

**A PRELIMINARY UNDERSTANDING OF DEEP
GROUNDWATER FLOW IN THE TABLE MOUNTAIN GROUP
(TMG) AQUIFER SYSTEM**

by

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A thesis submitted in partial fulfilment of the requirements for the degree of



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ABSTRACT

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Key words:

Deep thermal groundwater, shallow groundwater, groundwater classification, hydrogeology, Table Mountain Group, Thermal springs, TopoDrive, Particle tracking

The Table Mountain Group (TMG) Aquifer is the second largest aquifer system in South Africa, after dolomites. This aquifer has the potential to be a significant source of water for the people of the Western Cape. The occurrence of hot water springs in the TMG in relation with the main geological fault systems in South Africa shows that deep flow systems do exist. Little is known about these deep aquifer systems in South Africa (i.e. flow mechanisms).

To close the above-mentioned knowledge gap, this study was initiated. The current study gives a review of some of the aspects that need to be considered when distinguishing deep groundwater from shallow groundwater. These involve the use of hydrochemistry, temperature in relation to groundwater circulation depths and the general concepts of confined and unconfined aquifers. Temperature is regarded as the most suitable dividing line between shallow and deep groundwater for this study. The use of this dividing line in the TMG is limited by the lack of data regarding the distribution of the geothermal gradients.

The modelling package TopoDrive was used to simulate groundwater flow on a steady state with the aim of assessing the influence of the Cedarberg aquitard (C/S) on the deep flow. This was achieved by means of particle tracking, and multiple simulation runs with variable (i.e. from $K=1.0E-15$ – $K= 1.0E-02$) hydraulic conductivity (K) of the C/S. Simulation results show that C/S with K value of $1.0E - 02$ can allow shallow groundwater from the Nardouw Aquifer to recharge the deep Peninsula Aquifer. The lowest K value of $1.0E-15$ makes the C/S to act as a barrier, a situation where there is no water exchange between the upper and the lower deep aquifers. This situation is not realistic in the TMG since the area is structurally disturbed.

Flow paths with long travel times of about 405 years were estimated with the use TopoDrive. These ages are not correlating with the age estimated Carbon-14 data. These isotopic analyses have estimated several thousands years for most of thermal spring water in the TMG. The reason for this may be the fact that the groundwater model presented here does not account for all the complexities of the fractured rock aquifer. These include structural setting, distance from recharge to discharge point, and permeabilities of the aquifers. This analysis demonstrates the high level of uncertainty involved in calculating travel times from the recharge area to discharge area, given our current limited knowledge of the appropriate parameter values. It is therefore recommended that further detailed research studies be conducted on the aspects of deep groundwater reviewed in this thesis.

January 2007

DECLARATION:

I declare that *A Preliminary Understanding of Deep Groundwater Flow in the Table Mountain Group (TMG) Aquifer System* is my own work, that it has not been submitted for any degree or examination in any other university, and that all the sources I have used or quoted have been indicated and acknowledged by complete references.

NAME: **Khangweleni Fortress Netili**

SIGNED.....

DATE: **January 2007**

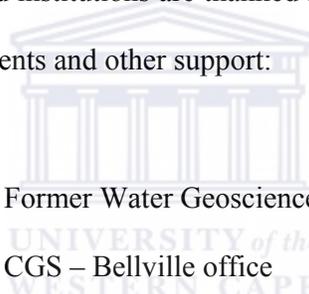


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CHAPTER 1 GENERAL INTRODUCTION

1.1 INTRODUCTION

The Table Mountain Group (TMG) Aquifer is the second largest aquifer system in South Africa, after the dolomites. This aquifer has the potential to be a significant source of water for the people of the Western Cape. But many uncertainties remain, considering that the need to move groundwater exploration in this country an order of magnitude deeper, from 100m to 1000m (Pietersen and Parsons, 2002). The occurrence of hot water springs in the TMG in relation with the main geological fault systems in South Africa, show that deep flow systems do exist. Little is known, flow mechanisms for example, about these deep aquifer systems in South Africa.

Although the aquifer system is used to some extent, a number of aspects relating to the aquifer system are poorly understood and unquantified. Deep groundwater movement in the TMG Aquifer is via fracture systems, which are either horizontal bedding planes or vertical joints (Diamond and Harris, 2000). Because of the intended development and use of these deeper groundwater resources, the occurrence of deep groundwater in the TMG aquifer system also needs to be investigated.

1.2 BACKGROUND TO THE STUDY

A short description of the current state of knowledge on groundwater flow in the TMG area is provided here after which the objectives of the present study are presented.

In deep aquifers the complex flow pattern originating from the geological structure often leads to difficult predictions of water origin, determination of the main flow paths, and potential mixing of waters. All these uncertainties prevent an efficient management of the groundwater resource.

The Table Mountain Group aquifer system is used to some extent for water supply and other uses, but a number of aspects relating to the aquifer system are not well understood. Deep groundwater flow in this aquifer is one of the aspects that are currently debated. The water demand is also increasing with increase in population. The increased pressure on the water resources in the TMG area will inevitably imply that more information will be required in the near future for groundwater exploration. The Water Research Commission (WRC) has initiated various projects on the study of the TMG aquifer (Pietersen and Parsons, 2002; Weaver *et al.* 1999; Cavé and Clarke, 2003; Kotze, 2002). There are a number of projects that are underway on the TMG area. A project, for example, ‘A feasibility study of the TMG’, is also underway. The TMG Aquifer Feasibility Study and Pilot Project aims, as stated on its website, to determine: 1) The feasibility of using the TMG Aquifer as a water source to augment Cape Town's water supply; and 2) The environmental impacts associated with using this resource. As part of this WRC project deep exploratory boreholes will be drilled. More information on this project is available at www.tmg-aquifer.co.za. Another WRC project entitled “Flow conceptualisation and storage determination in the Table Mountain Group Aquifer System”, which this study is part thereof, was also initiated.

The hydrogeological community is currently debating the occurrence of deep groundwater flow in the TMG. This has encouraged the current study which focuses on the number of aspects relating to this topic.

There are a number of methods used to study groundwater dynamics, including flow patterns. Hydrochemistry has long been used to understand the evolution and movement of groundwater, including flow paths. In the last four decades the use of isotopes, either naturally occurring (environmental isotopes) or intentionally injected (artificial isotopes) have proved their value in the studies related to water resources assessment, development and management. The applications in a large variety of hydrological problems are based on the concept of *tracing* (Yurtsever 1999). Geochemistry and isotopic study conducted at the Agter-Witzenberg area (Weaver *et al.*, 1999) has proved that this integrated approach is powerful in groundwater resource evaluation. For the development and management of deep groundwaters in the TMG, proper investigation of different aspects is crucial (i.e. deep groundwater recharge, flow mechanisms, and mean residence times). Simulation of groundwater flow paths is also important.

1.3 STUDY AREA SELECTION

This research forms part of the WRC project (K5/1419/1) entitled “Flow conceptualisation and storage determination in the Table Mountain Group Aquifer System”, which has started in 2001. The former phases of this project resulted in the subdivision of the TMG into 19 hydrostratigraphic sub areas. The details of the

methodologies used to subdivide the TMG into hydrostratigraphic subareas are discussed in Lin (2003).

For the groundwater flow modelling purpose this study will focus on the Montagu area (falling in Subarea 10). This area was selected because of the following reasons:

- The well field recently developed at Koo Valley, where slightly deep boreholes (>240 m) were drilled as part of the development of the Montagu groundwater scheme.
- The occurrence of two thermal springs (Montagu and The Baden), which are regarded as discharge points for deep flowing groundwater.

The above-mentioned reasons make this area to be of great interest in terms of groundwater resource investigation and development because some historical data is available.

1.4 AIM AND OBJECTIVES

The main aim of this thesis is to investigate deep groundwater flow mechanisms in the study area to develop a conceptual model of deep groundwater flow discharging in thermal springs. The following, but not limited to, underpin the specific objectives of this research project:

- To review of related groundwater concepts in order to distinguish between deep and shallow groundwater.
- To estimate depth of groundwater circulation
- To simulate deep groundwater flow paths

- To estimate groundwater travel times from selected springs.

1.5 STRUCTURE OF THIS THESIS

The introductory information is given in Chapter 1. The general information given includes background to the study, aims and objectives of the thesis. Methodology followed throughout the study is also introduced in this chapter.

Relevant literature has been reviewed in Chapter 2. The information presented in this chapter gives insight into some of the concepts and available techniques (i.e. hydrochemistry, geothermometry and isotopes), and results from various related case studies. Some of the data derived from the literature review has been presented as tables, figures in the context.

Chapter 3 provides study area description; location and extent, physiography, drainage, geology and hydrogeology, and methodology. Chapter 4 gives insights into aspects related to the occurrence of deep groundwater in the TMG area, including some case studies and conclusions. Chapter 5 deals with the conceptual modeling of deep flow and particle tracking simulation and discussions based on the results of current study are presented. Conclusions and recommendations are outlined in Chapter 6.

1.6 METHODOLOGY

The study comprises three phases, namely; (1) Desktop, (2) Fieldwork, and (3) Data analysis and interpretation coupled with report writing. The desktop study includes literature search and review and/or collection of available data. Sources of data used in this phase include published and unpublished reports, both local and international. Major local sources include Department of Water and Forestry's National Groundwater Database (NGDB), Council for Geoscience, and consulting firms such as SRK Consulting. The type of data acquired includes borehole data and reports. The data collected from reports and fieldwork gave aid to the knowledge of concepts and the situation of the groundwater resource in the study area, and it was used for planning of the fieldwork. This study focused much into this data set, as it was the available reliable source of data.

Fieldwork was limited and only water levels were measured in the field. The data obtained from reports and National Groundwater Database (NGDB) was interpreted, and water levels were compared. Water levels were used to produce groundwater contour and flow directions maps for the understanding of the flow behavior in the field.

A simple conceptual flow model was developed with TopoDrive (Hsieh, 2001) with the aim of identifying sensitive parameters Particle tracking was performed on the flow model in order to compute flow paths and travel times, and therefore in conceptualizing deep groundwater flow. An interpretation of available data and results of particle tracking led to the conceptualization of deep groundwater flow.

CHAPTER 2 REVIEW OF DEEP GROUNDWATER FLOW

2.1 INTRODUCTION

The term '*deep groundwater*' is widely used by the hydrogeological community. The use of this term is normally dependent on the objectives of the study. This section tries to give differences, for a better understanding, between shallow and deep groundwater. Varieties of aspects concerning the subject are also outlined.

2.2 ORIGIN AND DEFINITION OF DEEP WATERS

The issue of the origin of deep thermal or hot waters has long been discussed among geologists and hydrogeologists. There are at least two possible sources of these waters:

- Water may rise from (or from the cooling of) the hot magmas at depth or,
- Water may originally be rainwater which has percolated down deep into the earth's crust, there to be heated by geothermal gradients.

The most important process that causes heating of rainwater is increase in temperature with depth or geothermal gradients. The relationship of temperature with depth can be described by a linear geothermal gradient for the upper regions of the crust. The temperature depth relationship is described the equation 2.1 (Cave and Clarke 2003) below:

$$T_z(^{\circ}\text{C}) = G \times z + T_s(^{\circ}\text{C}) \quad (2.1)$$

Where T_z is the temperature at depth below surface, G is the geothermal gradient, typically expressed in units of $^{\circ}\text{C}/\text{km}$, and T_s is the temperature at surface ($z=0$). There is a general tendency on the change of temperature with depth. The average increase worldwide is on the order of $2\text{-}3^{\circ}\text{C}$ for every additional $90\text{-}100\text{m}$ on average air temperature, and is due to the release of heat by the natural radioactive decay of potassium, uranium, and thorium in the crust as well as from compression of the planet's interior by gravity (National Park Service, available at www.nps.gov).

When rainwater percolates downward, through fractures, cracks or any other preferential pathways, it can reach greater depths. The journey downward is slow, and as the water becomes heated it dissolves silica and other minerals from the surrounding rocks. The amount of silica present in the water provides the geochemists with an accurate indication of the maximum temperature to which the waters are heated (Kent, 1949).

Major and minor fractures, normally associated with faults, provide a ready route to the surface for hot or thermal waters collected from a broad area. The heated water uses these routes upward and then discharged at lower elevations as thermal or hot springs. The trip is normally so rapid so that there is very little cooling of the water.

There are at least three factors that cause the flow of heated waters to the surface:

- Much of the recharge normally takes place at slightly higher elevation than springs. That is water will normally descend to great depths but will be discharged at lower elevation than its recharge elevation.
- Heating at depth causes water to become buoyant

- In case of the fractured-rock environment, large amounts of rainwater are collected from a broad area, where it infiltrates through fractures and fissures with low vertical conductivity, descends to great depth and then directed rapidly up through a much narrower escape route.

There is always a possibility that thermal water rising from great depths will be affected in some of its characteristics by the natural processes including decrease in temperature, interaction with cold water and atmospheric effects (Imbach, 1997). It should be noted that not only the temperature can be used to characterize deep groundwater but aspects such as age, rate and velocities, chemistry and isotopes can be of great importance when classifying deep groundwater from shallow groundwater as shown in section 2.3.

2.3 CONCEPT OF SHALLOW- AND DEEP GROUNDWATER

After consideration of available reports, it was found that there is no formal and clear definition or 'strict dividing line' for distinguishing between shallow and deep groundwater. This section outlines the criteria for distinguishing between shallow and deep groundwater. It normally depends on the objectives of the study, for example, for an agricultural engineer, >25 m might be deep, 500 m might be deep for water supply whereas for mining activities deep can be several thousands of meters. In the Oil and Gas industry, the term *deepwater* means 400 m or greater (McLennan and Williams, 2005).

2.3.1 International context

This section provides common approaches to distinguish between shallow- and deep groundwater in different parts of the world. Few case studies and views of few professionals are highlighted. The approaches include use of hydrochemistry and isotopes, depths of circulation, and concept of confined and unconfined aquifers are also highlighted.

2.3.1.1 Hydrochemistry

2.3.1.1.1 *Isotopes*

Seiler and Lindner (1995) argued that groundwater mobility and age, deduced from radionuclides tritium and carbon-14, may also be used to subdivide groundwater into shallow and deep groundwater.

As these radionuclides enter the groundwater system through precipitation infiltration, they make groundwater flow dynamics measurable in tens (tritium) and thousands (carbon-14) of years (Seiler and Lindner, 1995). Their concentrations in groundwater decrease with time due to radioactive decay. The classification of groundwater in this case relies on the concept of '*tritium zero limit*', which is the depth at which tritium concentrations become undetectable. Below this limit, the water ages exceed 50 years and rise over a short depth ranges to ages of several thousand years.

In a study conducted in Germany (Seiler and Lindner, 1995), where groundwaters are not overexploited, the '*tritium zero limit*' referred in the text as TNF (Tritium-Null-

Flache) was found at an average depth of 40 and 100 m. These depths were found to be dependant on the infiltration capacity of soils and/ or unsaturated zone, the quantity of groundwater recharge, and the porosity and permeability of the aquifer. They further noted that observations of vertical hydraulic gradients of groundwater containing tritium discharge to the nearest river whilst groundwaters below TNF are directed to the one of the deeper systems in a region (see Figure 2.1).

2.3.1.1.2 General chemistry

Groundwater chemistry can also play an important role in distinguishing shallow groundwater from deep groundwater. Due to high change in chemistry with depth in crystalline rocks (i.e. Total Dissolved Solids (TDS) and temperature), groundwater gradually changes into a hot saline called crustal deep-fluid (Stober and Bucher, 2005). For example, in fractured rock aquifers near the surface, waters are typically low in TDS, low in sodium concentration and are of a Ca-HCO₃ type whilst at great depth deep fluid (groundwater) is of a Na-Ca-Cl type (Stober and Bucher, 2005). This hydrochemical stratification has been reported in most parts of the world. The review of the hydrochemistry of these deep crustal fluids is given in Frapé et al. (2004).

According to Stober and Bucher (2005) the general increase of temperature with depth is about 2 to 3 degree Celsius per every 90 to 100 m. At 4-5 km the temperatures of these deep fluids range from 100 to 200 °C. Temperatures of this magnitude were measured in two deep wells; 12.5 km Kola in Russia with bottom hole temperature of 240 degree Celsius and 9.1 km KTB in Germany, with 270 degree Celsius (Stober and Bucher, 2005)

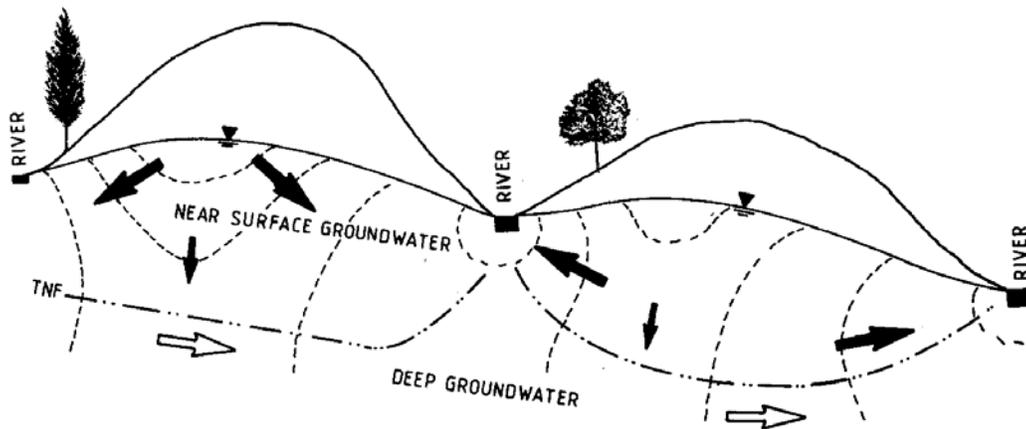


Figure 2.1 Schematic local (dark arrow) and regional (light arrow) groundwater flow patterns and the position of the tritium zero limit (TNF) (Seiler and Lindner, 1995).

