcrop of the TMG are presented in the Appendix XV.

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

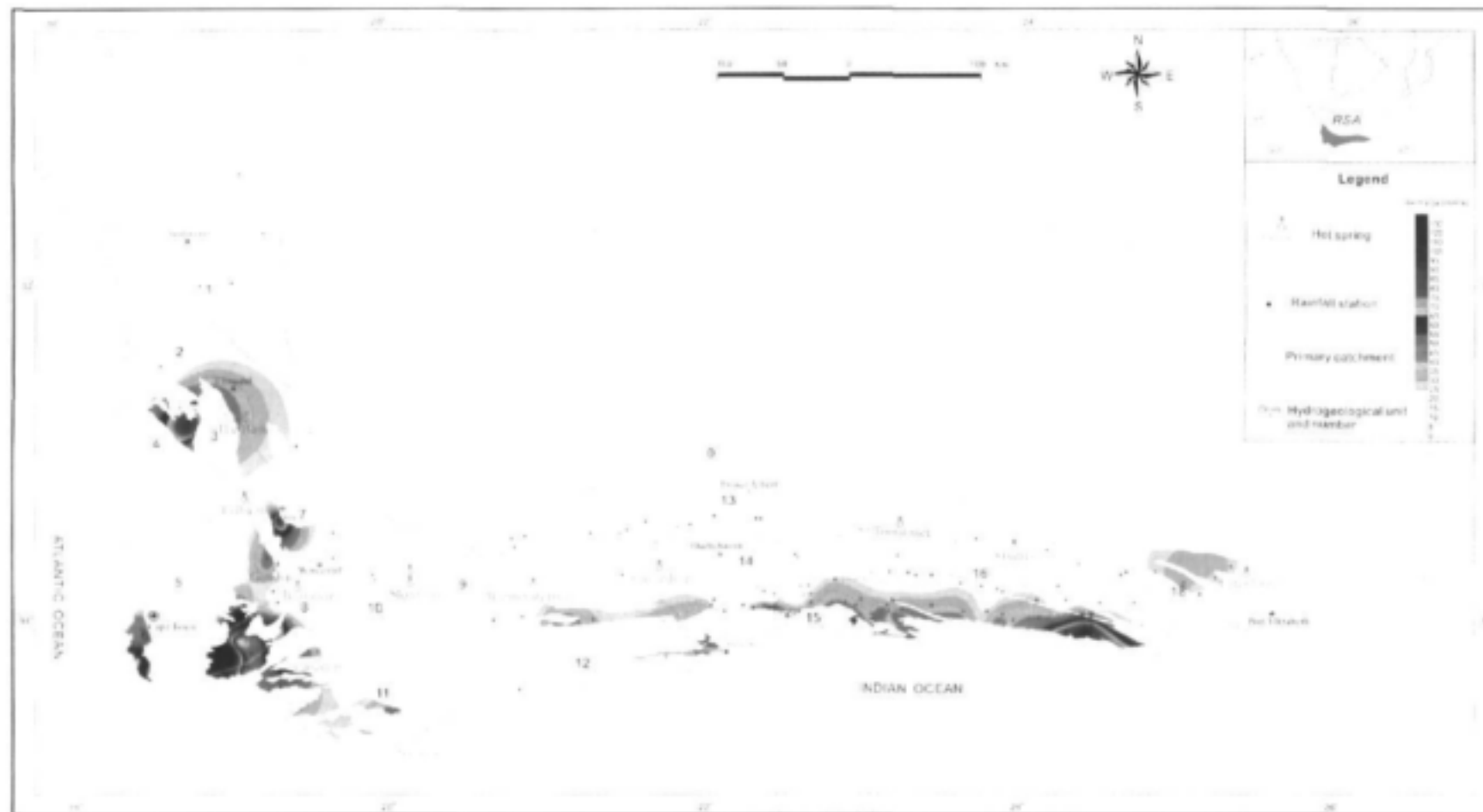


Figure 6.7 Annual average recharge rate in the TMG area

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

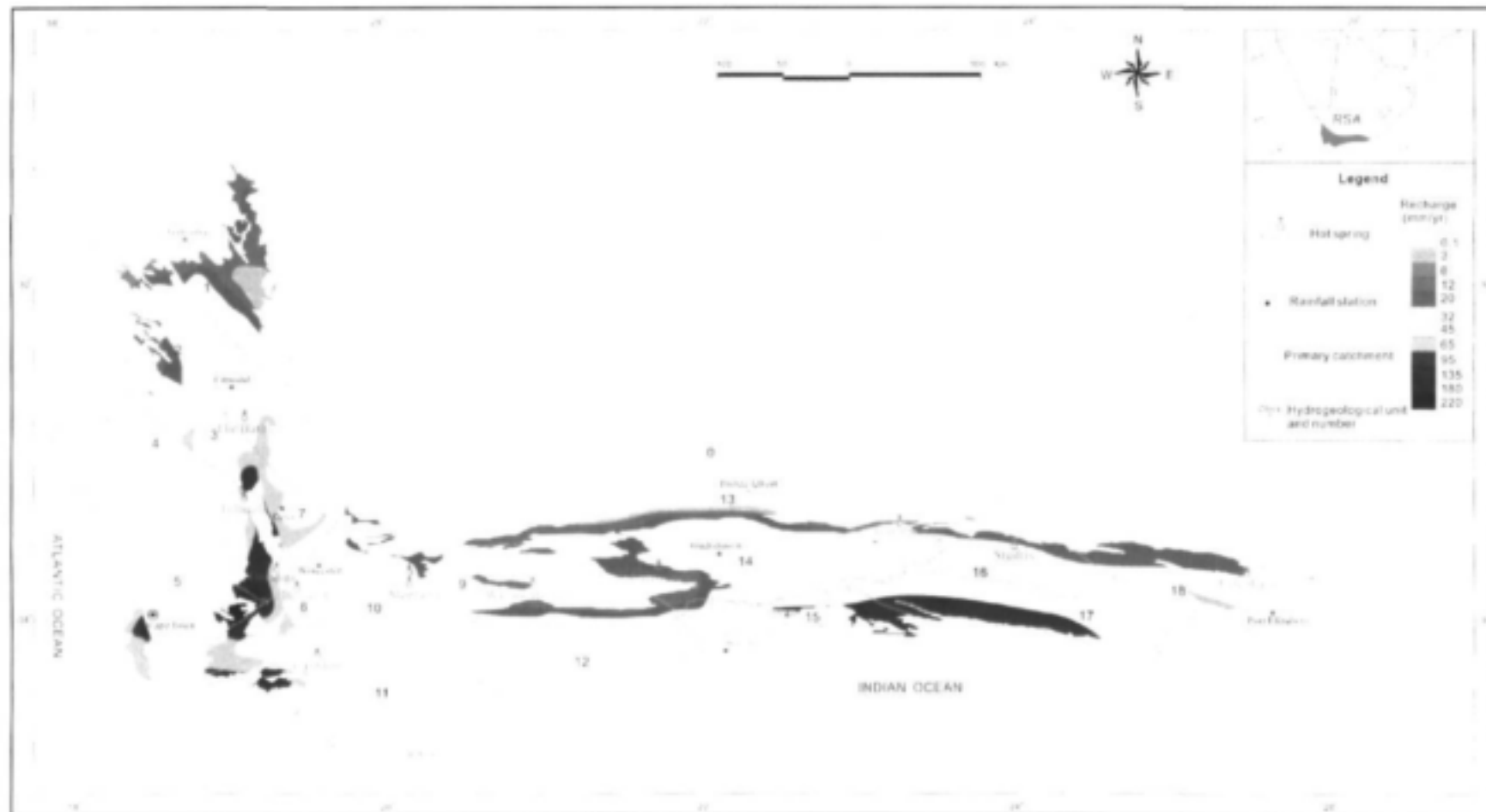


Figure 6.8 Annual average recharge rate in the TMG area (After Vegter, 1995)