To accomplish the results obtained from the isotopic data a simulation model has to be developed to test the results. Seiler and Lindner (1995) set a two-dimensional model to simulate groundwater recharge over depth, flow velocities and ages. Below the active recharge zone the calculated water ages increase very rapidly with depth as observed in the field using radioactive isotopes, especially tritium and carbon-14. These results proved that the TNF separates the approximate zone of active high flow velocities (shallow) from a very slow flow velocities (deep groundwater). They then argued that although all interfaces between zones of different activities have to some extent an arbitrary character, the TNF seems to be quite useful in separating shallow- from deep groundwater, because tritium traces the water molecule, enters the subsurface system only with precipitation (with very few exceptions) and is present in precipitation in both hemispheres.

In Germany, shallow groundwater ranges in depths between 10 and 100m. Deep groundwater, in contrast, participates in the water cycle only at intervals of hundreds of years and reaches depths of several hundred meters (Seiler and Lindner, 1995).

2.3.1.2 Concept of unconfined and confined aquifers

In the United States, Thomas Harter (*pers. comm*) agrees that shallow groundwater systems and deep groundwater systems commonly refer to the unconfined aquifer and confined aquifers respectively. The unconfined aquifer is often considered the shallow aquifer, because it would be the uppermost system in a multiple aquifer system. Confined aquifers are typically considered deep aquifers. "I have used 'deep' aquifer for confined aquifers at depths of 100 to 500 m below ground surface"- said Thomas Harter of the UCCE Groundwater Hydrology Program, University of California, Davis (*pers. comm, e-mail*). This may be completely different from different geologic environments. On the other hand, he used the term "shallow groundwater" for the uppermost 10 m of an unconfined aquifer. This shallow groundwater is the youngest and most recently recharged groundwater in areas studied.

According to De Vries (2005, *e-mail*), of the Faculty of Earth and Life Sciences, Vrije Universiteit Hydrology Department in The Netherlands, deep groundwater in most cases would be constrained within the confined aquifer system. This groundwater hardly participates in the present day water cycle. Such aquifers are filled with old water that might have been infiltrated under different conditions, i.e. (paleo-) climatic and/or (paleo) topographic conditions. This water is subjected to processes that are important on a geologic time scale like diffusion, thermo-chemical convection

(density-driven) and compression-driven flow, and residual flow. This water participates in the water cycle only at intervals of a few hundreds to thousands of years and reaches great depths. They are characterized by ages of more than 50 years, slow flow velocities and thus long residence times (from recharge area to discharge point).

2.3.2 South African context

In South Africa, like in other countries, there is no strict definition of "deep" and "shallow" ground waters. The objectives of a study normally drive the classification scheme. This section outlines the proposed considerations towards the criteria for distinguishing between shallow and deep ground waters.

The issue of classifying groundwater into shallow and deep groundwater in South Africa tends to follow a global trend of using general methodologies such as the ones discussed above. Much focus in this section will be put on the classification based on temperature related to depth of circulation.

2.3.2.1 Temperature

The earth's temperature increases with depth. This increase in temperature is due to natural processes; volcanic activity, radioactive decay of some radioactive isotopes and from the compression of a planet's interior due to gravity. Temperature may be used as an indicator of water that has circulated to a certain depth below the surface. High water temperature, normally at thermal spring for example, may therefore

represent deep (regional) flow path from recharge to the point of discharge. James *et al.* (2000) found that temperature measurements at springs and boreholes are one means of assessing the relative scale of groundwater flow because deeply circulating groundwater will acquire more geothermal heat than groundwater circulating at shallow depths. Generally, the water in the thermal springs is at least few degrees warmer than the mean annual air temperature for the localities where they occur. Some investigators used 6 °C above annual air temperature as an arbitrary temperature divide between cold springs and thermal springs (Mazor, 1991; Diamond and Harris, 2000). This means that classification between shallow and deep waters may be done based on the depth at which water has circulated which is portrayed by water temperature.

In this study 25°C, where a mean air temperature of 18°C and the springs show water temperatures by more than 7°C, is set to classify shallow from deep waters, using equation:

$$D = \frac{T_{clct} - T_{avg}}{G} \quad (2.2)$$

where G is the mean regional geothermal gradient (°C/km), D is depth (km), T_{calc} is the temperature calculated by a plausible geothermometer (°C), and T_{avg} is the yearly average air temperature of the sampling locality (°C). This means that this dividing isotherm will be variable depending on the topography, local mean air temperature and geothermal gradients. For example, Diamond and Harris (2000) argued that in Western Cape valleys and coastal plains average annual air temperatures range from

15 °C to 20 °C whereas Jones (1992) noted geothermal gradients ranging from 10°C/km to 20°C/km.

The limitation of using water temperature as a dividing line between shallow and groundwater is that in some areas (e.g. TMG aquifer) geothermal gradients are not well established. In a feasibility study on the use of chemical geothermometers for tracing deep groundwater flow in the TMG, Cavé and Clarke (2003) used the minimum geothermal gradient of 10°C/Km and the highest geothermal gradient of 20°C/Km to calculate circulation depths of spring waters. They reckon that 20°C/Km seemed more suitable for the Cape Fold Belt. The data from this study indicated that groundwater supplying the Nardouw Subgroup boreholes may circulate from 0 to 500 m below surface, while the water feeding the Peninsula Formation could circulate to depths of 200 to 3000 m. These estimates were found to be generally deeper than minimum depths previously given by Weaver and Talma (1999).

2.3.2.2 Thermal spring occurrences in the Republic of South Africa

Kent (1949) reported that there are 74 springs that are regarded as thermal; issuing at temperatures of >25 °C. Figure 3.2 shows the distribution of thermal springs in the TMG area. Diamond and Harris (2000) reported a different number (>84) of thermal springs that occur in South Africa, ranging from 25°C to 64 °C. The geology and chemical compositions of these springs have been given in Kent (1949) and Hoffmann (1979) respectively, whilst Mazor and Verhagen (1983) and Diamond and Harris (2000) give isotopic data of selected springs in the Western Cape Province. These springs are regarded as the manifestation or issue points for the heated meteoric

water that has circulated to deeper depths, hence deep groundwater as referred to in the context of this thesis.

Heated waters, irrespective of the source of heat, may be classified into different classes depending on their temperatures. In the South African context, a division between non-thermal (cold) and thermal waters at 25°C and where the mean annual air temperature ranges from 12.8°C to 21.1°C (18°C taken as average) , was proposed and the following classification was done by Kent (1949):

Warm	25° C - 37°C
Hot	37°C - 50°C
Scalding	>50°C.

2.4 DEEP UNDERGROUND GEOLOGIC ENVIRONMENT

Deep groundwater flow is largely controlled by the hydrogeological properties of the deep underground geologic environment. This section gives an overview of the important aspects of the deep geologic environment that influence groundwater flow at depth. These aspects should be considered when assessing deep groundwater.

2.4.1 Hydraulic properties at depth and groundwater circulation

In a geographical setting groundwater flow through rock mass or structures is generally driven by changes in elevation. This generally means that water will tend to flow along hydraulic gradients from the inland areas towards the coast, through more

permeable sections of the rock mass. In case of the geological environments where complicating diagenic and tectonic effects do not prevail, permeability will tend to decrease with increase in depth. This is due to the increase in the confining pressure with increasing depth, leading to closure of pores and fracture aperture. Thus, groundwater flow will decrease at great depth as compared to the flow in the near-surface environment where the confining pressure is minimal. This tendency is not always the case in the fractured rock environment where open fractures also prevail at great depth. Kotze (2002) states that fractures occur in the TMG that are still open in the depths of more than 200 m.

For fractured rock environment borehole yields may increase with depth, instead of decreasing. This character has also been witnessed in the TMG aquifer system (Weaver et al., 1999). This is normally evident where the rocks possess a brittle form of deformation and in the absence of clay or cementing minerals. In this type of environment it is important to understand how groundwater flow is controlled by the local and regional presence of transmissive fractures. Understanding the distribution of hydraulic properties with depth is important in assessing deep groundwater flow. The following properties should be considered:

- Spatial variations in hydraulic gradients and distributions of permeability with depth
- Hydraulic conductivities of the underground rock mass
- Transmissivity and storage
- Distribution of recharge at depth
- Water balance

In a fractured rock aquifer, groundwater flow is largely dependent on hydraulic properties of the geological structures such as folds and faults. In case of faults, their orientations and interconnectivity remain a major influence of groundwater flow. Within a single fracture, dispersion within the fracture and diffusive transport out of the fracture and into the surrounding rock mass are both possible. It is also important to understand fracture geometry, fracture networks particularly for those that are thought to be more significant in controlling groundwater flow at depth.

2.4.1.1 Hydraulic gradients at great depth

Hydraulic gradient is determined by dividing the difference between the hydraulic heads at upstream and downstream points by the distance between the two points. Information on these gradients is important in assessing deep groundwater flow. However, there are few, if any, available in situ measurements of gradients for the deep geological environment of the TMG area.

2.4.1.2 Hydraulic conductivity of the deep underground rock mass

As mentioned earlier that there is limited information on the hydraulic properties of the deep underground rock mass, this also include the hydraulic conductivity. Information (i.e. Lugeon test data) provided by the civil engineering projects reports can be of great help. In Lugeon tests, a rock with 1-3 Lugeon (1 Lugeon is equivalent to hydraulic conductivity of $1.33 \times 10^{-7} \text{ m s}^{-1}$) is generally considered to have low permeability. Unfortunately, data are not available for deep TMG geological environment. Alan Shelly (*pers. comm.*) of the Ninham Shand (Pty) Ltd. (Engineers)

reckons that the only data available for TMG area in their databases is for the dam construction where they drilled to the depth of 50 m.

Understanding the hydraulic properties of these fractures (including fold geometry, fault and fault zones) is required when trying to assess deep groundwater flow. The hydraulic conductivity of fault crushed zones is known to depend partly on the occurrence and distribution of low permeability minerals, such as clay minerals (Okamoto et al., 1998). For example, the hydraulic conductivity of a fault filled with clay is typically in the order of 10^{-8} to 10^{-5} m s^{-1} , whilst that of the fault breccia is typically of 10^{-6} to 10^{-3} m s^{-1} (see Table 2.1). The effect of the infill material of the fault can inhibit the hydraulic conductivity of the fault completely causing the fault to act as a hydrogeological barrier (Bense et al., 2003). The variations in fault permeability can be revealed by variations in the distributions of hydraulic heads. According to Bense et al. (2003) sealing faults will normally cause very high hydraulic head gradients whereas some will show little or no significant impact on the distribution of hydraulic heads. In the TMG area, the Cango Fault has been reported to be of the latter nature (Kotze, 2002).

The relationship between hydraulic conductivity and depth is also important. Snow (1968) argued that the depth-dependence of the hydraulic conductivity in granite is explained by a decrease in fracture aperture due to increase in the in situ confining stress with increasing depth. The data used to draw this conclusion was measured at depths of less than 100 m. Measurements at depth of more than 100 m were done in crystalline rocks of Sweden and Canada (Almen et al., 1986; Davidson et al., 1994; Neretnieks, 1993). The data from these measurements also supported the depth-

dependence of hydraulic conductivity, and that greater decrease in hydraulic conductivity is witnessed at depths greater than 500 m.



Table 2.1 Hydraulic conductivity of a fault and fault zones (Tanaka and Inoue, 1995)

Rock type	Geological Time	Filling width of fault and fracture zone	Cohesion (kgf cm ⁻²)	Internal friction angle (°)	Hydraulic conductivity (m s ⁻¹)
Sandstone, slate	Mesozoic	Clay 10~15cm	0.25	35	1x10 ⁻⁵ ~ 1x10 ⁻⁶
Granodiorite	Mesozoic	Clay 1~30cm	2~3	30~40	
-	-	Sandy 5~6m	5~6	40~45	
Varved clay	Paleozoic	Breccia 10~15m	-	-	1x10 ⁻³ ~ 1x10 ⁻⁶
-	-	Clay	-	-	1x10 ⁻⁶ ~ 1x10 ⁻⁸
Granite	Mesozoic	Clay 9m	0.5	20	1x10 ⁻⁶ ~ 1x10 ⁻⁷
Hornfels	Paleozoic	Braccia -	-	-	1x10 ⁻⁵
-	-	Sandy -	0.5	20	1x10 ⁻⁶
-	-	Clay -	-	-	1x10 ⁻⁷
Sandstone	Paleozoic	Breccia -	-	-	1x10 ⁻⁵
-	-	Clay -	-	-	1x10 ⁻⁵ ~ 1x10 ⁻⁶
Slate, chert	Paleozoic	- -	-	-	1x10 ⁻⁶

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2.5 ISOTOPES IN GROUNDWATER STUDIES

Isotopes are atoms of the same element, in which the nuclei of the atoms have the same number of protons but different numbers of neutrons. In the last four decades the use of isotopes, either naturally occurring (environmental isotopes) or intentionally injected (artificial isotopes) have proved their value in the studies related to water resources assessment, development and management. According to Verhagen and Butler (2004), W.F. Libby and co-workers established the presence of radioactive carbon-14 and tritium and Craig (1961) developed the systematics of the stable isotopes of hydrogen and oxygen in the hydrological cycle. The applications in a large variety of hydrological problems are based on the concept of *tracing* (Yurtsever 1999).

The summary of the main hydrological use of environmental isotopes (after Lloyd and Heathcote, 1985; Weaver *et al*, 1999) is given below:

- To provide a signature to a particular water type or
- To monitor the origins of water (Harris *et al*, 1999).
- To determine the mean recharge elevation (James *et al.*, 2000)
- To identify the occurrence of mixing of two or more water types
- To quantify interactions between groundwater and surface water
- To provide residence time information about groundwaters
- To provide information concerning travel times and groundwater velocities
- To provide information on water rock-interaction

2.5.1 Stable isotopes

Normally, stable isotopic abundances are reported as ratio to some abundant isotope of the same element or as positive or negative deviations of these isotope ratios (R) away from a standard. The relative difference is designated by delta (δ) (normally expressed as parts per thousand, permil ‰).

$$\delta_x = [R_x - R_{std} / R_{std}] \times 1000 \quad (2.3)$$

In the case of the ratio of oxygen-18 to oxygen-16 equation (2) becomes:

$$\delta^{18}\text{O}_{\text{sample}} = \{[(^{18}\text{O}/^{16}\text{O})_{\text{sample}} - (^{18}\text{O}/^{16}\text{O})_{\text{std}}] / (^{18}\text{O}/^{16}\text{O})_{\text{std}}\} \times 1000 \quad (2.4)$$

where R is the particular isotope ratio (i.e. $^{18}\text{O}/^{16}\text{O}$ or $^{13}\text{C}/^{12}\text{C}$ or D/H) for the sample and standard. The standard used for Carbon is PDB (Pee Dee Belemnite) and for oxygen and hydrogen is SMOW (Standard Mean Ocean Water).

In practice, each laboratory has its own standard or set of standards, which have been calibrated against the SMOW scale (Drever, 1988).

The naturally occurring stable isotopes hydrogen (H), deuterium (D) oxygen-16 (^{16}O), oxygen-18 (^{18}O), carbon-12 (^{12}C) and carbon-13 (^{13}C) have become useful tools in groundwater studies (Weaver *et al.*, 1999).

2.5.1.1 Stable isotopes of oxygen and hydrogen

As water molecules are made up of hydrogen and oxygen atoms, hydrogen (deuterium) and oxygen (oxygen-18) isotopes are of key importance to most groundwater studies. These isotopes are commonly used to determine average elevation of recharge for a particular mass of groundwater (Christodoulou, et al., 1993; Scholl *et al.*, 1996; Williams and Radoni, 1997; James, et al., 2000). Hydrogen and oxygen isotope ratio are highly conservative tracers for monitoring the origins of water (Harris *et al.*, 1999). One of the assumptions made in applying deuterium studies to hydrology is that the climate of the earth has not changed appreciably during the last two hundred years (Cindrich and Gudmundssori, 1984). The deuterium and oxygen-18 content of water varies with altitude, geographic location, and evaporation. These variations in stable isotope content can be used with conventional data, and tritium and C-14 results, to define recharge and discharge areas, trace mixing patterns, and evaluate the importance of evaporation in the chemical balance of the aquifer. These data are important in understanding controls on water quality and in regulating activities that may pollute groundwater resources.

After considering the complexity of the hydrological cycle, Craig (1961) found that $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in fresh waters correlate on a global scale. This is because hydrogen and oxygen occur together in water molecules and both experience the same sequence of events during the migration of air masses. Craig's "global meteoric water line" (GMWL) defines the relationship between ^{18}O and ^2H in worldwide fresh surface waters:

$$\delta^2\text{H} = 8 \delta^{18}\text{O} + 10 \text{‰} \quad (2.5)$$

According to Faure (1998) the slope of the meteoric water line is related to the isotope fraction factors of the H and O during evaporation or condensation of water. The intercept at $\delta^2\text{H}=+10 \text{‰}$ for $\delta^{18}\text{O}=0\text{‰}$ is called the *deuterium excess*. Both the slope and the intercept vary depending on the local climatic conditions (Hoefs, 1996; Faure, 1998).