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

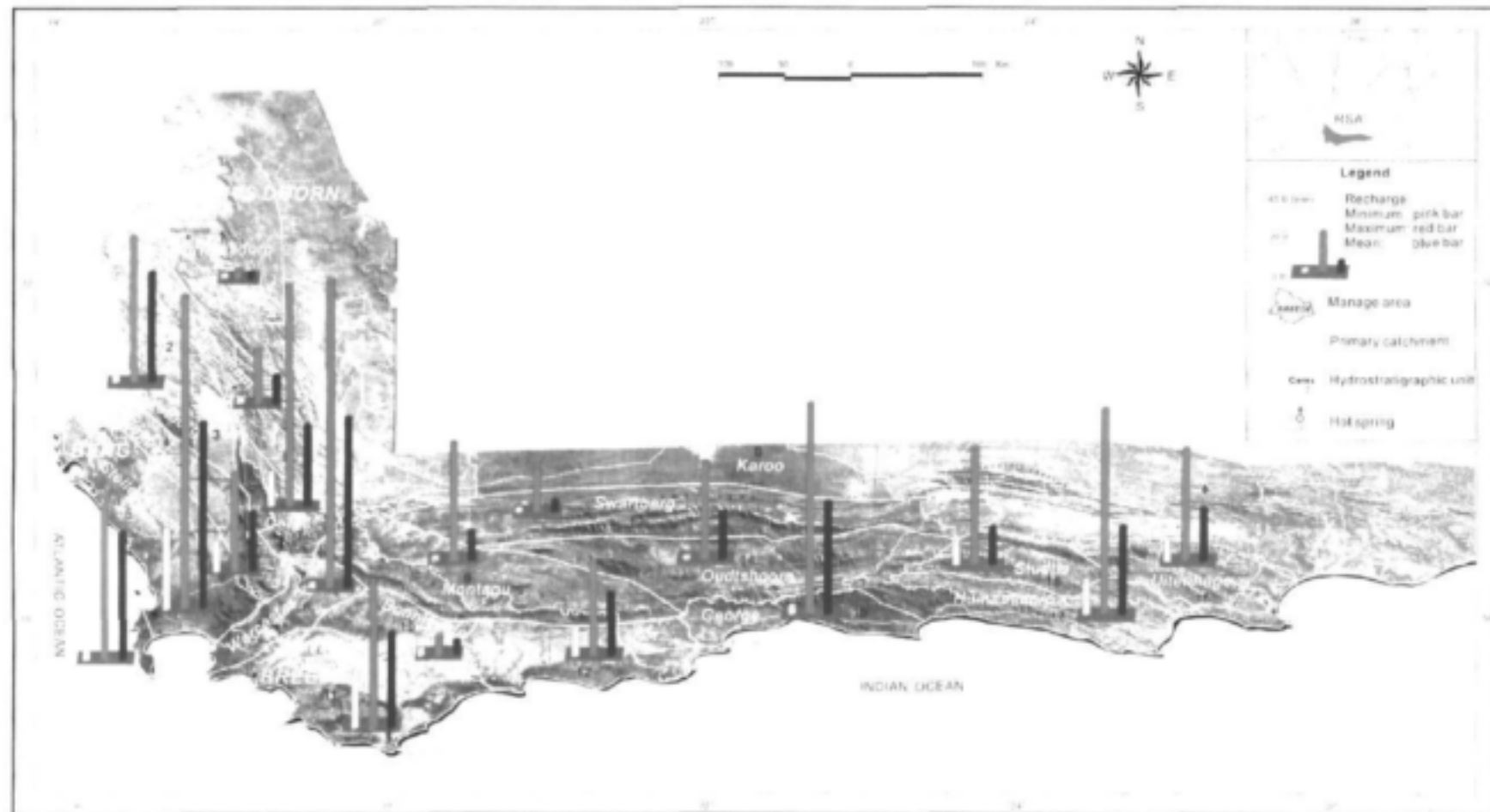


Figure 6.9 Recharge ranges in the hydrogeological units of the TMG

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

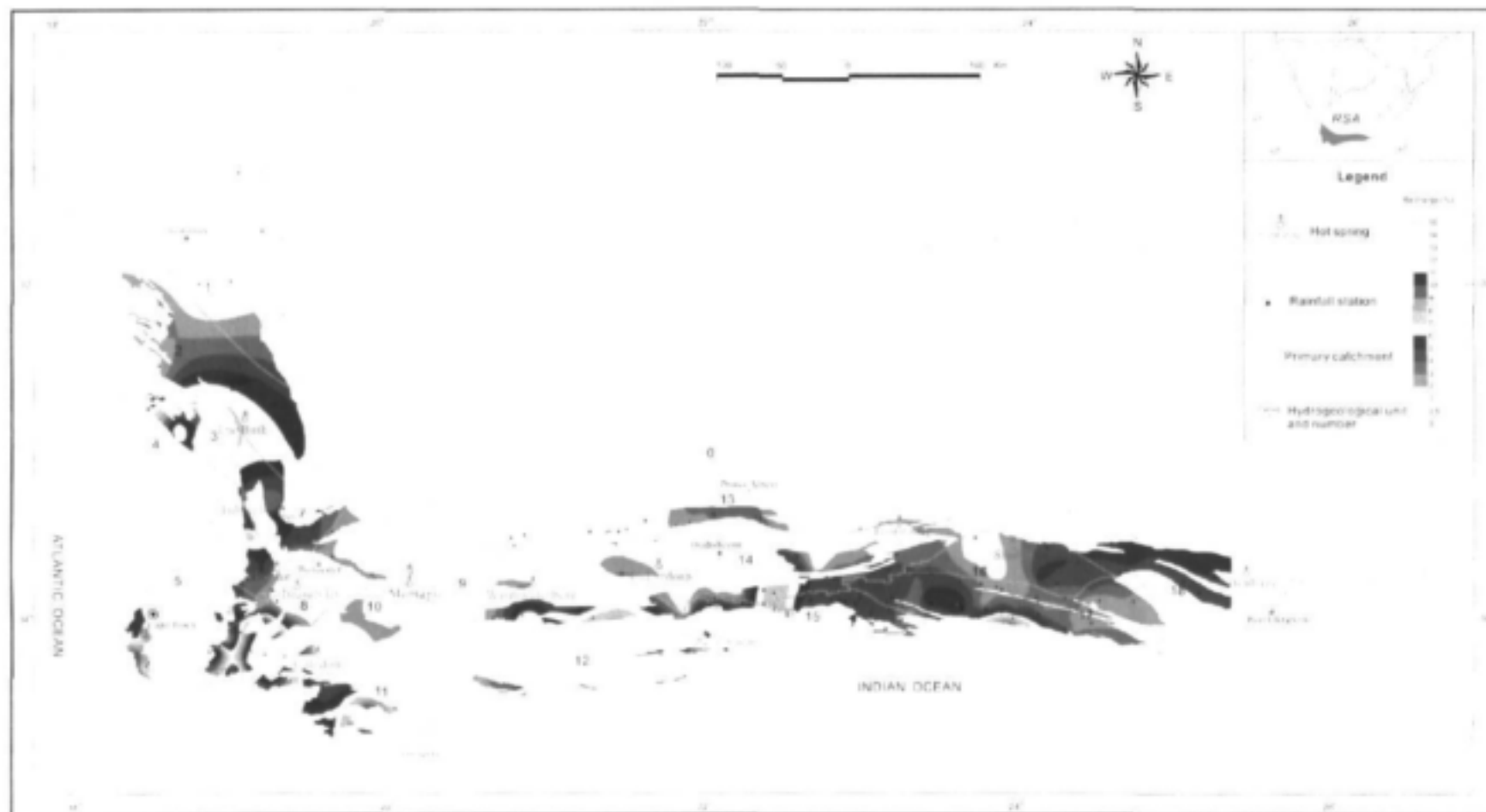


Figure 6.10 Annual recharge percentages in the TMG area

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

6.4.2 Relationship between the recharge rate and the MAP

In analysing the relationship between the rainfall and the recharge rates, the results are grouped into 11 rainfall ranges (Table 6.5). A large range of the recharge rate within the same rainfall interval occurred (Figure 6.11). The observations are as follows:

- The recharge rates from 0.5 mm·a⁻¹ to 14.5 mm·a⁻¹ are related to the MAP of 104.9 mm-298.5 mm. The recharge rates range from 2.66 mm·a⁻¹ to 32.94 mm·a⁻¹ in the MAP of 300 mm-400 mm.
- A similar range of percentages occurs in the MAP of 400 mm – 700 mm. The recharge rates of 7.19 - 57.91, 9.23 - 67.68 and 6.25 - 60.07 mm·a⁻¹ are associated with the MAP of 400 - 500, 500 - 600 and 600 - 700 mm, respectively.
- The recharge rates varying from 19.83 mm·a⁻¹ to 69.25 mm·a⁻¹ relate to the MAP of 700 mm - 800 mm·a⁻¹
- Recharge rates of 35.03 - 96.52 mm·a⁻¹ is related to the MAP of 800 - 900 mm, but higher recharge rates of 19.02 - 115.69 mm·a⁻¹ are related to the rainfall of 900 - 1000 mm·a⁻¹.
- A similar recharge rate range is also for the MAP higher than 1200 mm. The recharge rates of 35.00 - 135.2, 45.69 - 130.03 and 109.91 - 137.42 mm·a⁻¹ are related to the MAP of 1000 - 1200, 1200 - 1600 and higher 1600 mm, respectively.