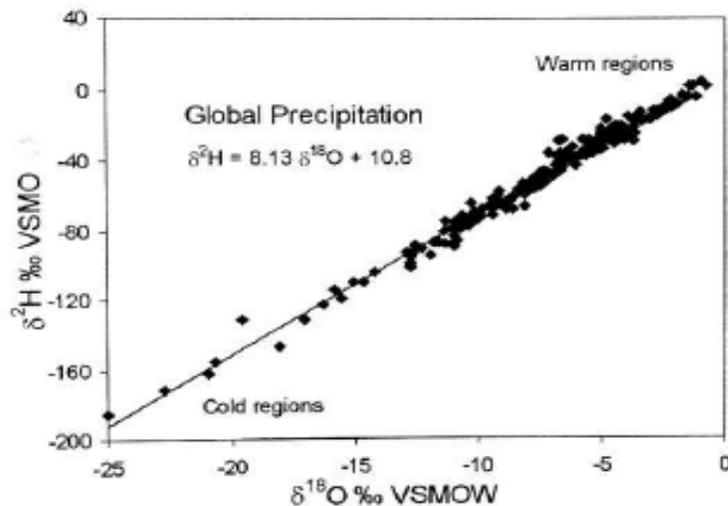


Figure 2.2. The relationship between ^{18}O and ^2H (Clark & Fritz 1997)

Analysis of δD and $\delta^{18}\text{O}$ can be used in conjunction with Figure 2.2 to identify the probable source of underground water. This is because the δD is generally unaffected by reaction with aquifer material at low temperatures. Craig (1963) showed that for most geothermal water the deuterium content is approximately equal to that of the local meteoric water. Where deuterium levels are depleted relative to the local precipitation, it is probable that the system is fed from an aquifer originating in higher terrain where deuterium levels are lower. The $\delta^{18}\text{O}$ is generally unaffected by reaction with silicates at low temperatures for short period of time (1 million years or so). Thus if the isotopic composition of an underground water plots close to the meteoric water line in a position similar to that of present-day precipitation in the same region, the water is almost certainly meteoric. If the water has the same δD value as local precipitation by a slight heavier $\delta^{18}\text{O}$ value, the water is probably meteoric, but has been affected by exchange with calcite (Craig 1961).

If it plots close to the seawater value, the seawater value is far from local precipitation range, the water is not meteoric, and may perhaps be connate (Drever, 1988).

Variation in the equation for the meteoric line at a specific location is a function of its climate, geographic location and the source region of the evaporation. The IAEA (1981) has reported southern African meteoric water line as $\delta D = 6.1\delta^{18}O + 5$. In a recent study, Diamond and Harris (1997) found that Western Cape meteoric water defines a meteoric water line with approximate equation $\delta D = 6.1\delta^{18}O + 8.6$. The isotope composition of water that recharges groundwater in the Western Cape should approximate the weighted annual δD and $\delta^{18}O$ values.

2.5.2 Radionuclides

Naturally occurring radioactive isotopes of hydrogen (tritium) and carbon (carbon-14) are used to determine the age (time since recharge) of the ground water. The presence of tritium is generally used to identify younger ground water (water recharged after 1952). Carbon-14 data is used to estimate the age of older ground water that did not contain tritium.

2.5.2.1 Tritium

Tritium (^3H) is a radioactive isotope of hydrogen. It is a short-lived isotope with a half-life of 12.43 years. It has attracted considerable interest during the era of thermonuclear bomb testing (Clark and Fritz, 1997), which led to higher tritium concentrations during the 1960s.

Small but measurable amounts of ^3H are also produced naturally in the stratosphere by cosmic radiation on ^{14}N . Both natural and anthropogenic tritium enters the hydrological cycle through precipitation. Its presence in groundwater provides evidence of active recharge.

The absolute concentration of tritium in groundwater is given in tritium units (TU). One TU corresponds to one ^3H atom per 10^{18} atoms of hydrogen. Measurable tritium in groundwater usually signifies modern recharge. High tritium (>30 TU) indicates recharge in the 1960s while low values (<1 TU) usually signify paleogroundwater (older groundwater) that has mixed with shallow modern groundwater. The depth to which tritium concentration becomes immeasurable is called the '*tritium zero limit*'. Below this limit, water ages exceed 50 years and rise over short depth ranges to ages of several thousand years (Seiler and Lindner, 1995).

Although qualitative and quantitative approaches to dating groundwater is also possible with tritium (Clark & Fritz 1997), the direct age determination of groundwater accurate to the year is somewhat uncertain partly due to the unknown extent of mixing of each year's recharge with that of the previous year and partly because of high local and temporal variability of the input values in precipitation. Mazor (1991) and Weaver *et al.* (1999) gave guidelines that should be considered when investigating groundwater age with tritium:

- Water with zero tritium (in practice <0.5 TU) has a pre-1952 age.
- Water with significant tritium concentrations (in practice >5 TU in the southern hemisphere) is of a post-1952 age.

- Water with little, but measurable, tritium (between 0.5 and 5 TU) seems to be a mixture of pre- and post-1952 water.

However, by measuring ^3H together with its daughter ^3He , true age determination is possible by calculations not based on the complicated tritium input function (Clark & Fritz 1997).

2.5.2.2 Carbon-14

This radioactive isotope is also produced in the atmosphere by cosmic rays interaction with nitrogen and was introduced in large amounts by nuclear weapons testing. The natural production in the atmosphere is balanced by decay and burial to maintain a steady-state atmospheric $^{14}\text{CO}_2$ activity of about 13.56 disintegration per minute (dpm) per 1 gram of C, or about 1 ^{14}C atom per stable 10^{12} C atoms (Clark and Fritz, 1997). It has a half-life of 5730 years, making it useful tool for dating waters as old as 50,000 years. Carbon-14 produced in the atmosphere is introduced into the groundwater via precipitation.

Carbon-14 undergoes radioactive decay (to ^{14}N), so once isolated from the atmosphere the amount of Carbon-14 decreases with time according to the equation:

$$(^{14}\text{C})_t = (^{14}\text{C})_0 e^{-kt} \quad (2.6)$$

Where $(^{14}\text{C})_t$ is the amount at present time t , $(^{14}\text{C})_0$ is the amount present at time $t=0$, and k is the decay constant ($k= 1.2 \times 10^{-4} \text{ year}^{-1}$ for ^{14}C) which is related to half life $T_{1/2}$ by the equation

$$T_{1/2} = \ln 2/k \quad (2.7)$$

To determine the time since water was in last contact with the atmosphere, it is necessary to know $(^{14}\text{C})_0$ (Drever, 1988). Measurements of ^{14}C are reported in terms of *per cent modern carbon (pmC)*, where

$$A = [(^{14}\text{C}/^{12}\text{C})_{\text{sample}} / (^{14}\text{C}/^{12}\text{C})_{\text{modern}}] \times 100 \text{ pmC} \quad (2.8)$$

Radiocarbon contents greater than 100 pmC show the presence of bomb- ^{14}C and indicate ages less than 44 years (post AD). However, there are some complications in the behavior of ^{14}C during recharge, so the “absolute” age of groundwater cannot be determined reliably. If the ^{14}C concentration is measured at several points along the flow line within an aquifer, the difference in age between the points and hence the flow velocity can be determined (Vogel, 1967; Drever, 1988; Weaver *et al.*, 1999).

Drever (1988) mentioned that one of the complications is the dissolution of carbonate minerals or oxidation of organic matter within an aquifer which may add “old” or “dead” (no detectable ^{14}C) carbon to the water, giving an erroneously old age. The contribution of carbon from these sources can sometimes be estimated from $^{13}\text{C}/^{12}\text{C}$ measurements and chemical arguments, so corrections can be made. Another complication is the mixing of different waters of different ages within the same aquifer. A low ^{14}C concentration may mean that we are looking at a relative “young” water and “dead” water. The proper interpretation of ^{14}C measurements in groundwater requires knowledge of the flow of water, the source of the dissolved inorganic carbon in the water and geohydrological observations, and measurements are required to establish this (Drever, 1988; Weaver *et al.*, 1999).

Dating groundwater with radiocarbon cannot be done on the water molecule itself but will rely on the dissolved inorganic and organic carbon in the water (DIC and DOC). Both forms enter groundwater from atmospheric $^{14}\text{CO}_2$ via the soil zone. Old groundwater can be dated if sufficient time has passed for measurable decay of the initial ^{14}C activity (Clark and Fritz, 1997).

2.5.2.2.1 Correction of Carbon-14 age

By various interactions of recent and fossil carbon an initial ^{14}C value between 50 and 100 *pmC* results. Other, partly bacterial processes can also influence the isotopic composition of a groundwater. During the development of the ^{14}C method several correction models, from *statistical models* and *mixing models* to *process-oriented models* were set up, which reduce the initial activity of samples below 100 *pmC*. The result is a corrected age. Several correction models exist (Vogel 1970, Tamers 1975, Fontes & Garnier 1979, Maloszewski & Zuber 1991, Kalin 1999). A brief review was given in Clark and Fritz (1997). Other work have been done by Garrels and Christ (1965), Freeze and Cherry (1979), Stumm and Morgan (1996) and Drever (1997). The resulting ages are often compared to conventional ages, which are obtained with initial values of 100 *pmC*.

Weaver, *et al.* (1999) ruled out the possibility of carbon exchange during water-rock interactions since the TMG rocks usually do not contain any carbonates. The ^{13}C content of groundwater therefore remains close to those of the soil produced CO_2 ; usually around B22 to B18‰, as suggested by Talma *et al.* (1984).

2.5.2.3 Chlorine-36

Chlorine-36 is a radioactive isotope of chlorine whose application to hydrology has attracted much interest in the last decade. It is naturally produced by cosmic rays interacting with the atmospheric argon (^{40}Ar), and finds its way into the hydrological cycle either as dry fallout or in precipitation. Thermonuclear bomb testing of 1960s contributed significant amount of ^{36}Cl , thus elevating its concentrations above the natural atmospheric abundance. ^{36}Cl behaves conservatively in most hydrological environment, and like bomb tritium, it is useful in delineating recharge rates. However, unlike tritium, its use for dating modern groundwater is unrealistic. But with a half-life of 3001,000 years, ^{36}Cl is a useful tool in groundwater age determination in the range of 10^5 - 10^6 years. This isotope is not expected to yield realistic results in this study, and has been included with a purpose of highlighting its importance in groundwater studies.

2.5.3 Water-rock interactions (Isotopic Exchange)

The isotopic composition of groundwater is in most cases controlled by meteorological processes. There exist, however, some extreme geological environments where reaction between groundwater and the aquifer matrix or subsurface gases can modify the water's "meteoric" signature. Effects of such water-rock interaction dominate at high temperature and over geological time scales, but can also be observed in shallow groundwater flow systems at low temperature

According to Drever (1988) the deuterium content of many thermal waters is the same as that of the local precipitation, but the $\delta^{18}\text{O}$ is heavier than that of the local precipitation

The interpretation of the graph is that the waters are of meteoric origin, and the $\delta^{18}\text{O}$ has been shifted by isotopic exchange with rock. The reason why the $\delta^{18}\text{O}$ is shifted but not δD value is that rock contains large amount of oxygen, but little hydrogen. In a water-saturated rock, 50 % or more of the total oxygen in the system is in the rock, but almost all hydrogen in the system is in the water. In the isotopic exchange reaction the total mass of each isotope is conserved.

In geothermal system, some of ^{18}O is transferred from the rock to the water, and some ^{16}O from the water to the rock. The rock becomes isotopically lighter and the water becomes isotopically heavier. This is because there is so little hydrogen in the rock, however, that no change in the δD of the rock could be large enough to balance a significant change in the δD of the water. Thus alteration of rock in hydrothermal systems may cause a significant change in the $\delta^{18}\text{O}$ of the water, but usually no detectable change in the δD of the water (Drever, 1988).

In case of the TMG hot springs, water-rock interaction is unlikely to take place above about 70°C where mineral-water fractionations are such that any change in $\delta^{18}\text{O}$ value would have been to lower not to higher values with exchanged waters plotting to the left of the meteoric water line (Harris and Talma, 2002). Considering the recent work done by Harris and Diamond (2001) on TMG hot springs, it can then be argued that oxygen isotope exchange at such low temperatures is likely to have been sufficiently slow that water-rock interaction has no effect on isotope ratios of the water.

2.6 CONCLUSIONS

Deep groundwater occurrences are known in most parts of the world. The term deep groundwater is not easy to define as shown above that there is no distinct definition of deep groundwater. Methods of studying deep groundwater range from hydrogeochemistry, isotopes, temperatures and hydrodynamics. The definition depends on the objectives of the study. There are also many ways in which deep groundwater can be defined from shallow groundwater. In this context, deep groundwater is referred to as water that has circulated to significant depths and acquired at least 7 degree Celsius as compared to the recharge or source temperature. These waters are known to occur at the depth of more than 100 m in primary aquifers and up to 10-15 km in deep fractured hard rock/basement aquifers, where they are normally referred to as deep fluids (Manning and Ingebritsen, 1999). Due to high change in chemistry with depth in crystalline rocks (i.e. TDS and temperature), groundwater gradually changes into a hot saline called crustal deep-fluid (Stober and Bucher, 2005).

Deep groundwater has slow velocities and flow rates, leading to long residence times and therefore old water. Recharge of deep groundwater flow system is generally not of current climatic and topographic conditions.

These waters are characterized by high temperatures (greater than ambient air temperatures) and this is because of the geothermal heat they acquired as they descend to greater depth. The occurrence of these deep groundwaters is controlled by permeability which is further known to decrease with depth. In basement aquifers,

permeability gradually decreases to the so-called ductile-brittle zone (DBTZ). DBTZ depends on the composition and structure of the rocks, their water content and geothermal gradient, and is at 10-15 km (Manning and Ingebritsen, 1999).

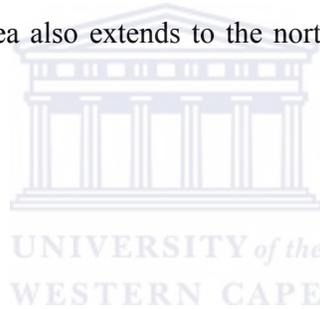
Below this level, water occurs only in isolated pores (Frost and Bucher, 1994; Stober and Bucher, 2005). This groundwater is termed connate water and is very old in age.



CHAPTER 3 STUDY AREA DESCRIPTION

3.1 GEOGRAPHICAL LOCATION AND EXTENT

The location of the study area is shown in Figure 3.1. Geographically, the area lies generally within an area of 19°44'15" to 19°46'20" longitude and 33°37' to 33°38'30" latitude. The large part of the study area lies within the Gouritz Water Management Area (WMA) and small portion forms part of the Breede WMA. The area is bordered by the Hex River on the western side, and by the Worcester Fault (WF) and the basement in the south, extends to the east next to Oudtshoorn where it is bordered by another river course. The area also extends to the north where it is bordered by the Cango fault (CF).



3.2 CLIMATE

The semi-arid climate of the Western Cape is characterized by warm dry summers and mild, moist winters. The climate of the region is largely influenced by the topography. The annual rainfall varies greatly within the region – from between 500 mm and 1700 mm on the Cape Peninsula, to between 500 mm and 800 mm on the Cape Flats, and ranges from 800 to over 2600 mm in the mountains. The Western Cape receives 80% of its annual rainfall during winter (May to September) and in the summer months (December to March) it could be as low as about 10 – 20mm and about 50mm in the mountains (Diamond and Harris, 2000). The Worcester and Ceres areas receive precipitation of > 2000 mm/a in the mountainous areas where the elevation also reaches >2000m. The average daily maximum temperatures vary from

minimum of about 28° C in mid-summer and 17° C in mid winter. Little snow is experienced during winter period.

In the Koo Valley area, the overall precipitation is minimal. Two automatic rainfall recorders were installed by DWAF in November 2003 on the mountains bordering the Koo Valley, near the Telkom towers on Rooihoogte in the north and Langeberg in the south. Table 3.1 below shows the rainfall data obtained during the period of December 2003 to April 2004. Recent records could not be found.

Table 3. 1: Comparison of Rainfall Figures (mm) of three Gauges in the Koo Valley

Month	Rooihoogte	Langberg	Tot-u-Diens
Dec-03	5	22	5
Jan-04	6	19	11
Feb-04	0	0	0
Mar-04	0	61	13
Apr-04	31	28	60
Cumulative	42	130	89

The table shows that for the five months since the automatic recorders were in operation, the Langeberge gets considerably more rain than Rooihoogte and Tot-u-Diens.

3.3 GEOLOGY

3.3.1 Regional Geology

The area of study lies largely in the Cape Fold Belt (CFB), which is composed of the rocks of the basement and the Cape Supergroup. These rocks include Malmersbury Group, Cape Granites, TMG, Bokkeveld Group, and Witteberg Group (see Figure 3.1). The detailed geology and stratigraphy of the CFB fall beyond the scope of this research (see Broquet, 1992; and De Beer, 1995; 2002).

The geology of the TMG is well understood. The TMG occurs within the Western and Eastern Cape Provinces of South Africa, extending just from north of Nieuwoudtville (see Fig.3.2) to Cape Agulhas and then eastwards to Algoa Bay, a linear distance outcrop of over 900km (Pietersen and Parsons, 2002). The TMG is divided into the Nardouw Subgroup, consisting of the Goudini, Skurweberg, Rietvlei/ Baviaanskloof Formations, and of the Piekenierskloof, Graafwater, Peninsula, Pakhuis and Cedarberg Formations. According to De Beer (2002a), the TMG rocks have been modified by the two major tectonic events, namely the Permo-Triassic Cape Orogeny and the fragmentation of south-western Gondwana during the Mesozoic.

Existence and occurrence of any groundwater system(s) in the hard-rock provinces in Southern Africa will indisputably be influenced and controlled by the structural history of the sub-continent.

3.3.2 Local geology

The study area lies on the fractured and folded rocks of the Cape Fold Belt (CFB) of the Cape Supergroup. The geology of the study area is shown in Figure 3.1. Lithostratigraphically, the area is comprised of the different rocks of the following groups (see Table 3.2):

- Cape Granite Suite
- Malmersbury
- Table Mountain
- Bokkeveld
- Witteberg



Table 3. 2. Lithostratigraphy of the study area (after De Beer, 1995; 2002).

GROUP	SUBGROUP	FORMATION	THICKNESS (M)	ROCK TYPES
WITTEBERG	WELTEVREDE	Blinkberg	250	Quartzitic sandstone, minor siltstone
		Wagendrift	330	Siltstone, mudstone and shale, micaceous, purple red brown
BOKKEVELD	Bidouw	Karopoort	90	Shale, siltstone
		Osberg	75	Siltstone, shale
		Klipbakkop	170	Siltstone, shale
		Wuppertal	90	Sandstone, siltstone
		Waboomberg	200	Siltstone, shale
	Ceres	Boplaas	80	Sandstone
		Tra-Tra	120	Shale, siltstone
		Hex River	100	Sandstone, Siltstone
		Voorstehoek	200	Shale, Siltstone
		Gamka	35	Sandstone, siltstone
		Gydo	160	Shale, siltstone
TABLE MOUNTAIN	Nardouw	Rietvlei	400	Feldsparthic quartz arenite, shaley Keerom Member at base
		Skurweberg	400	Quartz arenite
		Goudini	300	Red-brown-weathered arenite, minor siltstone, shale
		Cedarberg	80	Shale, siltstone
		Pakhuis	20	Sandy diamictite
		Peninsula	1600	Quartz arenite
MALMESBURY				Limestone, shale, subarkose
CAPE GRANITE SUITE		Robertson pluton		Granite

The rocks of Cape Granite Suite and the Malmesbury Groups form the basement of the study area, and are generally exposed on the southern side of the study area, where they form part of the study area boundary on the south. These outcrops are encountered just north of the Worcester Fault (WF).

The Table Mountain Group (TMG) rocks are the most dominant within the area of study. The TMG is quartz arenite complex (Weaver *et al.*, 1999). These rocks bound the area from the north and from the south. According to the recent study by De Beer (2002b), refer to Table 2.1; Piekenierskloof and Graafwater Formations are absent in the study area. The present formations include Peninsula, Pakhuis, Cedarberg, and all formations of the Nardouw Subgroup.

The Peninsula Formation, which is comprised of well-sorted, thicker-bedded, white quartz arenite, is separated from the Nardouw by the shaley Cedarberg and Pakhuis Formations. According to De Beer (2002a), the Nardouw Subgroup is a thick (maximum 1 200m) unit of sandstone that varies between quartz arenite, silty and feldspathic arenites, accompanied by minor interbedded conglomerate and shale. This lithological diversity, together with textural, grain size and bedding thickness differences, lead to pronounced differences in weathering, structural and hydrological characteristics.

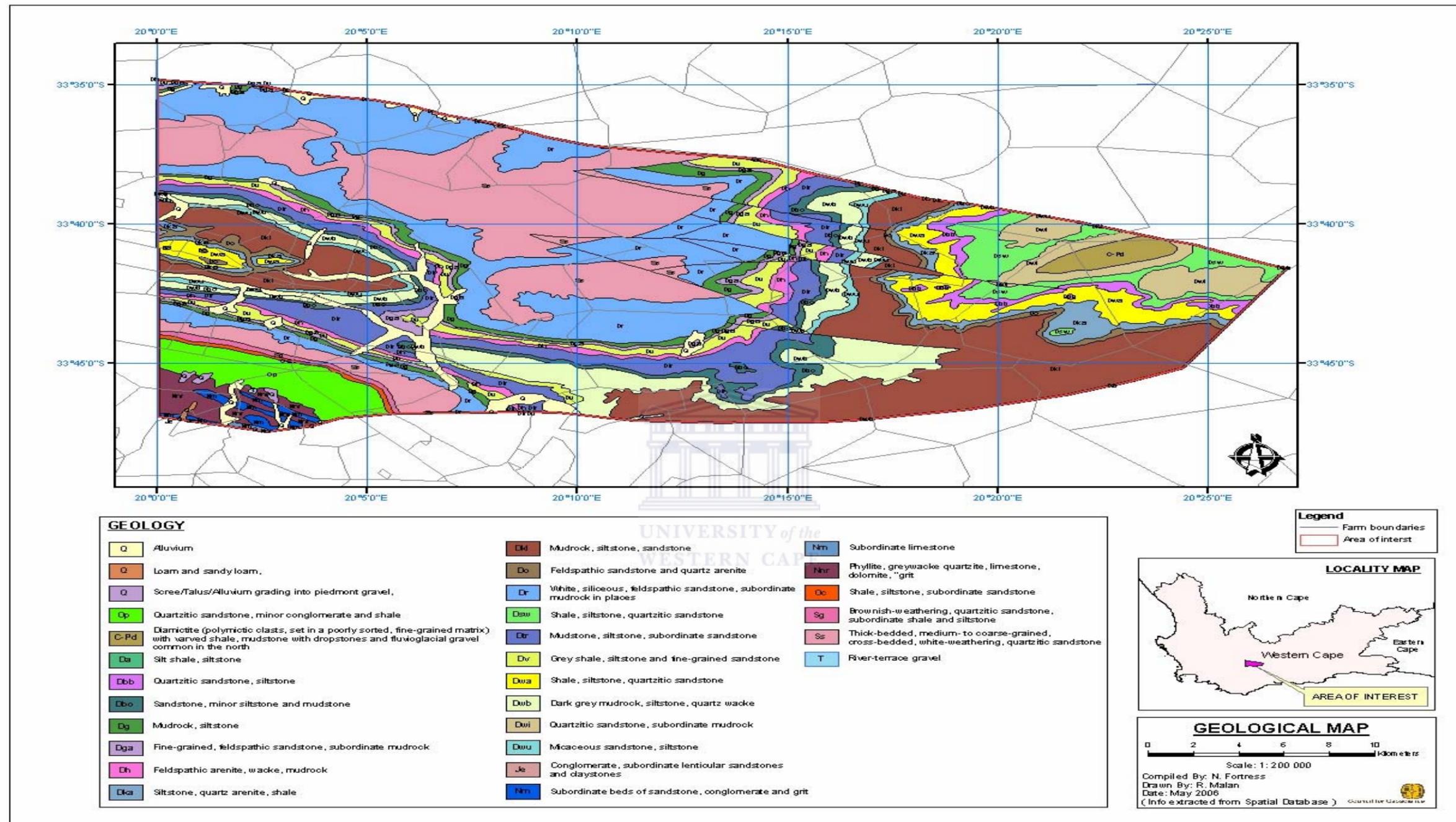


Figure 3. 1 Location and geology of the area

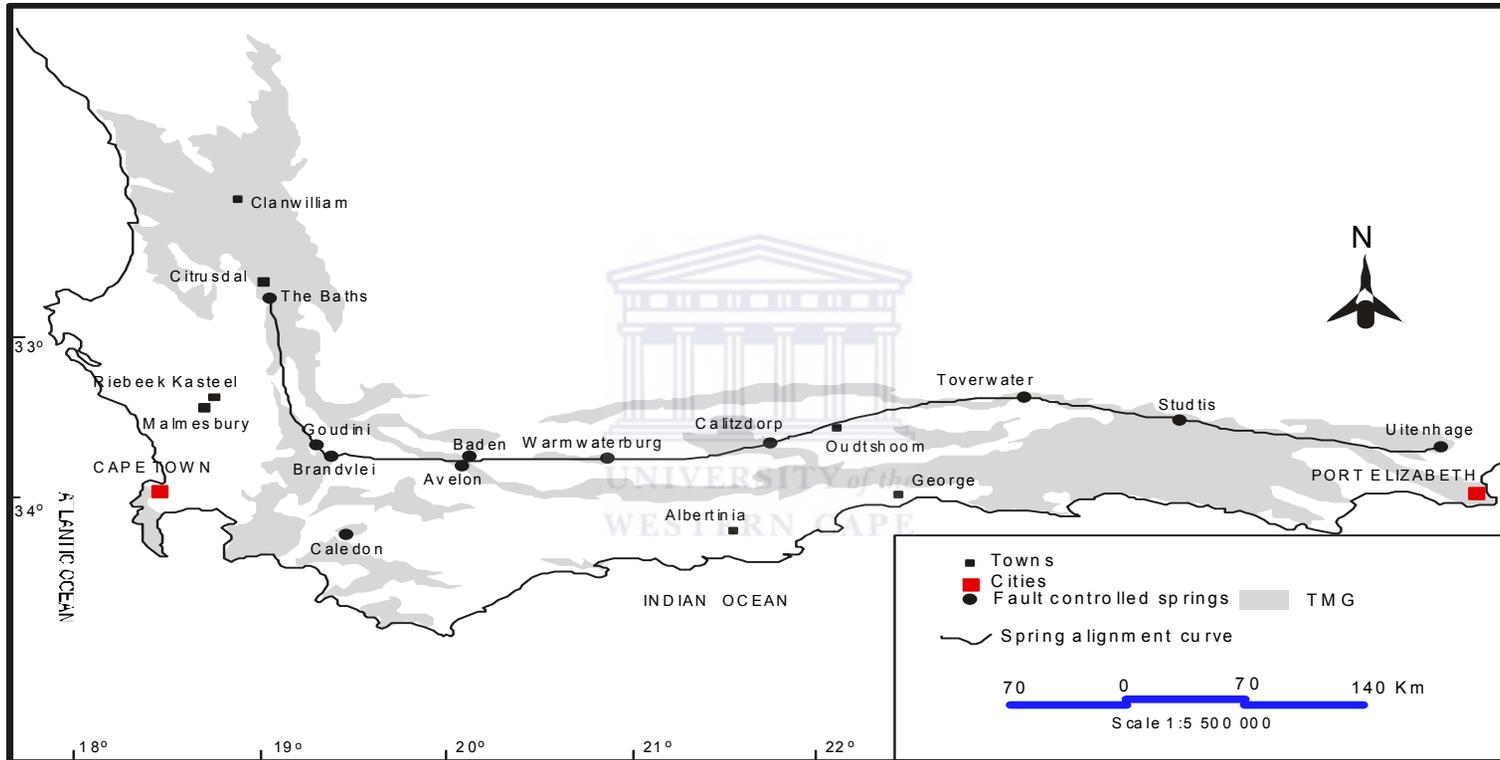
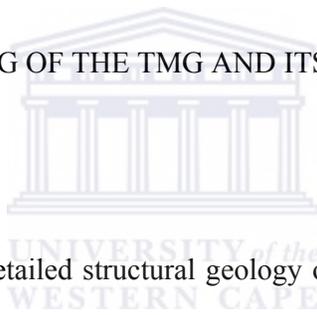


Figure 3. 2. Distribution of the TMG and locations of thermal springs (modified from De Beer , 2002).

The Skuweberg Formation of the Nardouw Subgroup was targeted during the drilling by SRK Consulting, and artesian borehole yield of about 30 ℓ/s were observed (De Beer, 2004. *pers.comm*).

The Bokkeveld Group overlies the TMG. This group is comprised of alternating shale and sandstone units formed by fluvial and shallow water delta processes. The argillaceous nature of this 1500-4000 m thick off-lapping, southward-thickening sequence contrasts strongly with the TMG (Broquet, 1992). De Beer (2002a) states that the contact between the Bokkeveld Group and the Nardouw Subgroup is usually abrupt.

3.4 STRUCTURAL SETTING OF THE TMG AND ITS INFLUENCE ON DEEP GROUNDWATER

The logo of the University of the Western Cape, featuring a classical building facade with columns and a pediment, with the text 'UNIVERSITY of the WESTERN CAPE' below it.

De Beer (2002a) gives the detailed structural geology of the TMG. According to De Beer, the presently exposed structure and thickness of the TMG rocks are the result of initial deposition within an east-trending basin along the southern and south-western Cape region, as modified by the two major tectonic events, namely the Permo-Triassic Cape Orogeny and the fragmentation of south-western Gondwana during the Mesozoic. The Cape Orogeny deformed a basement of previously deformed and metamorphosed Neoproterozoic rocks and the Cape Granite Suite, together with its cover sequence of Ordovician to Triassic rocks (Cape Supergroup and part of the Karoo Sequence). These deformation events of the CFB have resulted in two branches, namely the Western and Southern Branches. These branches meet at the Syntaxis (the highly deformed area). The Syntaxis comprises a complex zone of NE-

trending folds and faults stretching from Ceres in the north to Cape Hangklip on the coast (De Beer, 1992).

According to Kotze (2002), the TMG is composed of three types of fractures (depending on the scale of extent): 1) The first type is the small scale of the order of tens of meters deep. In the Kammanassie area, these fractures generally strike NW-SE, N-S and NE-SW, and correspond with the mechanical failure of rocks near the surface. These fractures control infiltration and circulation of water to shallow depths only, except where they are connected to the medium-scale fractures. The water flowing in these fractures are normally discharged as seasonal seepage in the mountain ranges at perched water tables during the winter months.

2) The second type of fractures is the intermediate or medium scale fractures related to the structural failures along major fault-lines, resulting from folding or later extensional fault movement. These fractures can penetrate to great depths (> 100 m) and are restricted to mountain catchments.

3) The third type is the largest fractures that are associated with deep-seated tectonic movements in the earth's crust. They allow large volumes of groundwater to penetrate and circulate to greater depths, and also provide a route for deep circulating groundwater under artesian pressure to the surface leading to the formation of thermal springs. The intermediate and regional groundwater flow paths are also visualised as inter-connected fracture sets along folded bedding planes on the catchments and geological basin scales, respectively (Kotze, 2002).

Sedimentary features such as grain size, matrix composition, sedimentary structures and formation thickness are inevitably modified by folding and faulting processes, the

end result of which will ultimately determine the structure and resultant hydrogeological potential of these rocks (Pietersen and Parsons, 2002).

A large percentage of its total thickness of >2000 m consists of quartzitic sandstones. These sandstones are of Ordovician to Silurian age (500 Million years) and because of their age and mild regional metamorphism, essentially possess zero primary hydraulic conductivity. However, due to combination of favourable factors, such as structure and climate, they form one of the major fractured aquifers in South Africa.

3.5 GEOHYDROLOGY

Over 30 major users of the TMG aquifer have been identified, ranging from municipalities to agriculture and it has major direct and indirect beneficial impacts on the hydrogeology of the Western and Eastern Cape. Borehole blow-yields of up to 120 ℓ/s have been recorded and hot-spring flows of 127 ℓ/s (Rosewarne, 2001).

3.5.1 Potential aquifer systems

Three geological groups that are found within the study area are regarded as aquifers. These include the Witteberg, Bokkeveld and the TMG Aquifer systems. However, they possess different hydrogeological characteristics. Lithology, stratigraphy, structural geological setting of these rocks largely influence groundwater occurrence in this area.

3.5.1.1 Table Mountain Group Aquifer system

The TMG in the study area is comprised of Peninsula, Pakhuis and Cedarberg Formations, and the Nardouw Subgroup. These geological formations and subgroup do not display similar or uniform aquifer parameters. Lithology and tectonic/structural controls mean that these parameters will vary considerably from place to place (Rosewarne, 2002). Meyer (2001) argued that an intricate network of fissures, joints, fractures and cavities largely governs infiltration, storage and transmission of groundwater in the largely competent and brittle-natured arenaceous units of the TMG.

Peninsula Formation, which attains a thickness of 1600m, and the Nardouw Subgroup are the main aquifers followed by the Bokkeveld. According to Kotze (2002), the Peninsula Aquifer is postulated to be purely fractured rock aquifer, i.e. rock mass consisting of large fractures and some matrix blocks with no micro-fissures. The system consists of interconnected network of fractures and the rock matrix, comprising the blocks surrounded by fractures, is impervious to flow. However, a good interconnection exists between smaller matrix fractures and larger scale fractures.

The Peninsula Aquifer generally constitutes the mountainous areas, which in turn influence precipitation to a significant extent. The fractured nature of the sandstones in these areas tends to favour groundwater recharge. Taking the orographic rainfall pattern, the elevation of the outcrop areas of the Peninsula Aquifer and its weathered and highly fractured surface into account, it is evident that the Peninsula Aquifer receives the higher amount of recharge (Ninham Shand, 2004).

Rosewarne (2002) reported Transmissivity and Storativity of 10 to 200 m²/d and 10⁻⁴ to 10⁻² for the Peninsula Formation and Nardouw Subgroup respectively, in the Klein Karoo area. These values were estimated using the Flow Characteristic (FC) Method. In the Hex River valley Transmissivity was estimated at 55 m²/d for the Rietvlei and Gydo Formations using Gringarten and Witherspoon Method. The bulk Storativity of 0.01 or 1 x 10⁻³ is a fair estimate for the Peninsula and Nardouw Formations (Rosewarne, 2002).

The Nardouw Aquifer has a scale effect introduced by lower fracture frequency and more ductile nature of the formation due to the presence of thicker shale layers. The scale effect refers to the different scale of heterogeneity introduced by micro-fractures in the matrix blocks themselves subdividing the matrix blocks. Alternatively, the rock is also referred to as fractured porous rocks (Kotze, 2002). According to Weaver *et al.* (1999), the boreholes drilled in the TMG are located entirely within this aquifer. The Skurweberg Formation was targeted for drilling during the Koo Valley Groundwater Scheme (De Beer, 2002b). Rosewarne and Weaver (2002) regard the Skurweberg Formation as a subaquifer with sufficient overall thickness and massive thick-bedded zones to support large-scale groundwater abstraction. As opposed to Peninsula Aquifer, the Nardouw Aquifer outcrops generally at lower elevations and receives much less direct recharge. Considerable amounts of indirect recharge are possible via fractures connecting Peninsula and Nardouw Aquifers.