Table 6.5 Ranges of the recharge estimates related to the precipitation in the TMG area

Range of P (mm)	Recharge rate (mm)			Precipitation (mm)			Station size
	Minimum	Maximum	Mean	Minimum	Maximum	Mean	
<300	0.5	14.25	3.64	104.9	298.5	232	41
300-400	2.66	32.94	14.7	301.7	392.1	355.8	14
400-500	7.19	57.91	24.67	404.1	495.6	444.9	36
500-600	9.23	67.68	31.45	501.5	593.5	542.1	21
600-700	6.25	60.07	36.27	602.3	682.8	640.8	14
700-800	19.83	69.25	37.16	704.2	792.5	752.8	12
800-900	35.03	96.52	55.56	806.8	892.9	844.4	11
900-1000	19.02	115.69	59.77	922.7	999.0	964.1	8
1000-1200	35.00	135.52	65.97	1002.7	1126.3	1085.7	12
1200-1600	45.69	130.03	93.01	1208.4	1604.8	1420.5	5
>1600	109.91	137.42	125.36	1842.1	2063	1975.2	3

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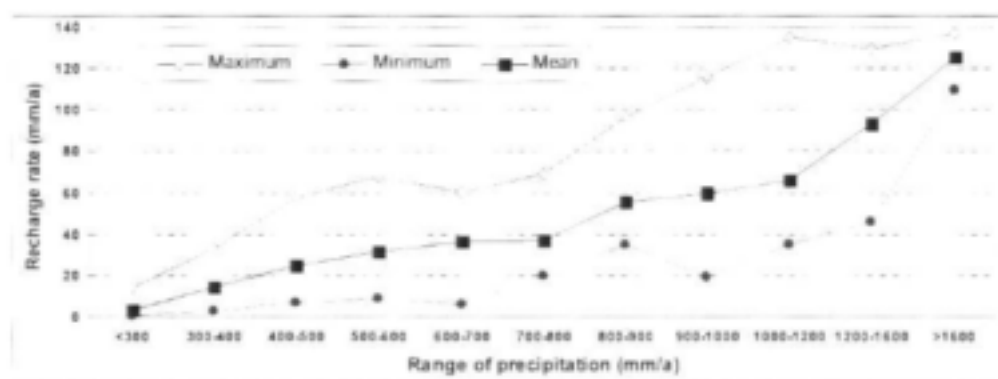


Figure 6.11 Relationship between the recharge range and the MAP range

The recharge rates appear to have non-linear positive relationship with the MAP through the scatter diagram (Figure 6.12). Most high recharge percentages are related to ranges of 300 mm to 1100 mm rainfall (Figure 6.13). The percentage is low if the MAP is greater than 1100 mm, while the low recharge percentages related to rainfall less than 300 mm could be attribute to high evapotranspiration and low effective infiltration. Actually, the recharge rates are spatially related.

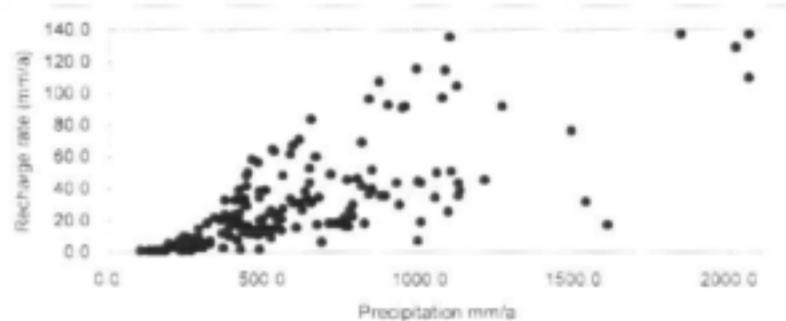


Figure 6.12 Scatter diagram of the recharge rate and the MAP

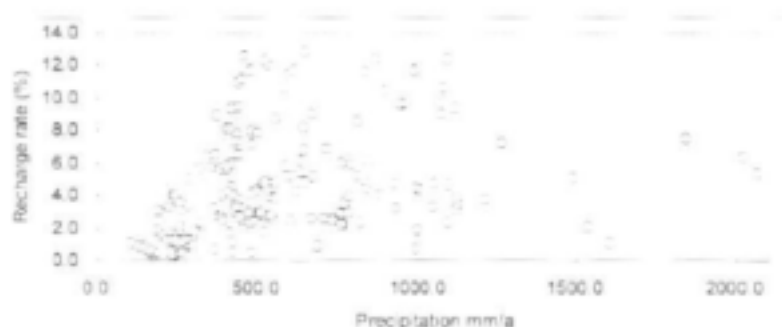


Figure 6.13 Scatter diagram of the recharge percentages and the MAP

6.5 Discussion

6.5.1 Accuracy of the model

Uncertainties and inaccuracies of the model arise from several sources: spatial and temporal variability in processes and parameter values, measurement errors, and the validity of assumptions upon which different methods are based (Healy and Cook, 2002). It is not possible to evaluate the accuracy of $R_E(t)$ estimated with this procedure. However, short-term $R_E(t)$ may be estimated from measured $P(t)$ and simple estimates of $E(t)$ and $R(t)$. Even if the accuracy of $E(t)$ and $R(t)$ is inadequate, they still contain useful information regarding the relative values of individual elements within the distributions. Ideally, recharge estimates should be presented along with statistical confidence levels or as a range of likely values (Healy and Cook, 2002).

The accuracy of recharge estimates based on the water balance approach is controlled by many factors, such as temperature, soils, vegetation and rainfall intensity. In the case of the Kammanassie, the results are calibrated with the critical value of annual recharge range of 10.88-16.09 mm (1.65-3.30%) derived from the numerical model in Chapter 5. The recharge rates based on the water balance, the CWD-CRD, the regression of the cumulative spring flux and the mixing model methods are 2.32-3.27%, 0.24-7.56%, 3.29-4.01% and 1.18-2.03% of the precipitation in the Vermaak's area, respectively. These results reflect identical levels in recharge rate, therefore, the recharge rates derived from the evapotranspiration and the runoff model seems to be reliable.

6.5.2 Sensitivity analysis

6.5.2.1 Temperature

The evapotranspiration model is dependent on temperature. The maximum and minimum temperatures are considered. Maximum temperatures range from 17°C to 25.5°C in July and from 24°C to 42°C in December, respectively. The high temperatures occur in the Buffelsklip, Wildalsvlei, Kammanassie and Montagu-Oudtshoorn but low temperature in the Uniondale and Cape Town areas in the same season. Minimum temperatures vary from -4°C to 8.8 °C in July and from 2.5 to 18°C in February. There are similar low levels of temperature in Buffelsklip, Wildalsvlei Kammanassie, Wildebeesvlakte, Elandsvlakte and Paardevlakte area, but similar high levels of the temperature in the Uniondale, Clanwilliam, Langebaanweg, Cape Town, Greyton, Montagu-Oudtshoorn, Vrolijkheid and George areas in the same season. The temperature pattern in Port Elizabeth is different from the others.

The recharge rate decreases sharply with increasing monthly average maximum temperatures, while it increases gently with the increasing monthly average minimum temperatures. It is the relative difference in temperature that causes effect on evapotranspiration. The large difference in temperature leads to more evapotranspiration. Approximately 0.01% of the precipitation should be reduced while the difference in temperature increases by 0.2°C in the Wildebeesvlakte area.

6.5.2.2 Other factors

The recharge rate is negative in relation to factors of soil, vegetation cover and intensity of precipitation. The degrees of the sensitivity of the model related to the factors are listed in Table 6.6. Under similar landscape and hydrogeological settings, the recharge rate reduces about 0.62% if the soil factor increases by 0.01; the recharge reduces by 0.31% when the coverage rate of vegetation increases by 1%; the recharge reduces 0.09% while the rainfall duration increases by 1 hour in a single month.

Table 6.6 Sensitivity of the factors

Factor	T maximum (°C)	T min (°C)	Soil Factor No dimension	Vegetation Factor No dimension
Delta	0.1	0.1	0.01	0.01
$\delta Re\%$	0.02	-0.31	-0.62	-0.31

The errors of the model in the Wildebeesvlakte are listed in Table 6.7. The extremely high recharge rate is 5.46% of the precipitation, and the extremely low rate is 1.05% of the precipitation.

Table 6.7 Estimated errors of estimates in the Wildebeesvlakte rainfall station

Condition of recharge	Delta maximum (T °C)	Delta minimum (T °C)	Soil (No dim.)	Vegetation (No dim)	Rainfall time (hrs)	$\delta\%$	% of the Precipitation	RE (mm)
Extremely low	0.25	-0.25	0.26	0.46	15	-2.21	1.06	5.12
Extremely high	-0.25	0.25	0.24	0.44	5	1.19	5.46	26.41
Model	0	0	0.25	0.45	10	-	3.27	15.80

6.5.3 Snow impacts on recharge in the TMG area

6.5.3.1 Spatial distribution of snow

The distribution of snow is associated with topography, season and temperature in the TMG area. Winters in the mountain areas are cold and wet, night temperatures drop sharply and heavy frost and snow may occur. The distribution of snow occurs above 1 000 to 1 200 m a.m.s.l of the TMG mountain area in some winter seasons. The distribution of snow in the TMG area is shown on the Landsat 7 Enhanced Thematic Mapper (ETM) images as seen in Figure 6.14 (Wu and Xu, 2005). Different distributions occurred in different seasons or periods, and snow cover remained on the gentle slope sides, e.g., the northwest slope of the Hex River Mountain and the northern slope of the Swartberg Mountain. This was also related to wind field and direction. As seen in Figure 6.14A, snow remains slightly longer on the mountain peaks due to snow thickness. On 8 August

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2002, a wide distribution of snow occurred from the Gydoberg in the north, the Hex River Mountain in the middle to the Du Toitsberg Mountains in the south as can be seen in Figure 6.14B. A small area of snow cover distributes on the peaks of the Kammanassie Mountain and Marloth as shown in Figure 6.14C.

The thickness of snow was up to 600 mm in the Langeberg at the Protea Farm and Marloth Mountain, as can be seen in Figure 6.15 (Wu and Xu, 2005), and about 300 mm of snow occurred on the Villiersdorp Mountain in August 2003 as can be seen Figure 6.16 (Wu and Xu, 2005). The more grass was present; the less snow cover was present. Less snow occurred on the side sheltered from the wind. The interflow from snowmelt could be observed. This means percolation rate was smaller than the snowmelt rate. Based on ETM images and climatic analysis, the snow-influenced recharge areas in the TMG are shown in Figure 6.17 (Wu and Xu, 2005). The influenced areas are much wider than the snow cover area. Snow is distributed widely on the mountainous peaks and places of high elevation, but snowmelt can influence the entire catchments or outcrops of the TMG. Therefore, recharge from snowmelt must be considered in these mountainous catchments.



Figure 6.14 ETM images of snow (blue) on the mountain peaks in the TMG

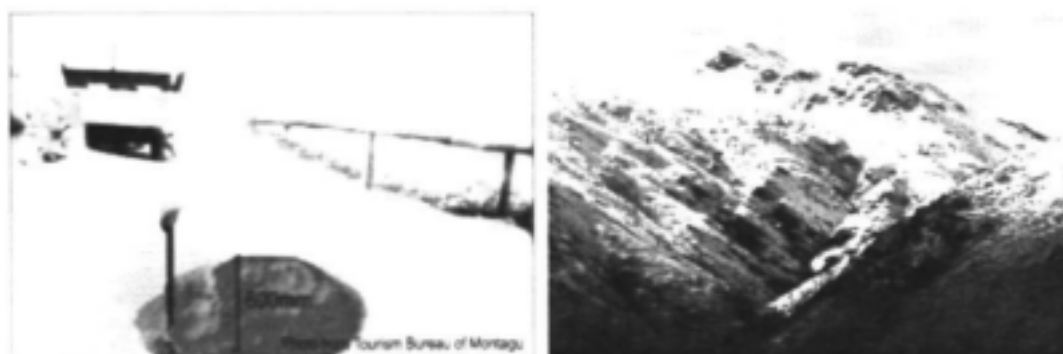


Figure 6.15 Snow on the Langeberg of the Protea Farm (left) and peaks above Franschoek Pass (right) in 2000



Figure 6.16 Snow and interflow (right) from the snowmelt in the Villiersdorp Mountain

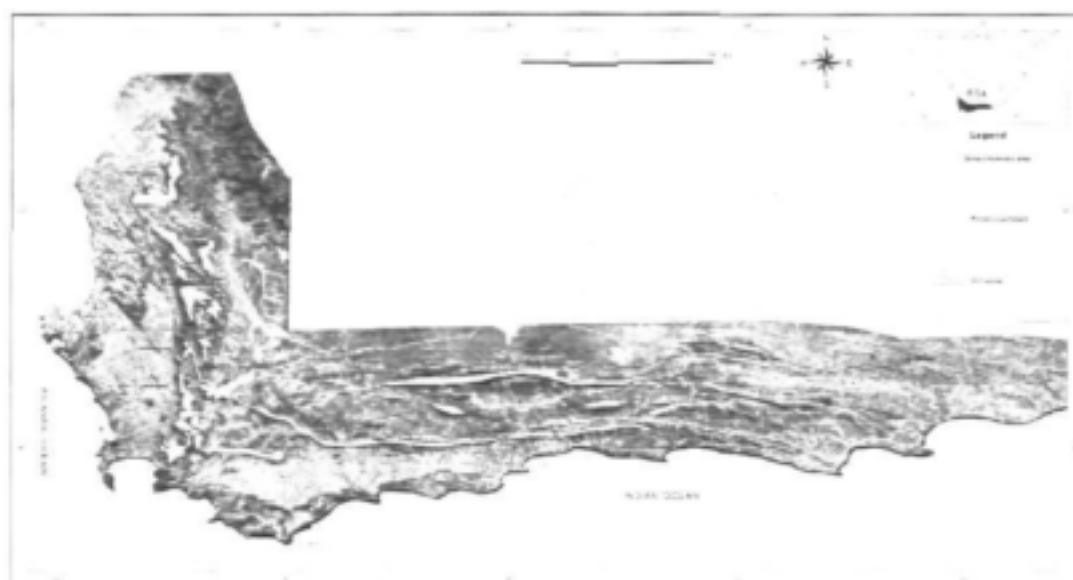


Figure 6.17 Snow influenced area in the TMG area

6.5.3.2 Impact of snowmelt on recharge

Physical processes within the snow pack involve mass and energy balances as well as heat and mass transport. Snowmelt is an energy-driven process. Recharge processes are associated with fractures, soils and vegetation, especially characteristics of fractures at depth. Groundwater recharge from snowmelt is related not only to snowmelt mechanism and processes but also to recharge processes. Snow water equivalent is used to characterize the snow pack and snowmelt rate. An infiltration experiment is used to compute the percolation rate.

Snowmelt was computed as follows, using a variable degree-day melt factor (γ) determined as a function of snow pack density (ρ_s) and vegetation cover (open/forested) (Kuusisto, 1984):

$$\gamma = 0.0104\rho_s - 0.70, (1.4 < \gamma < 3.5 \text{ in forested areas (mmd}^{-1}\text{)}) \quad (6.9)$$

$$\gamma = 0.0196\rho_s - 2.39, (1.5 < \gamma < 5.5 \text{ in open areas (mmd}^{-1}\text{)}) \quad (6.10)$$

New snowfall density ρ_{fall} was assumed to be a function of air temperature (T_{air}), and was calculated as follows (Hedstrom and Pomeroy, 1998):

$$\rho_{\text{fall}} = 67.9 + 51.3 \exp(T_{\text{air}} / 2.6) \quad T_{\text{air}} \leq 0^\circ\text{C} \quad (6.11)$$

$$\rho_{\text{fall}} = 119.2 + 20.0T_{\text{air}} \quad T_{\text{air}} \geq 0^\circ\text{C} \quad (6.12)$$

The hourly melt was computed using a standard temperature index approach as follows (Ross *et al.*, 2003):

$$\omega = \frac{\gamma}{24} (T_{\text{air}} - T_{\text{melt}}) \quad (6.13)$$

where ω is the hourly melt rate, T_{melt} is the threshold air temperature for snowmelt. There is considerable variability in published values of T_{melt} and a value of less than 0°C is not physically unrealistic since radiation melt can take place when air temperatures are below freezing (Ross *et al.*, 2003). Kuusisto (1984) obtained T_{melt} values of -1.3°C and -1.2°C for open and forested sites in Finland. These values are used for calculating hourly melt rate in the TMG mountainous area.

Based on equation (6.13), the hourly melt rates calculated at air temperatures of 4 and 6°C are 0.328 mm and 0.689 mm respectively. A higher hourly melt rate should occur in cases of higher air temperature. In the TMG mountainous area, the melt time is 6.04 d if the average snow thickness is 100 mm, which equals to rainfall of 21.9 mm. Considering infiltrating conditions, more runoff should occur if the snowmelt time is shorter than time of the equivalent rainfall, because recharge is constrained by fracture characteristics while snowmelt rate is not. Note that recharge from snowmelt is very complicated, it is related to snowmelt mechanisms and infiltrating processes. Field studies are necessary in the future.