The Cedarberg Formation, which acts as an aquitard, and the Pakhuis Formation separate the Peninsula and Nardouw Aquifers from each other.

Fracturing in the TMG may extend down to several hundred meters in many areas and deep groundwater circulation is one of the notable groundwater characteristics of the TMG (Meyer, 1999). Figure 3.3 shows that borehole yields are often greatest when the borehole is greater than 200m deep. Hartnady and Hay (2002a) confirmed that experience during drilling of BK3 and BK4 showed that yields increase dramatically below depth of 200 m. This is possibly related to a stress-generated confining zone within the uniformly quartzitic rock type (Hartnady and Hay, 2002a).

Groundwater associated with the TMG is characterized as being amongst the purest occurring in South Africa in terms of Electric Conductivity (ranging between 5 and 70 mS/m) and low Total Dissolved Solids (TDS). The quality of the water has been related to its source, namely the frontal systems which bring rain-bearing clouds from the Atlantic and Indian Oceans. However, the pH is as low as 5 which is very acidic and therefore corrosive. This pH range is common however, since streams fed from fynbos-dominated catchments in the Western Cape mountains typically have their pH between 4.5 and 6.5.

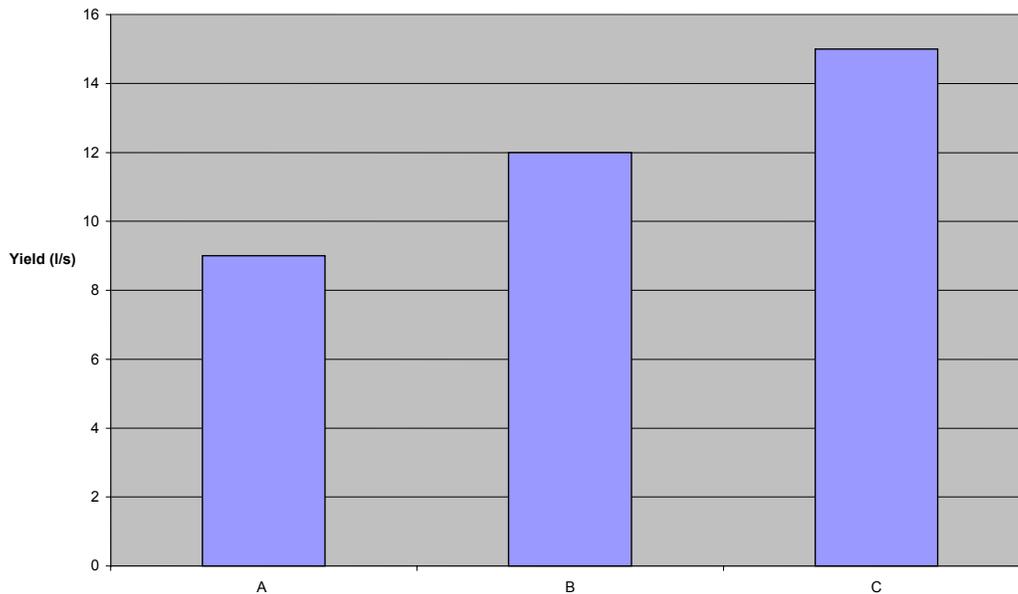


Figure 3.3. Yield increase with depth in the TMG (Weaver et al. 1999; Rosewarne, 2002)

Notes: A – Average yield of boreholes less than 99m deep

B – Average yield of boreholes between 99m and 199m deep

C – Average yield of boreholes exceeding 200m in depth

3.5.1.2 Bokkeveld Group (BG) Aquifer

Besides the argillaceous nature of this group, these rocks possess hydrogeological properties to some extent. According to De Beer (2002b) the occurrence of some cleavages in the BG rocks south of the Hex River valley is thought to have influence on the hydrogeological characteristics of the BG pelites and some silty sandstone, and shales that behave as aquitard to become aquifers.

De Beer (2002b) noted that many boreholes in the Koo Valley and surrounding area were drilled in BG rocks, where recharge is through fractures from the TMG, but they are overexploited. Groundwater in these rocks is generally of poor quality as compared to the water from the boreholes that tap water from the TMG aquifer systems. Meyer (1999) states that boreholes with maximum yields of 5 ℓ/s for sandstone-rich Ceres Subgroup, and seldom exceed 5 ℓ/s but usually below 1 ℓ/s in the sandstone-poor Bidouw Subgroup. Groundwater ECs varies between 30 to 400 mS/m, and are in excess of 400 mS/m for these two subgroups respectively. Groundwater in the BG is generally of a sodium-chloride nature (Meyer, 1999).

3.5.1.3 Witteberg Group Aquifer

A borehole yield analysis done by Meyer (2001), taking boreholes from all different geological units into account indicated that 30% of boreholes yield less than 0.5 ℓ/s , and 26% yield more than 5 ℓ/s . Brackish groundwater with ECs ranging between 200 and 700 mS/m can be expected in shale components, whereas groundwater from sandstone units generally possess ECs that range between 70 and 150 mS/m.

CHAPTER 4 DEEP GROUNDWATER OCCURRENCE IN THE TMG

4.1 INTRODUCTION

The occurrences of deep groundwater in the TMG have been confirmed, not only by the occurrence of thermal springs, but boreholes drilled in different localities. These localities include, amongst others, farm Boschkloof near Citrusdal (Hartnady, 1998; Hartnady and Hay, 2002a), Agter-Witzenberg (Weaver and Talma, 1999), and at the Blikhuis, Olifants River Valley (Hartnady and Hay, 2002b). Deep fresh water was also intersected by a borehole some 60 km offshore Port Elizabeth (Deep Offshore Water Research, 2006).

4.2 ORIGIN OF DEEP THERMAL GROUNDWATER

South African thermal springs are not associated with recent volcanic activity (Kent, 1949; Diamond and Harris, 2000) like in other volcanic active countries like Japan, but rather associated with meteoric water that is heated by geothermal gradients as it descends. Recent isotopic studies in the TMG area (i.e. Diamond and Harris, 2000) have confirmed low δD and $\delta^{18}O$ values than ambient rainfall and plot next to the local meteoric water line. The same conclusion of the source of the thermal was reached by Mazor and Verhagen (1983).

4.3 OCCURRENCE OF THERMAL SPRINGS IN THE TMG

The occurrence of the springs is the distinctive character of the fractured rocks of the Cape Supergroup. There are at least eleven thermal ($>25\text{ }^{\circ}\text{C}$) springs in the CFB. These springs are treated in this context as the manifestations of deep groundwater in the area. Two of them, namely Montagu and The Baden occur within the study area. The Montagu spring is also known as Avalon but referred in the text as Montagu. Three other springs occurring at the vicinity of the study area include Goudini, Brandvlei on the western side and Calitzdorp and Warmwaterberg in the eastern side. The general data including locations, temperatures and lithology is presented in Table 4.1.

They are all fault related and are situated in a TMG sandstone-Bokkeveld Shale relationship (Meyer, 2001; 2002). These thermal springs are known to be the discharge areas for deep circulating groundwater. Circulation depths estimates (Cavé and Clarke, 2003) have shown a minimum depth of about 1500 m.

Lithologically controlled, relatively shallow circulating springs are also present in the study area. According to Meyer (2001; 2002) these springs issue due to the presence of the impeding shale layers such as the Cedarberg Shale Formation. Yields of these springs are less constant and seasonal fluctuations are a distinctive feature of the TMG. The TMG is also known for the occurrence of the springs that seep from numerous small fractures and joints. They are evident during and shortly after rainy periods and are responsible for the myriad of springs in the TMG (Meyer, 2001).

Isotopic studies (e.g. Diamond Harris, 1999; 2000 and Weaver *et al.* 1999) of groundwater from these thermal springs have shown that the water is of meteoric origin. Table 4.2 provides the historical isotopic data of these springs.

4.4 RECHARGE AND DISCHARGE

Ground-water recharge of the TMG aquifer system is primarily from infiltration of precipitation and the associated flow away from the water table within the saturated zone. Recharge from precipitation at the water table varies seasonally because evapotranspiration, which can intercept infiltrating precipitation, varies seasonally.

The outcrops of the TMG in the area constitute much of the mountainous area, which in turn influence precipitation to a considerable extent (Meyer, 1999). A network of joints and fractures, which are a characteristic of the most rocks of the CFB, therefore control infiltration, storage and movement of groundwater. These fracturing and jointing is known to be extensive even to greater depths, and may influence the infiltration of rainwater to the deep groundwater flow system in the area.

The recharge to the Table Mountain Group aquifer system is believed to be in the range of 7% to 23% of Mean Annual Precipitation (MAP). Some researchers consider that recharge can be as high as 40% in the high lying areas that comprise predominantly bare rock (Ninham Shand, 2004).

Meyer (1999) argues that due to the fractured nature of the TMG sandstones in generally high rainfall regions, recharge is favourable and infiltration rates of up to 15% of the mean annual precipitation in certain area are not unrealistic.

A recent study (Wu & Xu, 2005) shows that recharge in the TMG is also influenced by snowmelt. The snowmelt is dominant in the catchments areas that are above 1000 m.a.m.s.l. They argue that recharge rates are influenced by fracture properties and are much lower than the snowmelt rate. The results of the experiment show that less than 13.6 % of infiltrating water recharges the TMG aquifer system. The snowmelt recharge of TMG is estimated at between 14.1 l/s to 15.0 l/s in the Kommissieskraal River Catchments (Wu & Xu, 2005). They also argued that the snowmelt impact on recharge in the TMG aquifer has a regional effect rather than local one.

A recharge estimation study in Kammanassie area has recently been conducted (Wu, 2005). This study shows that from water a balance point of view, to maintain the springs with a total flux of 383 ℓ/s in the TMG area, it has a recharge of 4.91 mm/a against relative outcrop of $2.46 \times 10^4 \text{ km}^2$ of the TMG at least, which is the base line of the TMG recharge. This study estimated recharge rate of about 5 % of the MAP.

The results of the stable isotopes analyses show that depletion of the δD and δO^{18} and this is an indication of recharge at high altitude (Diamond and Harris, 2000). Analysis of the stable isotopic concentrations of different samples from different altitude show recharge altitude for Brandvlei and Goudini springs of up to 1200 m (Saeze, 2005). It is therefore believed that recharge of deep groundwater flow system is gained from high-lying parts of the TMG where low-lying area recharge contributes much to the recharge of the shallow groundwater flow or local flow systems.

Discharge in the study area is normally in the form of seep zones, wetlands, seasonal low flow, and perennial cold springs in case of shallow circulating groundwaters whilst deep groundwater issue as thermal springs and thermal boreholes, where drilling intercept deep penetrating fracture or deep flow path.



Table 4. 1 General information on some of the major fault controlled thermal springs in the study area and vicinity (after Meyer, 2002 and Harris and Diamond, 2002).

Name of spring	Co-ordinate system		Temp. (°C)	Altitude (m)	Yield (ℓ/s)		Probable depth of circulation (m)*	Classification of thermal water# (Kent 1949)	Remarks
	South	East							
Goudini	33°24'00"	19°15'53"	40	290	11	7	1700	Thermal	Geological setting: situated on a NNW-striking fault in Peninsula Formation (TMG) sandstone near the contact with Bokkeveld Group shale. Dominant chemical determinants: sodium, sulphate and chloride. Utilisation: recreation.
Brandvlei	33°43'46"	19°24'58"	64	220	127	8	3600	Hyperthermal	Geological setting: situated on a NE-striking fault in Nardouw Subgroup (TMG) sandstone or close to the contact with Bokkeveld Group shale. Dominant chemical determinants: sodium and chloride. Utilisation: under-utilised, partly used for domestic supply.
Avalon (Montagu)	33°45'57"	20°07'02"	43	280	38	11	2000	Hyperthermal	Geological setting: situated on an E-striking fault. Nardouw Subgroup (TMG) sandstone faulted against Bokkeveld Group shale. Dominant chemical determinants: sodium and chloride. Utilisation: recreation.
Baden	33°42'20"	20°07'33"	38	280	37	10	1500	Thermal	Geological setting: situated on an ill-defined E-striking fault in Nardouw Subgroup (TMG) sandstone on the contact with Bokkeveld Group shale. Dominant chemical determinants: sodium and chloride. Utilisation: recreation.
Warmwater berg	33°45'57"	20°54'08"	45	500	9	26	2100	Hyperthermal	Geological setting: situated on a NE-striking fault. Nardouw Subgroup (TMG) sandstone faulted against Bokkeveld Group shale. Dominant chemical determinants: sodium and chloride. Utilisation: recreation.
* Average geothermal gradient of 1°C/80 m, and ambient water temperature of 18°C.									
# Italian classification (Kent, 1949).									

Table 4. 2. Hydrogen and oxygen isotope data for some thermal springs (After Diamond and Harris, 2000)

Year	Baden		Brandvlei		Goudini		Montagu				
	δD	$\delta^{18}O$	$\delta^{14}C$	$\Delta^{13}C$							
1995											
March			-33	-5.6			-33	-6.4			
April			-27	-5.5	-26	-4.4					
May			-32	-5.8							
June			-30	-5.0							
July			-28	-5.5							
August			-31	-6.1							
Sept.			-31	-5.6							
Oct			-28	-5.9							
1997	δD	$\delta^{18}O$									
March	-37	-6.9					-34	-6.4			
Mean	-37	-6.9	-30	-5.6	-26	-4.4	-33	-6.4			
M&V			-31	-6.1	-21	-4.8		-7.1	49.1	-21.3	

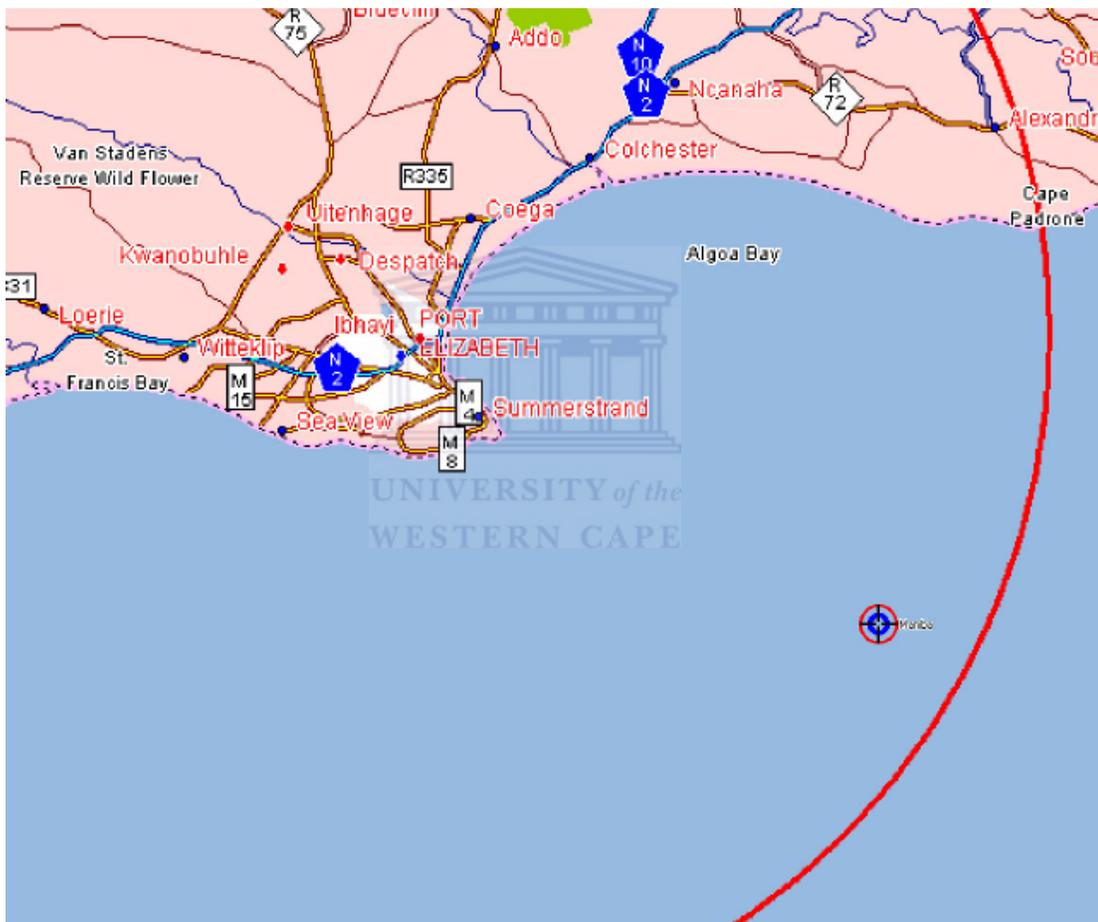
M&V Data from Mazor and Verhagen (1983) for samples collected in 1971-1972.

The table above shows low value (-21.3) ^{13}C that is an evidence of biogenic carbon. Since the concentration of ^{14}C at the recharge of this hydrothermal system is unknown, it may therefore be assumed that concentrations might have been as high as 80 to 90 pmC and that no significant exchange with free ^{14}C aquifer material has taken place. The concentration of 49 pmC for Montagu spring therefore suggests a turnover of several thousands of years (Mazor and Verhagen, 1983).

4.5 DEEP OFFSHORE WATER OCCURRENCE

Pioneer Natural Resources exploration well “Mamba”, drilled in 2000, intersected a fresh water resource some 60 km offshore Port Elizabeth. Location of the borehole is shown in Figure 4.1 and Figure 4.1(a). Discussions with geologists from Petroleum Agency of South Africa (PASA) and PetroSA confirm that wells drilled in the offshore basins of South Africa have encountered water. Some of this water may be ancient or connate water, some fresh being replenished through fissures and some very saline (Deep Offshore Water Research, 2006).