An infiltration experiment was performed at the Franschoek Mountain Pass. The source water was an overflow of $1.107 \text{ l}\cdot\text{s}^{-1}$ from a pipe (shown in Figure 6.18). The infiltration area was about 50 m^2 and the specific infiltrating rate was $0.022 \text{ l}\cdot\text{s}^{-1}\cdot\text{m}^{-2}$. The discharge sector was about 385 m^2 . The issuing zones were the bedding fractures on the side of the steep slope. A few more seepage zones were also observed at lower elevation but with less magnitude. A small temporal

weir was used to collect the discharge water in the main discharge area. The discharge rate was measured at $0.956\text{--}0.964\text{ l}\cdot\text{s}^{-1}$, which was not total discharge water. The recharge percentage was estimated at 12.9–13.6%. Therefore, most infiltrating water seeps out as interflow. A snowmelt of $0.689\text{ mm}\cdot\text{h}^{-1}$, which equals to $1.91\times 10^{-7}\text{ m}\cdot\text{s}^{-1}\cdot\text{m}^{-2}$, would still produce much interflow or runoff. Therefore, the recharge rate must be less than that of $1.91\times 10^{-7}\text{ m}\cdot\text{s}^{-1}\cdot\text{m}^{-2}$. Suppose 10% snowmelt is available for recharge in this case, the recharge rate would be 3.59 to 7.19% of the 21.9 mm rainfall (derived from 100 mm of snow) if snow can remain 1 to 2 d after the snow event. It is worth mentioning that infiltration or recharge cannot occur in whole ground surfaces but only in open fractures or fracture zones (Wu and Xu, 2005).



Figure 6.18 Infiltration experiment

7. RAINFALL VARIABILITY AND ITS IMPACT ON GROUNDWATER LEVEL FLUCTUATIONS

7.1 *Climate Variability and Climate Change*

Our climate has never been stable for any extended period of time. Natural external causes of climate variability include variations in the amount of energy emitted by the Sun, changes in the distance between the Earth and the Sun, and the presence of volcanic pollution in the upper atmosphere. Internal variations of the climate system also produce fluctuations, through the feedback processes that connect various components of the climate system. These variations arise when the more rapidly varying atmospheric conditions 'force' the slow components of the system like internal variations in the ocean, cryosphere, or biosphere. They are intensified by coupling of the components of the system that would not have such an effect on an individual basis (Appleton, 2003).

7.1.1 *Climate Variability in Southwestern Cape (SWC)*

The Southwestern Cape (SWC) region of South Africa is characterised by winter rainfall brought mainly via cold fronts and by substantial interannual variability. About 60% to 70% of the annual rainfall over the Southwestern Cape occurs during May – September. Previous work has found evidence that the interannual variability in SWC winter rainfall may be related to sea surface temperature (SST) in the South Atlantic Ocean and to large-scale ocean-atmosphere interaction in this region. Observations indicate that a warm-cold-warm sea surface temperature (SST) pattern across the subtropical-midlatitude South Atlantic is a robust feature associated with many wet Southwestern Cape winters (Reason and Jagadheesha, 2005).

Predictions based on results from international climatic change research indicate considerable impacts on climatic patterns of southern Africa. As a result significant changes in the rainfall distribution can also be expected over the next few decades. Climate change during recent geological times over southern Africa, and in fact the rest of the world, is reflected in the geological record of continents. The impact of these on groundwater conditions during recent geological times is, however, not known (Meyer, 2005).

The Department of Water Affairs and Forestry (DWAF) monitors groundwater levels across the country with a monitoring network of between 300 and 400 boreholes but this has only started earnestly a few decades ago. An analysis of long-term groundwater level records was recently made on selected boreholes across the country by Meyer (2005) for Water Research Commission (WRC) with the following objectives:

- To determine whether long term trends or cycles can be observed in SA groundwater level data and whether such trends, if present, can be linked to climatic patterns or whether other factors are contributing to observed trends;

- Collate information that can be used to correlate long-term water level fluctuations with rainfall, drought and climatic cycles, and document or develop methods to analyse this information; and
- Document analytical techniques available and used to analyse relevant water level information to assist planners in the longer term planning of groundwater utilization as part of the Integrated Water Resources Management programme.

Ultimately the aim of the project was to derive at indicators that can be used in for example groundwater recharge estimations and the long-term management of the important national groundwater resources (Meyer, 2005).

Meyer (2005) reports that several authors have published information relating to the climatic conditions in southern Africa over the last two centuries. He summarised the findings of these authors which include Tyson et al (2002); Muller and Tyson (1988); Currie (1993); Richard et al (2001); Vogel (1989,1995); Tyson and Patridge (2000); Nash and Endfield (2002) that a number of different climatic cycles of different duration appear to have been present or are postulated to have been present during the last 200 000 years (Meyer, 2005). The cycles are:

- An approximately 23 000 year cycle of major wet and dry periods during the last 200 000 years;
- A possible 400 – 500 year 'wet/dry' cycle;
- An ~120 year cycle;
- An ~18 year cycle;
- An 11 year cycle;
- A 3.5 year cycle; and
- A 2.3 year cycle.

The following records (Table 7.1) are quite significant in the southern Africa region during the last century (after Meyer, 2005).

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Table 7.1 Major wet and dry phases for Southern Africa for the 20th century.
(after Meyer, 2005).

Major wet periods	Major dry periods	Probable regional extent
	1905/06 – 1915/16	Southern Africa
1916/17 – 1924/25		Southern Africa
	1925/26 – 1932/33	Southern Africa
1933/34 – 1943/44		Southern Africa
	1944/45 – 1952/53	Southern Africa
1953/54 – 1961/62		Southern Africa
	1962/63 – 1970/71	Southern Africa
1971/72 – 1980/81		Southern Africa
	1982/83 – 1992/93	Southern Africa
1993/94 – 2002/3		Southern Africa

7.1.2 Groundwater Level Fluctuations

Records of groundwater levels in the number of boreholes that are periodically monitored by the Department of Water Affairs and Forestry (DWAF) show different water level fluctuation patterns which are broadly classified below (Meyer, 2005):

- Short term (~1 year) cyclical fluctuations in the water level data;
- Sharp increases in water level usually associated with extreme climatic events;
- Relatively steep and short term declining trend presumably associated with external factors such as abstractions;
- Medium term (~ 5-10 years) declining trends in the water level data; and
- Longer term usually declining trend which may be of a cyclic nature, but due to the short time series, it is not clear whether it does reflect a specific cycle.

A combination of the effects of abstraction and any of the periodic trends could produce a more complex trend that is sometimes difficult to classify. For the purpose of analysing groundwater records, Meyer (2005) suggested that the 18 and 11-year cycles are perhaps the most significant. In the TMG region significant correlation exists between rainfall and groundwater levels using data from a number of rainfall stations and boreholes in the Kammanassie area as discussed in section 5.2. Correlation coefficients ranged from 0.67 to 0.83, the differences in correlation attributed to a number of factors influencing water levels such as abstraction, temperature, evapotranspiration, geology, etc. The correlations come with different time lags as revealed by the regression analysis.

7.2 Application of Drought Indicators to Analyse Groundwater Fluctuations

Drought indicators have been used to predict drought throughout the world. They are derived from the analysis of rainfall, streamflow and other water supply indicators and measure how much precipitation or flow for a given period of time has deviated from historically established norms. Difficulties are primarily related to the time period over which deficits accumulate and to the connection of the deficit in precipitation to deficits in usable water sources and the impacts that ensue (McKee et al, 1993). The indices in use often are Percent normal, Palmer Drought Severity Index (PDSI), Standard Precipitation Index (SPI), Crop Moisture Index (CMI), Surface Water Supply Index (SWSI), and Reclamation Drought Index (RDI). No specific drought index has yet been developed to describe groundwater drought conditions.

7.2.1 The Standard Precipitation Index (SPI)

The understanding that a deficit of precipitation has different impacts on groundwater, reservoir storage, soil moisture, snow pack, and streamflow led McKee, Doesken, and Kleist to develop the Standardized Precipitation Index (SPI) in 1993. The SPI was designed to quantify the precipitation deficit for multiple time scales. These time scales reflect the impact of drought on the availability of the different water resources. Soil moisture conditions respond to precipitation anomalies on a relatively short scale. Groundwater, streamflow, and reservoir storage reflect the longer-term precipitation anomalies. For these reasons, McKee et al. (1993) originally calculated the SPI for 3, 6, 12, 24, and 48-month time scales based on the long-term precipitation record for a desired period (Hayes, 2002). The SPI calculation for any location is based on the long-term precipitation record for a desired period. This long-term record is fitted to a probability distribution, which is then transformed into a normal distribution so that the mean SPI for the location and desired period is zero. Positive SPI values indicate greater than median precipitation, and negative values indicate less than median precipitation. Because the SPI is normalized, wetter and drier climates can be represented in the same way, and wet periods can also be monitored using the SPI. This is the advantage that the SPI has over the other indices in analysing groundwater levels in addition to its applicability to different time scales. SPI values cover a range of approximately -3 to +3 in irregular steps. Positive SPI values indicate greater than median precipitation, and negative values indicate less than median precipitation (Table 7.2).

Table 7.2 Classification scale of SPI values

SPI Values	
2.0+	extremely wet
1.5 to 1.99	very wet
1.0 to 1.49	moderately wet
-.99 to .99	near normal
-1.0 to -1.49	moderately dry
-1.5 to -1.99	severely dry
-2 and less	extremely dry

7.2.2 Standard Precipitation Index and Groundwater Level Fluctuations

Meyer (2005) analysed South African groundwater level data using the SPI calculated for a 24-month period. In two examples where groundwater level data were correlated with the SPI a remarkable good correlation over the entire record period between the two datasets was present in both cases. Both the larger and smaller anomalies (amplitude and duration) correlate well and good correlation between groundwater level and both macro and micro changes in SPI were shown to exist. However, records of poor correlations also exist in some datasets and further study is recommended to establish the reasons for correlation whether good or poor.

In the Kammanassie region most of the water level data are so short to do any meaningful analysis using the SPI. In future when there is enough water level data from the monitoring wells it is hoped that this method could be employed to evaluate its applicability in this region.

7.2.3 Recharge Index

The good correlations observed between the SPI and groundwater level records of some of the South African datasets led to the development of a 'Recharge Index' concept (Meyer, 2005). The index is derived from the SPI values, groundwater level response and climatic indicators. Meyer (2005) believes that by using groundwater level fluctuations as an indication of recharge should result in a measure of the 'effective recharge' and that the development of a Recharge Index using the SPI approach will combine the effect of a number of variables and reduce the uncertainty in recharge estimates. Meyer (2005) put forward a number of factors to be taken into consideration in developing the SPI Recharge Index and proposes a way to rate the groundwater recharge though he acknowledges that considerable amount of work needs to be done to make the methodology more quantitative. He concludes that the method of developing a 'Recharge Index' derived from SPI data may not provide the required quantitative assessment of groundwater recharge, but together with other methods available may add the required information to provide more quantitative answers that can be to the benefit of the water resource planners in future.

8. CONCLUSIONS AND RECOMMENDATIONS

The TMG aquifer comprises approximate 4000 m thick sequence of sandstone with an outcrop area of 37,000 km² as key recharge areas. A wealth of recharge estimation methods for the TMG area is currently available with each method having its own limitations regarding temporal and spatial scale. Recharge estimates out of this study are noticeably much conservative than what are postulated in groundwater community. Confidence in the obtained recharge estimates in the TMG area has been improved by applying multiple methods, which included the CMB, CWD-CRD, SVF and water balance methods. The Table Mountain Group (TMG) is a regional fractured rock aquifer system with the potential to become important sources of future water supplies to meet both domestic, agricultural and urban requirements in the Western and Eastern Cape Provinces in South Africa.

8.1 Conceptual Hydrogeological Model

Higher fracture frequencies of the TMG during deformation forms secondary porosity. The storativity preferred is from 0.1 to 0.001. From the occurrence of aquifers, five types of aquifer systems, namely the horizontal terrain aquifer system, incline strata aquifer system, folded strata aquifer system, fractured rock aquifer system and weathered rock (crust) aquifer system, are classified. Based on geological and hydrodynamic characteristics, namely, geomorphology and boundary conditions of groundwater flow, nineteen hydrogeological units in the TMG area are identified, covering the area of 248,000 km².

The recharge processes are related to topography and fracture networks. Different topography types show different patterns of surface water runoff and infiltration mechanisms. Interflow occurs in the weathering and preferential fracture zones above the elevation of the local basal levels. Infiltration rate is related to characteristics of land surface; and the percolation rate to the aquifer is constrained not only by the infiltration rate but also by fracture characteristics at depths. In most cases, the percolation rate is very small as most of infiltration water seeps out in the form of interflow, especially in mountainous areas. The preferential recharge is dominated in the fractured TMG rocks. The recharge possesses diffuse pattern on the land surface with high fracture density, but preferential recharge occurs at depth. Six models of the recharge processes related to the landform were conceptualised, namely models of soil zones, horizontal strata zones, upright strata zones and three incline strata zones.

The hypothesis that the recharge of the TMG aquifer probably takes place within the outcrop area of the TMG was confirmed by analysis of the Landsat 7 ETM images. For local aquifer system, the Cedarberg shale is obviously the recharge boundary. More effective recharge occurs with the slope gradient less than 30°. The soil deposited on the bottom of the valley affects recharge water percolating to the fractured rock aquifer at depth. Part of the fault zone of some faults cemented by silicate minerals results in complicated recharge processes, in which low or no

infiltration occur in cemented zone, but high percolation concentrates in non cemented zones, such as Vermaak's Fault zone.

Groundwater in the Peninsula aquifer is characterised by its low TDS. The ranges of average TDS in the production boreholes of the Peninsula and Nardouw aquifers in the Kammanassie area are 40-60.4 and 86.8 $\text{mg}\cdot\text{L}^{-1}$, respectively. The groundwater type of the TMG aquifer is classified as Cl-Na type of water, as all other major ions are in very low concentrations. The source of the chloride in the TMG aquifer is from precipitation fallout, little additional chloride adds from rocks or formations. Based on the 98 samples distributed in the TMG area, the frequency is 32.7%, 24.5%, 15.3% for concentrations of 10-20 $\text{mg}\cdot\text{L}^{-1}$, 20-30 $\text{mg}\cdot\text{L}^{-1}$ and 30-40 $\text{mg}\cdot\text{L}^{-1}$, respectively. The frequency is 8.2% in both 50-60 $\text{mg}\cdot\text{L}^{-1}$ and 60-70 $\text{mg}\cdot\text{L}^{-1}$. The errors between arithmetic and harmonic, arithmetic and geometrical are 38%, and 18%, respectively. Therefore different errors could be introduced in the recharge estimation using different mean values. The work initiated by Meyer (2005) should continue and be applied in the TMG.

8.2 Recharge Related to Rainfall Pattern

Recharge is significantly influenced by seasonal variation of rainfall and extreme rainfall event. A gradual positive trend of rainfall occurred in the Hoekplaas and the Kammanassie rainfall stations in 1925-2002, but the one-tailed probabilities are low. There was a positive tendency in January, April, June, October, November and December, and a negative tendency in the other months in the Kammanassie Dam rainfall station. Negative CRD patterns occurred in the mountainous area from July 1997 to May 2002. Positive trend of rainfall does not mean positive recharge trend and neither does the seasonal trend. The variation of monthly rainfall results in more complicated water level response.

The same patterns of CRD are observed in the four rainfall stations, namely Parshall, V-notch, Wildebeesvlakte and Purification Works East located on the Kammanassie Mountainous area. Negative patterns of the CRD occur from July of 1997 to May of 2002. A significant correlation occurs in the CRD between the Parshall and Wildebeesvlakte stations. The extremely high rainfall events contribute to the effective recharge. The effective recharge may mainly occur in March, April and November in the Kammanassie mountainous area.

There is evidence to believe that correlation exists between long-term rainfall pattern and some groundwater datasets in southern Africa. A number of different climatic cycles of different duration appear to have been present or postulated during the last 200 000 years. For the purpose of analysing groundwater records the 18-year and 11-year cycles are the most significant. Analysis has shown that using the Standard Precipitation Index and other geological and hydrogeological factors a recharge index could be developed which could reduce the uncertainties in recharge estimates. The suggested methodology though subjective could be combined with other known recharge methodologies to provide more quantitative answers that can be to the benefit of water resource planners.

8.3 Recharge Estimates in the Kammanassie Area

The recharge rates in the Kammanassie region from 1994 to 2002 using the three-component mixing model of the CMB, CWD-CRD, regression of cumulative spring flux and water balance methods are estimated at 1.182 - 2.027%, 0.24 - 7.56%, 3.29 - 4.01% and 2.32 - 4.4 % of MAP (4,217 mm), respectively. The influence area of the abstraction and the concentrations of end-member input largely influence the recharge rates with the mixing model of chloride mass balance. Storativity is a very sensitive parameter for recharge estimation using the CWD-CRD method; and the actual recharge area is a very important factor in the spring flux regression method. The recharge estimates were integrated using weights based on the influences of the various methods in the study area. The integrated rates computed showed that about 1.63% - 4.75% of the MAP was recharging the aquifers in the Kammanassie area.

The recharge rate in the Kammanassie area is generally lower than 5%, which is much smaller than that presented in the earlier studies. The budget from the 3-D numeric simulation model confirmed that the recharge rate could not exceed 1.96% of the rainfall. The recharge estimates for the Vermaaks River wellfield area are compared with those from other areas in southern Africa in Figure 8.1. It is interesting to note that the recharge of the Vermaaks River wellfield is compatible with that of Botswana and the estimates have relatively higher values than those in Botswana. This may be attributed to the fact that highly fractured TMG formations and less evapotranspiration are present in the Vermaaks region.