It is also believed that the offshore fractured basement (and the TMG) is likely to be a very large storehouse for water. No research has been undertaken to determine the accuracy of any assumptions, volumes or viability of exploring for or producing offshore fresh water. The Offshore Water Investments is currently pursuing the idea for a research of these aspects (Deep Offshore Water Research, 2006).



**Figure 4.1 Location of Mamba borehole, 60 km offshore Port Elizabeth
(Not to scale)**

The cross-sections presented in Figure 4.2 show that area is structural disturbed and the structures can penetrate to greater depths of at least 1000m. This is an indication

that the fresh water intersected is fracture controlled and might have been through the TMG where it plunge into sea in the Port Elizabeth area.





OFFSHORE SOUTH AFRICA - MAMBA PROSPECT GEOLOGIC CROSS-SECTION ILLUSTRATING PLAY CONCEPT AND COLUMNAR SECTION ILLUSTRATING EXPECTED LITHOLOGY

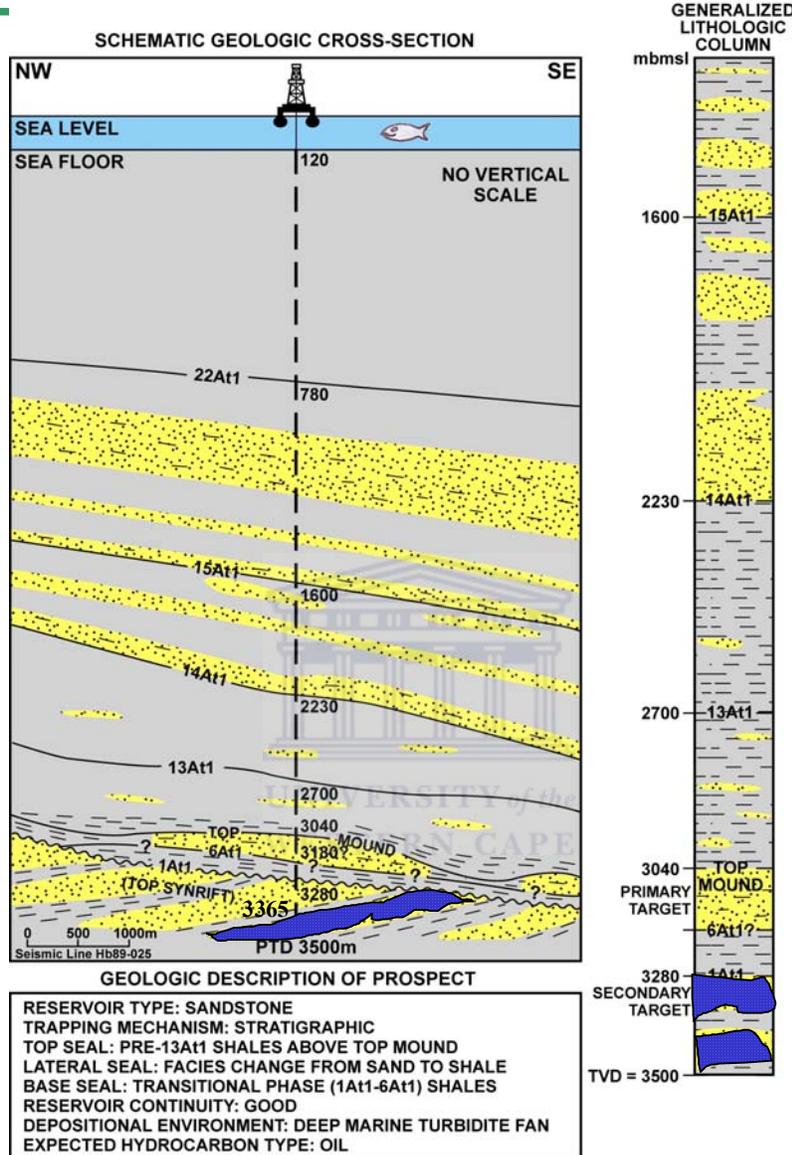


Figure 4.1(a) A cross section through Mamba borehole showing intersection of 185m fresh water column at -3365 meters

Figure 7

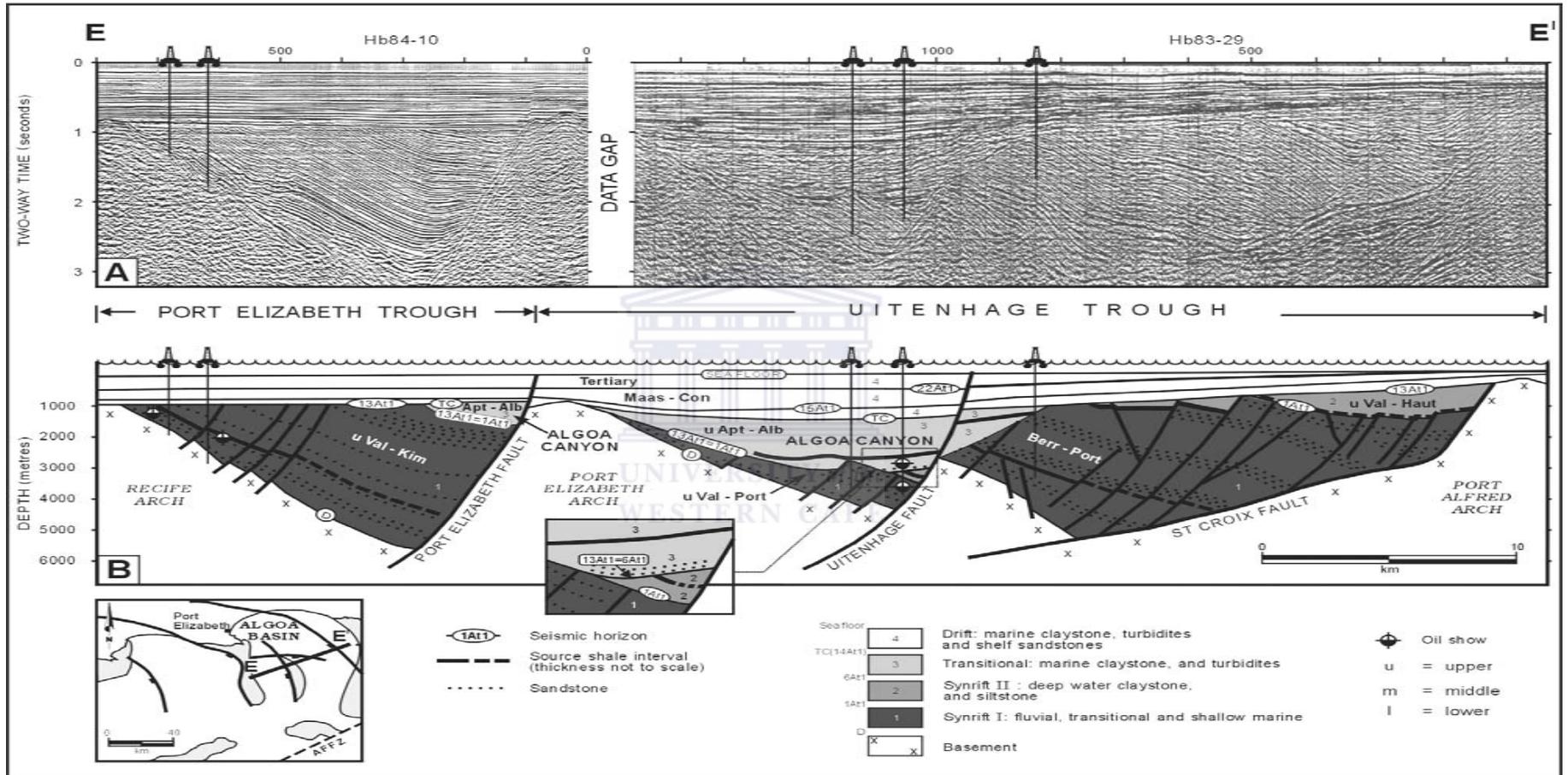


Figure 4.2 Geological cross-sections through Mamba borehole (left) to Shore (right).

4.6 GROUNDWATER TEMPERATURE AND CIRCULATION DEPTHS

According to Cavé and Clarke (2003) the depth of circulation may be estimated provided that the geothermal gradient and local mean air temperature of the area are known. In this case, the depth of circulation may be related to the subsurface temperatures estimated from the water temperature. The depth may then be calculated using equation:

$$D = \frac{T_{gw} - T_{avg}}{G}, \quad (4.1)$$

where G is the mean regional geothermal gradient ($^{\circ}\text{C}/\text{km}$), D is depth (km), T_{gw} is the groundwater temperature ($^{\circ}\text{C}$), and T_{avg} is the yearly average air temperature of the sampling locality ($^{\circ}\text{C}$). Similar equation has been used by Cavé and Clarke (2003) where T_{gw} is substituted by T_{calc} which is the temperature calculated by the plausible geothermometer ($^{\circ}\text{C}$). This depth estimate relies on the assumption that groundwater and the host rock are in local thermal equilibrium. This is generally justified because of the relatively high thermal conductivity of rock materials and slow, steady rates of subsurface fluid flow, but may be inappropriate for zones of rapid flow, for example geysers and or some fractured rock settings (Cavé and Clarke, 2003).

In cases where the circulation depth calculations are based on the increase of water temperature and geothermal gradients, two assumptions are generally made (Manga, 2001). The first assumption is that groundwater flow does not affect the subsurface temperature distribution and thus flow rates must be low. Secondly, the spring water rise rapidly enough that it does not cool.

The geothermal gradients of the CFB are not well established. Weaver and Talma (2000) used a geothermal gradient of 50 m/°C for calculation of circulation depths in three different localities in the TMG area, and their results are shown in Table 4.3.

Table 4. 3.Circulation depths inferred from borehole data (after Weaver and Talma, 2000)

Area	No. of Boreholes	Recharge temperature	Discharge Temperature	Change of temperature	Depth of Flow(m)
Piketberg	8	18.9	22.2	3.3	165
Oudtshoorn	11	16.2	23.4	7.2	360
Calitzdorp		16.2	22.9	6.7	335

Using the same geothermal gradient, the spring water temperatures of Montagu (43°C) and The Baden (38°C), and the mean air temperature of 18 °C, the water of these springs must have circulated at minimum depths of 1.3 km and 1.0 km respectively. Cavé and Clarke (2003) estimated a different depth of circulation of 1.6 km for the Montagu thermal spring using silica geothermometer (Chalcedony) and a geothermal gradient of 20 °C/km (50 m/°C).

4.7 CONSTRAINTS IN ASSESSING DEEP GROUNDWATER

The major constraint encountered when assessing deep flow in the TMG area, is the lack of data concerning the deep underground rock environment. These include lack of boreholes that penetrate the deep aquifer, the Peninsula aquifer, and unavailability of the information on the deep aquifer properties such as hydraulic conductivity and transmissivity at depth. These data sets, including recharge at depth, are crucial in

assessing the deep groundwater resource. However, it is believed that with the ongoing projects in the TMG area, these data will be available in the near future.



CHAPTER 5 CONCEPTUAL GROUNDWATER FLOW MODELLING

5.1 OBJECTIVES AND SCOPE

The numerical modeling plays an important role in the deep flow investigation. For this a ‘What-If’ type of modeling simulation approach has been adopted. This is an approach in which focus is placed on identifying sensitive parameters in the working scenarios. It is aimed at providing an understanding of the basic response of the flow system and visualizes most of the deep flow concepts discussed in the context of this thesis. The objectives of the groundwater modeling include the following:

- To conceptualization of groundwater flow of the area under consideration
- To do steady state particle tracking simulation
- To asses the influence of the Cedarberg on the flow system.
- To compare simulation results with interpretations of source, movement and travel times from existing data (i.e. isotopic data)

To achieve these objectives it is important to use a computer code for groundwater flow simulation on a steady state and that will allow particle tracking and visualization of flow paths. The current study employs TopoDrive (Hsieh, 2001) for that purpose (see 5.2.1.2 for code selection and data input). In this regard, TopoDrive is not intended to be a comprehensive modeling tool, but is designed for modeling at the exploratory or conceptual level, for visual demonstration, and for educational purposes (Hsieh, 2001).

According to Hsieh (2001) the theory that governs the use of TopoDrive is that for a two-dimensional vertical section, the boundaries of the flow domain (fig. 5.1) are as follows:

- The top boundary (AB) is the water table, which is assumed to lie close to land surface.
- The two vertical boundaries (BC and AD) and the bottom boundary (DC) are no flow boundaries.

The no-flow boundaries might represent ground-water flow divides or low-permeability bedrock that bounds the basin. Note that by specifying the position of the water table, it is assumed that the pattern of recharge and discharge is such that the water table is maintained at steady state.

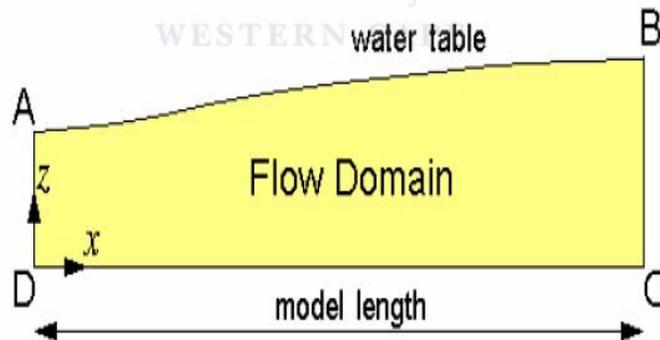


Figure 5.1. Flow domain of a topography-driven flow system.

Steady-state flow of ground water in the vertical section (fig. 5.1) is governed by the equation (5.1)

$$\frac{\partial}{\partial x} \left(K_{xx} \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial z} \left(K_{zz} \frac{\partial h}{\partial z} \right) = 0 \quad (5.1)$$

where h is hydraulic head, and K_{xx} and K_{zz} are the principal values of the hydraulic conductivity tensor. The principal directions are assumed to be parallel to the cartesian axes x and z . Assuming the position of the water table is known, the boundary condition along the water table (AB) is

$$h = z \quad (5.2)$$

where z is the elevation of the water table. Along the vertical boundaries BC and AD, the no-flow boundary condition is

$$\frac{\partial h}{\partial x} = 0 \quad (5.3)$$

Along bottom boundary DC, the no-flow boundary condition is

$$\frac{\partial h}{\partial z} = 0 \quad (5.4)$$

The computer model TopoDrive solves eq. (5.1) - (5.4) by the finite element method (see, for example, Huyakorn and Pinder, 1983). The flow domain is represented by a deformed rectangular mesh, and each quadrilateral cell is divided into two triangular elements. Linear basis functions are used in the finite element formulation. After

solving for hydraulic head h , the x and z components of the average interstitial velocity vector are computed by

$$v_x = -\frac{K_{xx}\partial h}{n \partial x} \quad (5.5)$$

$$v_z = -\frac{K_{zz}\partial h}{n \partial z} \quad (5.6)$$

where n is porosity. The velocity vectors are used for calculating flow paths and movement of fluid particles.

5.2 MODEL DEVELOPMENT



5.2.1 Model Set-up: Conceptual scenarios

5.2.1.1 Conceptual model

The first step in building up a groundwater flow model is to design a conceptual model (see fig. 5.1). The first phase in conceptualization comprised of layer schematization. Different layers were schematized according to the geological and hydrogeological properties of the study area. For the simplicity purpose, all the model area assumes a system comprised of three layers; Nardouw and Peninsula Aquifers, Cedarberg which is an aquitard (see fig. 5.2).

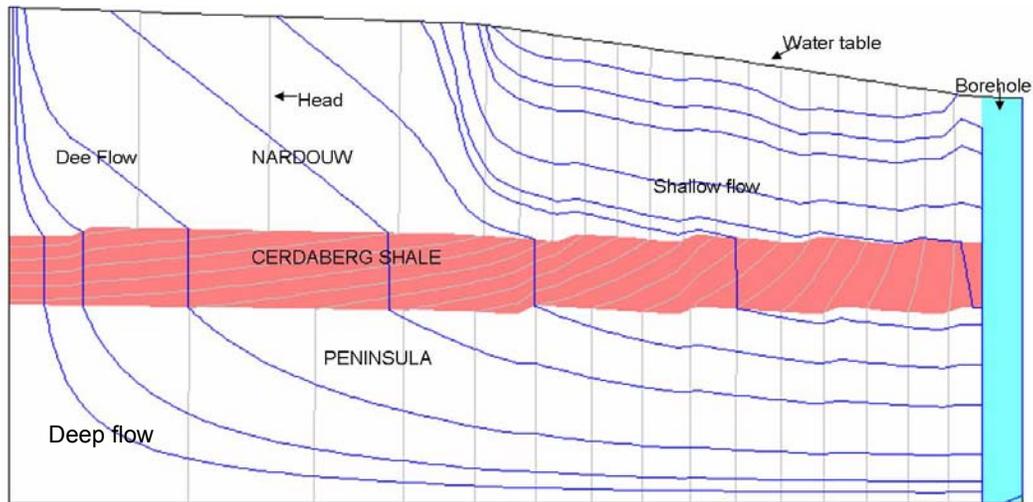


Figure 5. 2. A schematic representation of the conceptual model (Note:blue lines are flow lines)

5.2.1.2 Code selection and Data input

The package TopoDrive (Hsieh, 2001) was used to develop a groundwater flow simulation model and particle tracking. The model is based on a two-dimensional finite difference approach. The TopoDrive, is designed to simulate two ground-water processes: topography-driven flow and transport of fluid particles. In both cases, the flow is under steady state. The purpose of the model is to provide interactive simulation and visualization capabilities that enable the user to easily and quickly explore model behavior, and thereby better understand ground-water flow processes. In this regard, TopoDrive is not intended to be comprehensive modeling tool, but is designed for modeling at the exploratory or conceptual level, and for visual demonstration (Hsieh, 2001).

The following data input is required for the steady state groundwater flow modeling created by TopoDrive package:

- Model length/ geometry and vertical exaggeration
- Mesh (columns and rows)
- Hydraulic conductivity (K)
- Groundwater table
- Porosity
- Discharge and recharge rates

5.2.1.3 Boundary conditions

The top boundary is the water table, which is assumed to lie close to land surface. The two vertical boundaries and the bottom boundary are no flow boundaries. The no-flow boundaries might represent ground-water flow divides or low-permeability bedrock. It is assumed that the boundaries of the model area are zero flow boundaries ($Q = 0$).

5.2.2 Model domain

5.2.2.1 Model length, vertical exaggeration

The model length is 4000 m and the vertical exaggeration is 10. The grid of the model area has been designed as a grid with squares cells. The model area has been discretized with 50 columns and 50 rows to avoid long computational time.

5.2.2.2 Groundwater table

The water table is drawn in top left corner of model area, with high head and lowers towards the right top corner. This is to allow or direct water movement from left to right.

5.2.2.3 Hydraulic conductivity

TopoDrive offers five sets of hydraulic conductivity (m/s) and porosity (%) values are available for assignment to model elements. Each set is represented by a color. Default values are initially provided, but three of these values were altered in the edit boxes. Hydraulic conductivity of $K = 0.01$ and $K < 0.01$ were used. The former value was assumed for Nardouw and Peninsula (white) aquifers and the latter for Cedarberg aquitards (red) and $K=1000$ was allocated for a fully penetrating borehole (blue) (see fig. 5.2). Consequently, these values were changed during multiple simulation scenarios, especially K for the aquitard, in order to assess the influence of the aquitard on the deep flow.

5.2.2.4 Porosity

A constant value of $n = 0.20$ has been assumed for the porosity of the aquifers. Although this parameter is not needed for steady-state flow modeling, it is required for particle-tracking with TopoDrive.

5.2.2.5 Hydrological stress

5.2.2.5.1 Recharge through percolation

Groundwater recharge is assumed to take place at the water table. It was also assumed that contributions from other sources such as base flow and stream discharge do not have major significance in the flow system, and therefore ignored.

5.2.2.5.2 Well/spring discharge

Discharge in the study area is normally in the form of seep zones, wetlands, seasonal low flow and cold springs in case of shallow circulating system whilst deep groundwater issue as thermal springs and thermal boreholes, where drilling intersect deep penetrating fracture or deep flow system. To compensate for this discharge, a zone, representing a borehole (blue zone on the model area), is allocated a high hydraulic conductivity of 1000 m/d. water flows from high recharge area (top left) is topographically directed towards the borehole on the far top right zone with high hydraulic conductivity.

5.3 PARTICLE TRACKING SIMULATION

5.3.1 Purpose and scope

The main purpose of particle tracking in this study is to assess the groundwater paths and travel times. Particles were introduced on the borehole and tracked backwards to where recharge is believed to come from. Particle tracking was run on a steady state

groundwater flow model developed by TopoDrive. The program automatically calculates travel times to the discharge area. Results of particle tracking are discussed in the following section.

The particle tracker uses a regional-scale ground-water flow model that cannot account for all local flow effects or anomalies (Hinkle and Snyder, 1993). For instance, local vertical groundwater flow induced by pumping from wells, or resulting from annular flow or inter-aquifer flow through existing wells, is not accounted for by the flow model or particle tracker

5.3.2 Simulation scenarios

Particle tracking were performed on a different scenarios. Some data values were varied through multiple model runs. The aquifer parameters changed are K of the confining layer. The depth of the borehole was changed, and the fracture was introduced. This was done with the aim of assessing the influence of the confining layer on the deep flow. The hypothesis was that K of the Cedarberg (C/S) aquitard will have influence on deep flow. Various assumed scenarios are described below:

5.3.2.1 Scenario 1

The model parameters used in this scenario are the ones described in section 5.2. The only change is that the partially penetrating borehole is introduced and K of C/S is $1.0E - 4$ (see fig. 5.3).

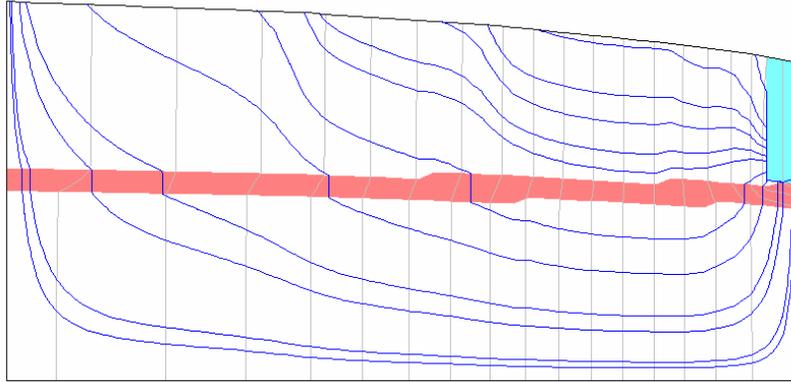


Figure 5.3 Simulation results for scenario 1.

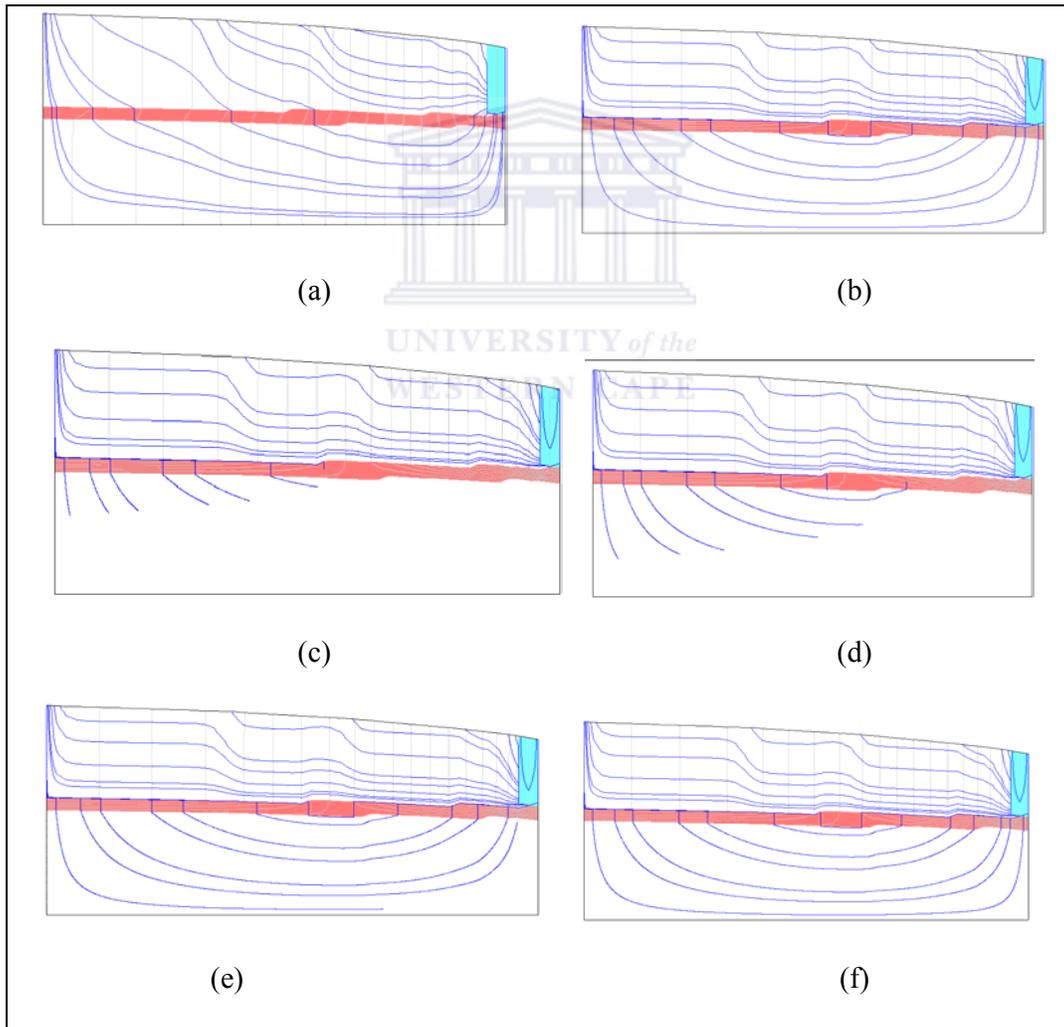


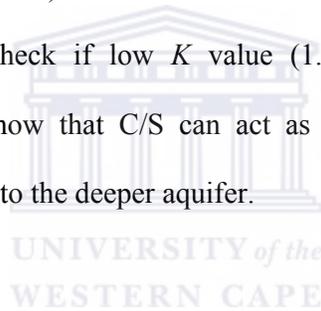
Figure 5.3.1 Simulation results of scenario 1 with variable K of C/S [i.e. (a) $K = 1.0E -6$, (b) $K = 1.0E -8$] and travel times with $K=1.0E -8$ [i.e. (c) 37 years, (d) 80 years, (e) 269 years, and (f) 405 years].

5.3.2.2 Scenario 2

Simulation results of this scenario are presented in Figures 5.3.2 (a) – (c). The simulation in this scenario maintained the assumption made in scenario 1, but the fully penetrating borehole is introduced. The objective of this scenario is to assess the change in travel times caused by the fully penetrating borehole.

5.3.2.3 Scenario 3

In this scenario, the model parameters are still kept in scenario 2. The only change is the impermeable C/S (fig. 5.4.a) and introduction of high transmissive fracture (see fig.5.4.b). The aim is to check if low K value ($1.0E-15$) of C/S can make it impermeable. The results show that C/S can act as a flow barrier from shallow groundwater to flow through to the deeper aquifer.



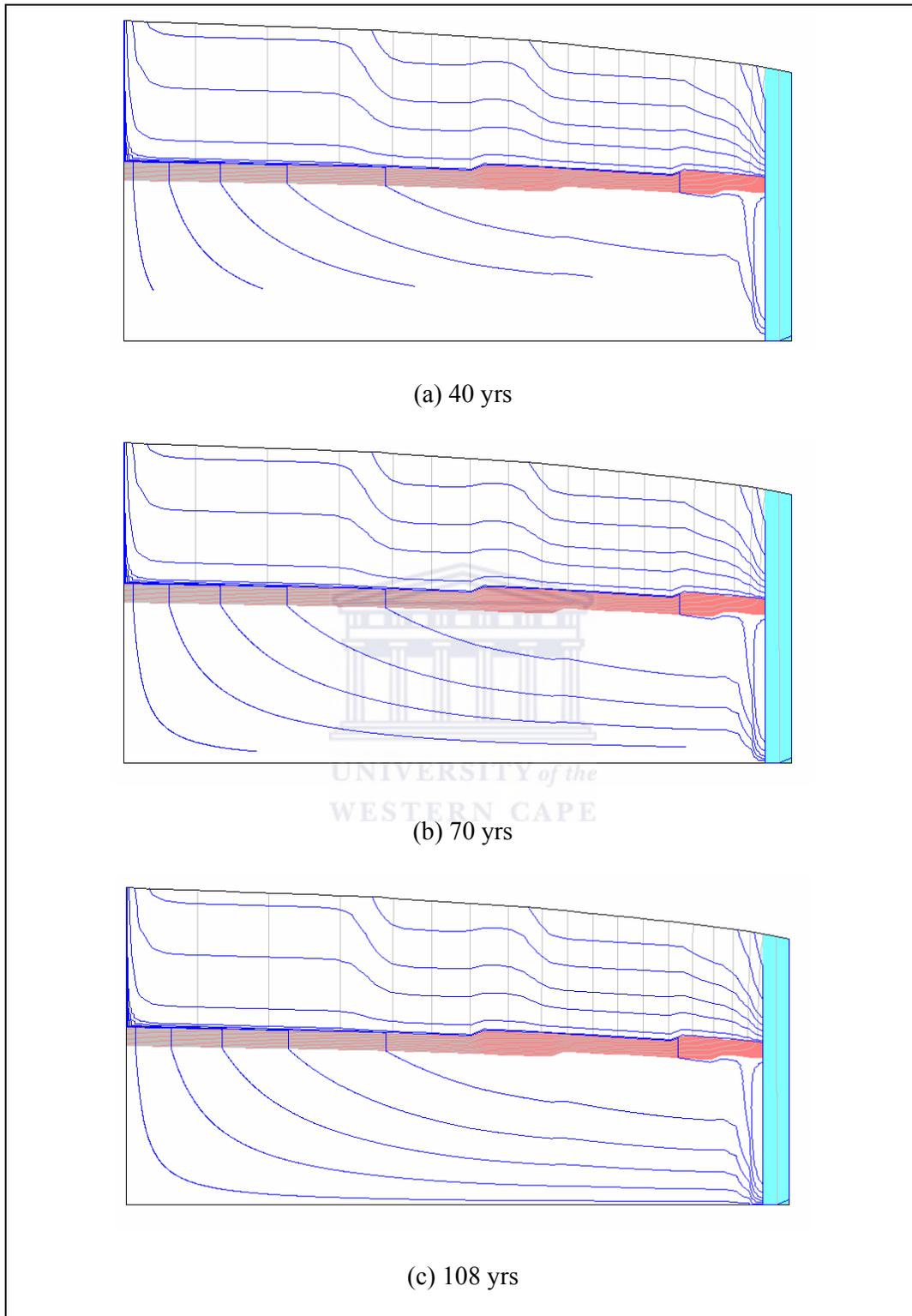


Figure 5.3.2 Simulation results for scenario 2 showing ages when a fully penetrating borehole is introduced.

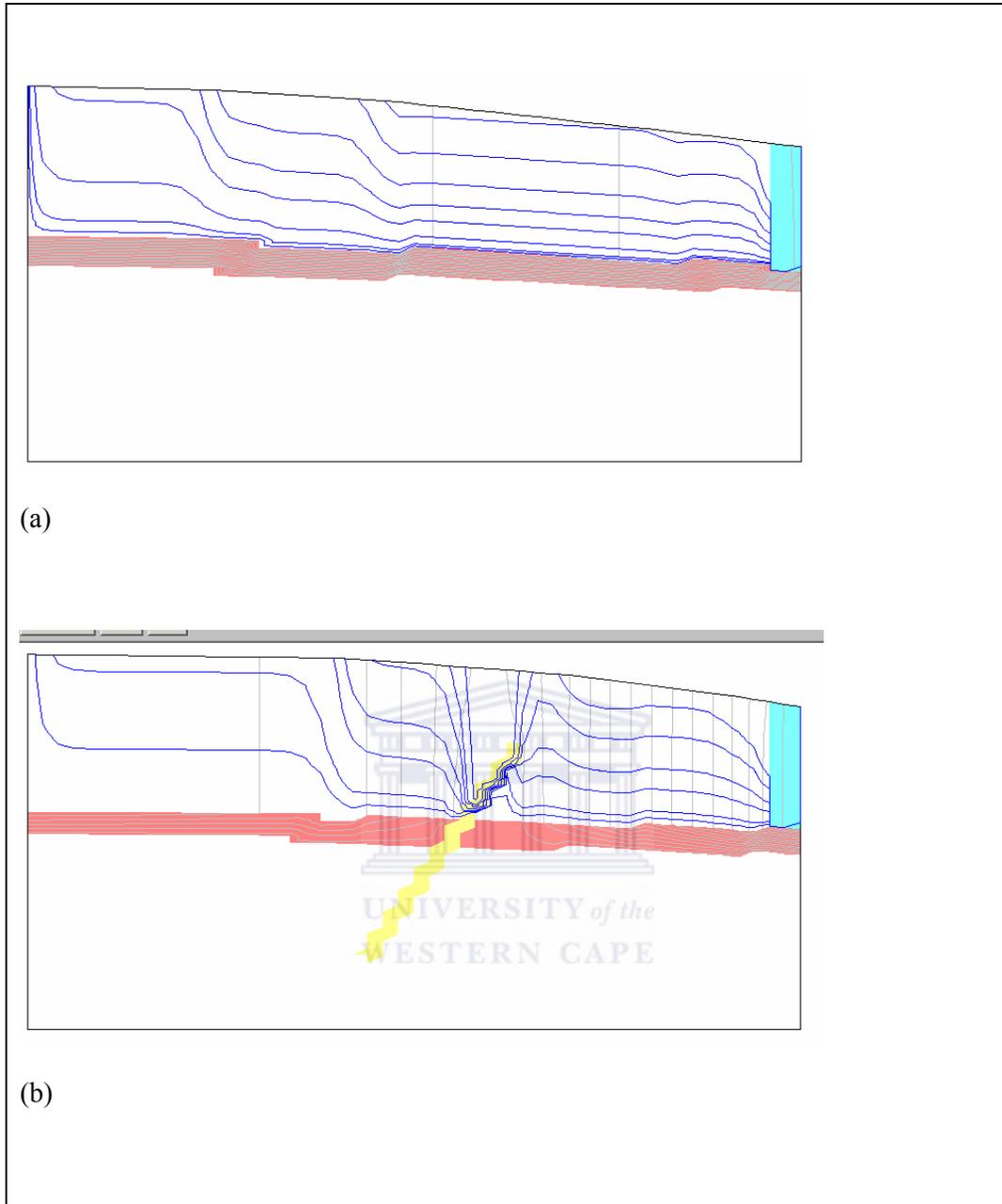


Figure 5.4 Simulation results for scenario 3 showing impermeable C/S (a) and a fracture (yellow) in (b).