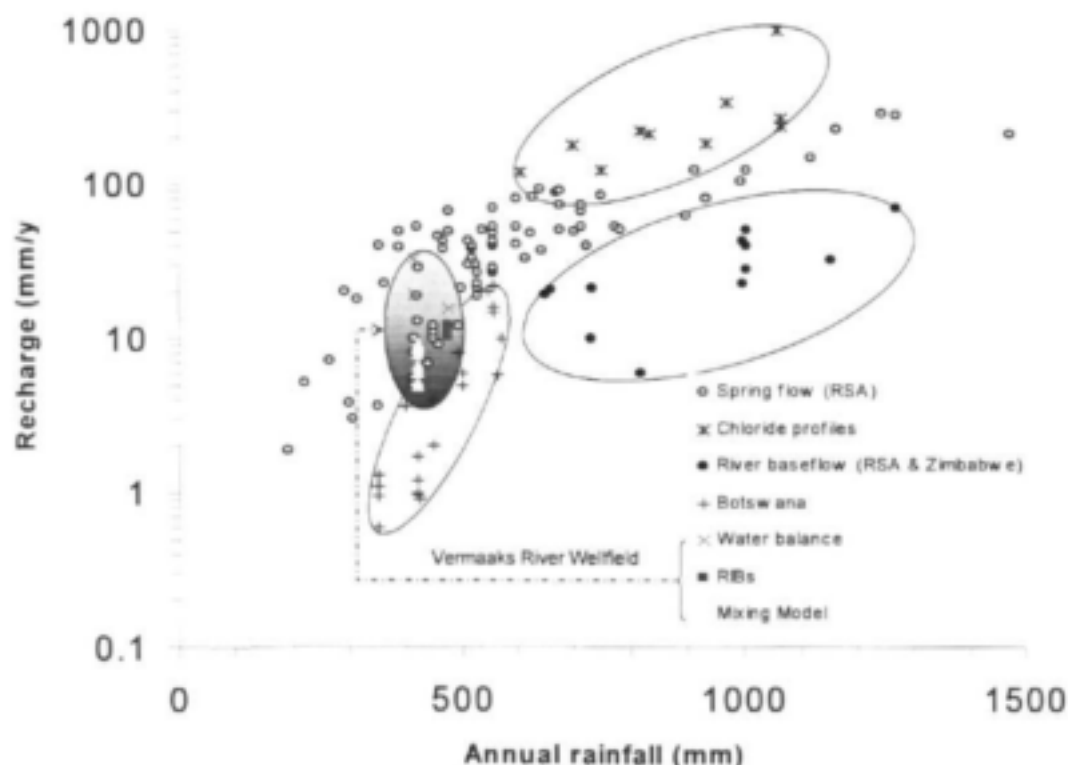


Figure 8.1 Recharge estimates in Kammanassie in comparison with estimates in Southern Africa.

It must be pointed out that the estimates excluded the likely recharge to deep flow, e.g. Calitzdorp hot spring.

8.4 Spatial Distribution of Recharge in the TMG Area

The spatial distribution of the recharge in the TMG outcrop area was presented based on the water budget models. The recharge rates from the water budget models in the Kammanassie area were calibrated with the results of the CWD, the SVF and the CMB mixing model. The impact of snowmelt processes on groundwater recharge was also discussed using results from an infiltration experiment and the snow distribution from the ETM images.

The average recharge rate in the TMG is $29.74 \text{ mm}\cdot\text{a}^{-1}$, which is generally smaller than the values estimated by Vegter (1995). The highest recharge rate of $137.17 \text{ mm}\cdot\text{a}^{-1}$ is related to the MAP of $1842.06 \text{ mm}\cdot\text{a}^{-1}$ at station No. 0022116 in the hydrogeological unit 5; the lowest recharge rate of $0.7 \text{ mm}\cdot\text{a}^{-1}$ relates to the MAP of $163.51 \text{ mm}\cdot\text{a}^{-1}$ at station No. 0048043 in the hydrogeological unit 13. The regions with a recharge rate more than $35 \text{ mm}\cdot\text{a}^{-1}$ occur in the hydrogeological units 5, 6, 8 and part of the hydrogeological unit 17. The recharge rate in the other areas is less than $35 \text{ mm}\cdot\text{a}^{-1}$. The recharge percentages are from 0.28% of the MAP at station No. 0107169 in the hydrogeological unit 1 to 12.6% of the MAP at station No. 0012220 in the hydrogeological unit 15.

The recharge rates based on the water balance method have a non-linear positive relationship with the MAP. The recharge rates of 7.19 - 57.91, 9.23 - 67.68, 6.25 - 60.07, 19.83 - 69.25, 35.03 - 96.52, 19.02 - 115.69, and 17.45 - 137.42 $\text{mm}\cdot\text{a}^{-1}$ are associated with the MAP of 400 - 500, 500 - 600, 600 - 700, 700 - 800, 800 - 900, 900 - 1,000 and 1000 mm upwards respectively. Most of the high recharge percentages are related to rainfall ranges between 300 - 1100 mm. The low percentages occur in the MAP greater than 1100 mm because more runoff occurs, and the low recharge percentages related to rainfall smaller than 300 mm can be attributed to high evapotranspiration and low effective infiltration.

8.5 Suggestions and Recommendations

The complexities of the hydrogeological settings make estimating local or regional recharge difficult. The spatial variation of rainfall and topographic conditions result in complex recharge mechanisms. Not all of the components of the mixing model can be measured accurately within a given catchment. Inputs such as precipitation and abstraction are quantifiable; however, the time lag changes significantly in relation to the residence time of the water. The solutions of the three-component model rely on the results from the two-component model, even the mixing model cannot include exponential flow, which occurs likely in the TMG system. The size of the recharge area is not easy to determine accurately in a fractured rock aquifer. Under these limitations, estimates of recharge from the mixing model are subject to large potential errors. In general, the mixing model of the CMB is unfit to be applied if recharge processes and recharge areas are not identified correctly.

The validity of assumptions of the methods is the basis of the actual recharge estimates. Errors would be introduced if different parameters were adopted in the empirical model. Uncertainties and inaccuracies arise from spatial and temporal variability of the data. The recharge rate, based on empirical factors in the water balance model, should be cross-checked with other methods before it is used for assessment of groundwater resources in the whole TMG area. Estimation of actual evapotranspiration should be cross-checked by application of remote sensing techniques (satellite imagery).

In the Vermaak's River Wellfield, the abstraction data are the touchstone for realistic estimates of the Peninsula fractured rock aquifer. Any recharge estimates obtained prior to actual groundwater abstraction are likely to be inaccurate. In TMG areas, all types of discharges should be considered together when recharge estimation is made with the methods related to discharge flux. Therefore, the spring flow, together with interflow within the aquifer system should also be considered for a total recharge. Low recharge rates estimated for the TMG area suggests that caution should be taken when a water supply strategy is formulated for a particular locality. The design of the borehole sustainable yield requires consideration of the recharge aspects as well as regional climatic change.

Future research should take into account snowmelt impact and types of vegetation. As the distribution of snow occurs above 1000 m-1200 m a.m.s.l of the TMG mountain area in some winter seasons, the recharge areas influenced by snowmelt are the catchment areas above 1000 m a.m.s.l.

Three types of vegetation, namely the Mountain Fynbos shrub, sparse vegetation and exotic plants, are dominant in the recharge areas of the TMG aquifer. The most common vegetation is the Fynbos shrub. The exotic plants consume more water than the indigenous plants and they could also reduce recharge. Several studies focusing on the impact of the exotic plant species on runoff suggest that the study of their impact on recharge would be necessary in the future.

Another shortcoming of the project is the recharge to deep flows including 12 hot springs (12 Mm³/a) and possibly groundwater discharge to oceans. These aspects also need further attention (Netili, 2007).

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10. APPENDICES

Appendix I: **Summary of the production boreholes data in the Vermaak's area**

Borehole	Depth (m)	Water Level (m)		Water Strike (m)*	Depth of pump (m)	Screen Depth (m)
		First	Now			
VG3	206.7	6.02	10	110-111(6.0), 190(3.0), 174 (3.0)	148	96.5-206.7
VR6	250	34.64	60	228-244(15)	165	108.7-230
VR7	177	63.3	90	78-81(8), 129-140(15)	159	53-177
VR8	251.3	100.5	125.4	113-117(5) 156-170(4) 234-240(4)	163	89.6-251.3
VR11	224.5	125.5	151	139(2) 183-194(8) 200-210(10)	180	18-224.5
DP10	210	114.07	90.2	183(7)	180	73-210
DP12	192	126.07	102.8	?(20)	180	66-192
DP29	240	120.6	97.09	160-170(2) 185-?(2)		
DP28	246	117.8	94.57	122-124(1.5) 151-160(10) 195-210(11)	170	121-207
DP15	224.5	103.8	86.04	110(3) 169(7) 187(11)	180	50-207
DP25	203	104.9	83.5	109, 166, 201	170	9-203
DG110	212	110.6	107.5 7	114-117(1.5) 137 200-203(6)	200	92-212
DP18	17	3.6	3.2	4.2-9(15)	14	2-9.4

The first water strike (fracture zone) depth (m), followed by yield, e.g. 110 – 111(6.0), after Kotze, 2002.

Appendix II: **Information of unsuccessful production boreholes of the KKRWSS**

Borehole	Depth (m)	Yield (l-s ⁻¹)	Geology	Water level (m)
WN101	243.5	>2.0	Baviaanskloof Quartzite, shale at 28m 188, 203 m, fractures at 120 m.	4.3 (1989)
VR5	215	15.0	0-12 boulders; 12-215 Baviaanskloof, shale at 37, 46, 57, 111 m, fractures at 82, 102, 169 and 184-186 m. (75-215).	4.3 (1989)
VG12	173	4.7	0-4 boulders; 4-230 Tchando Formation, fractures at 19, 78, 98, 14 and 171-173 (40-173)	2.93 (1992)
VG4	113	<2.0	0-2 scree, 2-113 Baviaanskloof	18.2 (1989)
DG107	210	4.1	0-22 boulders, 20-210 Kouga Formation	27 (1987)
DG104	250	5.7	0-8 Enon, 8-250 Kouga Formation	90 (1987)
DP27	249	135-140 (2) 140-195 (10) 240(8)	0-22 Enon 22-249 Baviaanskloof Formation, Fractures at 150 to 155 m, open joints with showing weathering at 241, 242 and 248 m	119 (1992)
DP20	220	0.9	0-18 Enon, 18-220 Baviaanskloof Formation	104.5 (1991)
DP14	167	12	0-6 Enon, 2-30 weathered sandstone, 30 to 167 sandstone	62.5 (1986)
DP13	184.6	30	Baviaanskloof with shale	112 (1990)

Appendix III: **Information of some monitoring boreholes: Eastern Section**

Borehole	Depth (m)	Depth of Water Strike#	Geology (depth in m)	Water Level (m)	Altitude (m)
G40171	50 (34-40)	?	0-17 boulders 17-50 Peninsula quartzite	5.085	660.516
G40172	16 (0-16)	(5.0)	Boulders	5.125	660.416
G40173	10 (0-10)	(3.0)	0-10 sandstone boulders and fragments in yellow sand matrix	4.925	660.216
G40174	147 (129-135)	127-133 (13)	0-12 sand and scree 92-147 sandstone and shale Fracture zone: 127-133 with dark brown Fe staining	105.98	469.66
G40175	126	84(5.0)	0-23 Boulders 23-49 C/S layer 49-126 Peninsula quartzite		
G40175A	84 (57-63)	27 (11) 60(0.75)	0-19 Sand, boulders and scree 19-45 C/S layer 27-30 fractured with quartz vein 48-84 Quartzite, fractured, Fe stained, pyrite quartz veins	Artesian	652.031
G40176	150 (39-45) (116-122)		0-17 sand and boulders 17-150 sandstone and shale weathering visible up to 119 m, reddish to yellow brown stains	6.71	644.081
G40177	150	92(2.4)	0-17 Sand and boulders 17-150 Sandstone (water strike in quartz vein)	51.94	420.036
G40178	120 (51-57)	54(8.0)	0-11 weathered sandstone 11-120 sandstone 94-114 fault zone with quartz veins containing pyrite	10.22	535.691

#Water strike yield is indicated within brackets ($\ell \cdot s^{-1}$), screen, if present in italics.

Appendix IV:
**Summary of water quality in the
Vermaaks area (mg·l⁻¹)**

GROUNDWATER RECHARGE ESTIMATION OF TABLE MOUNTAIN GROUP AQUIFER SYSTEMS WITH CASE STUDIES

Name	TDS	pH	K	Na	Ca	Mg	F	Cl	HCO ₃ ⁻	NO ₃ ⁻	SO ₄ ²⁻	PO ₄ ³⁻	Si	NH ₄ -N	Mn	Fe	Source
Rainfall	136.5	6.5	6.89	1.92	20.15	5.98	nd	1.74	63.4	48.45	8.36	11.26	nd	25	0.21	0.03	1
VG3	86.8	5.48	0.76	24.08	2.18	2.98	0.13	39.40	5.1	0.89	10.40	0.01	9.75	0.04	0.44	0.0043	1,2,3
VR6	48.0	6.01	1.01	12.89	1.89	1.90	0.11	19.50	10.1	0.20	3.08	0.02	5.46	0.03	nd	0.01	1,2,3
VR7	58.3	6.30	0.75	13.29	2.40	2.20	0.15	21.65	9.7	0.31	7.49	0.02	4.78	0.05	0.0055	0.0265	1,2,3
VR8	60.4	6.19	0.89	15.68	2.14	2.58	0.25	28.26	8.0	0.11	7.44	0.05	5.08	0.01	nd	0.37	1,2,3
VR11	58.9	6.29	0.84	15.14	1.93	2.42	0.13	25.48	8.8	0.07	7.30	0.02	4.95	0.02	0.0055	0.0265	1,2,3
V-notch	79.0	6.71	1.04	23.27	2.93	2.33	0.10	32.90	13.2	0.03	7.80	0.02	10.24	0.03	0.00	0.07	3
VG16	71.4	5.50	0.80	23.63	1.60	2.30	nd	30.50	9.5	1.77	6.00	0.06	0.00	0.02	nd	0.03	1
G40171	64.0	6.59	0.59	14.30	3.80	2.10	0.08	20.60	19.3	0.32	5.60	0.03	8.78	0.01	0.001	0.136	3
G40172	64.0	6.45	0.74	15.60	2.50	2.30	0.08	21.60	14.1	0.39	8.30	0.05	8.80	0.02	0.001	0.136	3
G40173	56.0	6.31	0.74	14.60	2.20	2.00	0.07	20.60	12.3	0.17	5.60	0.02	8.06	0.00	0.157	0.248	3
G40175	52.0	5.20	0.90	13.00	3.00	2.00	0.10	23.00	6.0	0.08	8.00	0.04	9.20	0.04	0.167	0.005	3
G40175 51M	94.0	6.97	1.12	14.50	9.80	1.80	0.09	21.90	40.4	0.01	11.70	0.02	9.14	0.02	0.114	0.003	3
G40175A	57.0	6.18	0.99	13.60	2.40	2.40	0.09	19.70	7.6	0.10	10.80	0.01	9.18	0.01	0.186	0.003	3
G40175A 23M	96.0	7.06	1.06	15.20	9.70	1.60	0.10	22.50	37.8	0.01	14.80	0.02	8.92	0.02	0.139	0.306	3
G40175A 27M	78.0	6.78	1.43	13.50	3.20	2.30	0.11	22.00	31.2	0.20	8.70	0.02	8.88	0.31	0.093	0.003	3
G40175A 60M	73.0	6.71	2.55	11.90	3.50	1.50	0.08	16.90	32.3	0.01	9.80	0.01	4.54	0.02	0.309	0.003	3
G40175A 84M	67.0	6.55	1.82	12.70	3.50	2.10	0.08	21.30	15.8	0.02	12.10	0.01	8.32	0.02	0.507	0.003	3
G40175B 115M	57.0	6.20	1.08	13.80	2.00	1.60	0.07	20.90	10.6	0.01	8.60	0.02	8.86	0.03	0.17	0.003	3
G40175C 126M	60.0	6.63	1.15	13.40	2.30	1.70	0.07	20.80	13.0	0.02	9.60	0.03	9.10	0.03	0.196	0.068	3
G40175	58.0	6.37	0.80	13.70	2.90	2.40	0.09	19.10	13.6	0.22	7.00	0.03	9.26	0.04	0.013	0.003	3
G40175 42M	69.0	6.99	1.07	14.10	2.90	2.00	0.07	23.20	20.8	0.21	7.30	0.01	5.24	0.07	0.502	0.003	3
G40175A 119M	111.0	7.43	3.03	15.50	10.10	2.20	0.05	22.70	53.2	0.40	12.30	0.07	10.02	0.08	0.001	0.003	3
G40175B 150M	109.0	7.16	2.74	15.10	11.20	2.20	0.11	21.50	58.8	0.01	8.10	0.02	10.00	0.03	0.023	0.224	3
VS 1	266.8	6.90	6.51	56.56	20.88	13.49	nd	80.30	152.2	0.97	10.26	0.31	0.00	1.07	1.02	1.42	1
VS 2	172.9	6.20	1.35	55.21	2.90	6.67	nd	83.70	15.9	2.07	11.57	0.11	0.00	0.18	0.11	1.32	1
VS 3	144.4	6.10	1.03	45.97	2.66	5.15	nd	74.10	12.7	0.02	8.77	0.18	0.00	0.02	nd	0.18	1

1 This study (WRC K5/1326) 2 NGDB, 3 WRC 98 and Kotze, 2002 nd not detected

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