5.4 RESULTS AND DISCUSSIONS

5.4.1 Flow paths analysis

Particles have been introduced mostly on the borehole (discharge area) and backward and forward tracking were done. Particles tend to emulate the direction of groundwater flow as controlled by the topography as expected. That is particles flow source, from high-lying areas towards the valley where they discharge at the drain.

The model results (senario1) show that the borehole gets water from both the upper shallow and the lower deep aquifers (see fig. 5.3.1.a, and b). Over 70 % of the groundwater will flow in the shallow aquifer and is directed towards the borehole before it recharges the deep seated aquifer. This might be caused by the hydraulic gradient together with the higher hydraulic conductivity of the upper shallow aquifer and compared to the underlying layer.

5.4.2 Groundwater age or travel times

Groundwater age is complex parameter to estimate. This is because it is normally estimated by analysis of the water sample from a borehole or a spring. These samples are mixture of water of various ages. This means that the particle tracking results cannot show the exact ages as estimated from other techniques, including carbon-14.

Travel times were also checked under assumptions made in section 5.3.2 above and results show that with K value of $1.0E - 8$, the particles on the shallow aquifer reaches the borehole after 37 years(yrs) [see fig. 5.3.1(c)]. This is the longest period taken by particles in this zone, whilst the particles in the lower zone (deep aquifer Peninsula),

particles that managed to flow through the C/S, almost doubled this period to reach the borehole. This situation is illustrated in Figures 5.3.1 (d) 80 yrs, (e) 269 yrs and (f) 405 yrs, which is the latest. This cause for this could be that the low permeability, influenced by the thickness, of the C/S, causes a delay when particles move deeper through the C/S. Another reason could be the partially penetrating borehole; that's particles are forced to flow through the C/S again from the deeper aquifer up to the discharge area, the borehole.

The simulation results for scenario 2 show that few particles from the deep aquifer reach the borehole after a short period of about 40 yrs (fig 5.3.2 a.). This means that the C/S delays water for about 10 yrs, and this is based on the period taken by particles from shallow zone at scenario 1.

The final arrival of all particles introduced takes only 108 yrs (fig. 5.3.2.c.) as compared to 405 yrs in scenario 1. This is because of the high transmissive zone introduced that taps water directly from the deep aquifer. That is groundwater will flow through the C/S and move directly towards the borehole without passing through the C/S again.

This analysis demonstrates the high level of uncertainty involved in calculating travel times from the recharge given our current limited knowledge of the appropriate parameter values.

5.5 CONCEPTUAL DEEP GROUNDWATER FLOW MODEL

Recharge of the study area occurs at high altitude particularly at the Waboomsberge and Langeberge mountains. The recharge area is covered by the rocks of the Nardouw Subgroup which are highly fractured and weathered. Percolation of water is dependent on the shallow, small-scale fracture systems. These fractures are partly closed by clayey material and this lead to the discharge of infiltrating water in the form of seepages, inflow and shallow cold springs. The other influencing factor to this discharge is the gravitational force.

The small-scale fractures, in some cases, are connected to the deep and/ or regional faults through intermediate fracture system.

Based on the groundwater temperatures at springs, there are two types of springs; cold springs (<25 °C) and thermal springs, those discharging water with temperatures greater than 25 °C. Cold springs are generally influenced by local or shallow flow systems. In the TMG area, this flow is dependent on the small-scale fracture systems and depth of weathering.

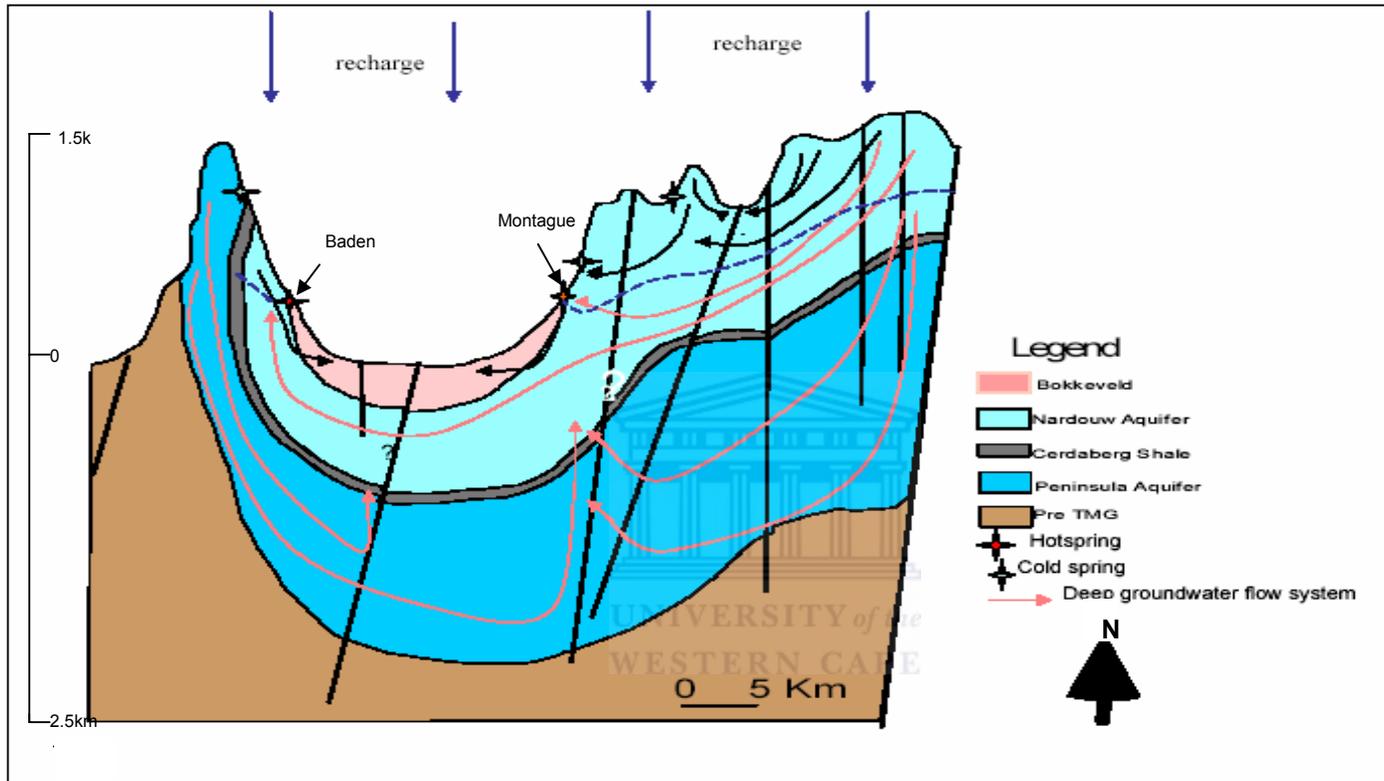


Figure 5.5. Schematic conceptual model showing deep groundwater flow system

In areas where these small-scale fracture systems are connected to medium-scale fractures, infiltrating water can be directed to intermediate flow systems. These intermediate systems have been confirmed by the isotopic study conducted in the Agter-Witzenberg valley (Weaver *et al.*, 1999). The groundwater circulating in these systems are often encountered in different parts of the TMG area in boreholes with elevated water temperature (i.e. 20 – 24 °C). Hartnady and Hay (2002b) reported some of the boreholes with elevated temperature in the Blikhuis area.

The deep penetrating faults, fed with water by the intermediate fractures, direct water to the deep parts of the TMG, particularly the Peninsula Formation (regional groundwater system). Groundwater then circulates to great depths (> 1000 m). The permeable faults cutting the regional aquifer system and small fractures that develop around them provide deep circulating water with an upward route to the surface where they discharge as thermal springs, or in the form of an artesian borehole. The discharge will be at lower altitude as compared to the recharge area.

CHAPTER 6 CONCLUSIONS AND RECOMMENDATIONS

6.1 DEFINITION OF DEEP GROUNDWATER

This study has found that there are many classification schemes that are used worldwide to distinguish deep groundwater flow from shallow groundwater flow. These schemes are guided by the objectives of the study. For example, for pollution studies 30 m might be considered deep whereas in researching fractured basement aquifers deep groundwater system may extend to the depths of several thousand meters.

These schemes are based on the number of techniques including hydrochemistry (i.e. basic chemistry and isotopes), temperature and some hydrogeological concepts such as confined and unconfined aquifers. These two hydrogeological concepts are normally considered as deep and shallow groundwater systems respectively.

Temperature, on the other hand, is used as measure of the circulation depths of descending groundwater where it acquires geothermal heat as it circulates to greater depths. The current study adopts temperature as a dividing line between shallow and deep groundwater flow systems. Thus deep groundwater will be the water that has circulated to deeper depths and acquired 7 °C as compared to the recharge or source temperature. These waters are known to occur at depth of more than 100 m in primary aquifers and up to 10-15 km in deep fractured hard rock/basement aquifers, where they are normally referred to as deep fluids (Manning and Ingebritsen, 1999).

Geothermal gradients, on which the adopted classification scheme is based, are not well established in the study area. It is therefore recommended that further studies be done on the distribution of geothermal heat, and therefore enhancing the level of confidence in application of proposed classification scheme. The heat exchange with groundwater flow in the aquifer system should also be investigated.

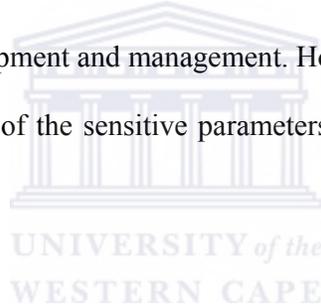
6.2 CORRELATION BETWEEN SIMULATION RESULTS AND EXISTING DATA

The groundwater flow model of the study area has been developed. During simulation, some parameters were varied through multiple runs to best characterise the site.

The origin or recharge area as explained in conceptual model in general recommendations is supported by the directions of the particles from point source towards the area of high hydraulic conductivity, the borehole. It has been concluded from recent studies in the TMG (i.e. Weaver et al. 1999; Weaver and Talma, 2000), that deep groundwater in the area possesses age of few hundreds to thousands years as indicated by carbon-14 concentration. The Carbon-14 concentration of 49 pMC of Montague spring is one example. Hartnady and Hay (2002b) reached the same conclusions after the study conducted in the Ceres area. It is shown in simulation results (see fig.5.3.1) that C/S, amongst other factors, has an influence on the travel times. For example, the shallow groundwater takes 30 years, latest, to reach the discharge point whereas it takes 60 years for the deep groundwater. However, travel times from simulation results are not enough compared to estimates made by authors

who conducted isotopic studies (i.e. several thousand years of age). The reason for this may be the fact that the groundwater model presented here does not account for all the complexities of the fractured rock aquifer. These include structural setting, distance from recharge to discharge point, and permeabilities of the aquifers.

The conclusions presented here are based on the little data that was used during the development of the groundwater flow model. The results give a basic conceptual understanding of the response of the flow system under assumed scenarios, and therefore the model is not comprehensive. This model cannot be used for pollution studies. However, it is the feeling of the author that more data should be gathered to test the validity of this model before it could be used as a predictive tool in groundwater resource development and management. However, the model results give a preliminary understanding of the sensitive parameters of groundwater flow system under investigation.



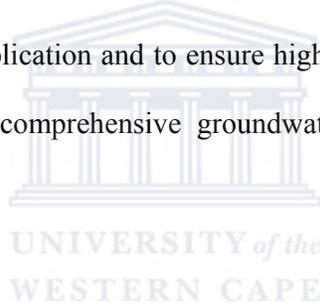
6.3 GENERAL RECOMMENDATIONS

There is still a need for further detailed research into the following aspects of deep groundwater:

- 1) The investigation of hydraulic properties of the deep fractured underground rock mass
- 2) The distribution of recharge with depths and/or the recharge mechanisms of the deep seated aquifer
- 3) Implications of using deep groundwater for water supply. This should include

- a) Assessment of the change of hydrodynamic regime due to abstraction of deep groundwater and,
 - b) Protection area for deep groundwater systems
- 4) Feasibility study on the potential of using deep offshore fresh water resources for commercial usage
 - 5) Detailed numerical groundwater flow modelling.
 - 6) Detailed study on the distribution of geothermal gradients.

It is however believed that more hydraulic data will be available since there are many ongoing groundwater projects in the TMG area. Because most of these projects are running independently, it is recommended that a more organised data flow structure be implemented to avoid duplication and to ensure high quality data that can be used in the development of the comprehensive groundwater flow model of the TMG aquifer system.



REFERENCES

Almen, K.E., Andersson, O., Fridh, B., Gustafsson, E., Nilsson, G. and Axelsen, K. (1986): Site investigation equipment for geological, geophysical, hydrogeological and hydrochemical characterization, SKB Tech. Rep. 86-16, SKB, Stockholm.

Bense V.F., Van Balen R.T. and De Vries J.J. (2003). The impacts of faults on the hydrogeological conditions in the Roer Valley Rift System: an overview. *Netherlands Journal of Geosciences/ Geologie en Mijnbouw* 82 (1): 41-54.

Broquet, C.A.M. (1992). A sedimentary record of the Cape Supergroup: A review. In: M.J De Wit and I.G.D Ramsome (eds.), *Inversion tectonics of the Cape Fold Belt, Karoo and Cretaceous Basins of Southern Africa*. Balkema, Rotterdam. ISBN 90 5410 0478.

Cave, L. and Clarke, S. (2003). *A Feasibility Study on the Use of Chemical Geothermometers for Tracing Deep Groundwater Flow*. WRC Report No. 1331/1/03.

Cherry, J.A & Freeze, R.A (1979). *Groundwater*. Englewood Cliffs: Prentice Hall.

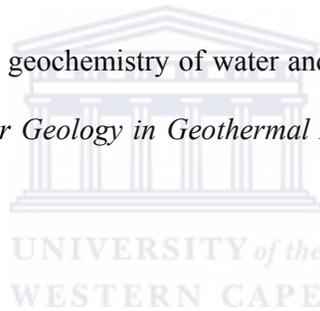
Christodoulou, T., Leontiadis, I.L., Morfis, A., Payne, B.R., Tzimourtas, S. (1993). Isotope hydrology study of the Axios River plain in northern Greece. *J. Hydrol.* 146, 391–404.

Cindrich, R.B. and Gudmundssori, J.S., (1984). Application of Stable Isotope Techniques to Tracing Recharge of Geothermal Systems on the Reykjanes Peninsula in Iceland, Department of Energy Contract No. DE-AT03-80SF11459 by the Department of Petroleum Engineering, Stanford University, California.

Clark, I.D., and Fritz, P, (1997). Environmental isotopes in hydrogeology: New York, Lewis Publishers, 328 p.

Craig, H. (1961). Isotopic variation in meteoric waters: *Science*, v. 133, p. 1,702-1,703.

Craig H. (1963). The isotope geochemistry of water and carbon in geothermal areas . In Tongiorgi E. (ed.) *Nuclear Geology in Geothermal Areas* , pp. 17– 53, Consiglio Nazionale delle Ricerche.



Davidson, C.C., Brown, R.A., Everitt, R.A., Gascoyne, M., Lodha, G.S., Martin, C.D., Soonawala, N.M., Stevenson, D.R., Thorne, G.A. and Whitaker, S.H. (1994): The disposal of Canada's Nuclear Fuel Waste, Site Screening and Site Evaluation Technology, AECL-10713, COG-93-3, p.154.

De Beer, C.H. (1992). Structural evolution of the Cape Fold Belt syntaxis and its influence on the syntectonic sedimentation in the SW Karoo Basin. In: M.J De Wit and I.G.D Ramsome (eds.), Inversion tectonics of the Cape Fold Belt, Karoo and Cretaceous Basins of Southern Africa. Balkema, Rotterdam. ISBN 90 5410 0478.

De Beer, C. H. (1990). Simultaneous folding in the Western and Southern Branches of the Cape Fold Belt. *S. Afr. J. Geol.* **93** (4), 583-591.

De Beer, C.H. (1995). Fault interference from simultaneous shortening in different directions: the Cape Fold Belt syntaxis. *J. Afr. Earth Sci.*, **21** (1), 157-169.

De Beer, C.H. (2002a). The stratigraphy, lithology and structure of the Table Mountain Group, In: K. Pieterse and R. Parsons (eds.), A synthesis of the hydrogeology of the Table Mountain Group - formation of a research strategy. Water Research Commission (WRC) Report, TT 158/01, pp. 9-18.

De Beer, C.H. (2002b). Structural investigation of the Koo Valley for the Koo Valley Water Scheme. Unpublished Council for Geoscience (CGS) Report No. 2002-0067, Bellville.

De Beer, C.H. (2004). Council for Geoscience. Bellville. Personal communications.

Deep Offshore Water Research (2006). A Project Proposal (July 2006) by the Offshore Water Investments. Contact person: Jack Holliday at; E-mail [jackh@mighty.co.za]/ [office@deepwater.co.za] or Cell: +27739131291. Research Proposal presented to the Groundwater Group of the University of the Western Cape.

Diamond R.E. and Harris C. (1997). Oxygen and hydrogen isotope composition Western Cape meteoric water. *South African Journal of Science*, Vol.93, pp 371 – 374.

Diamond R.E. and Harris C. (2000). Oxygen and Hydrogen isotope geochemistry of thermal springs of the Western Cape: recharge at high altitude? *Journal of African Earth Sciences*, Vol. 31, No. 3/4, pp. 467-481.

Domenico, P.A., and Schwartz, F.W. (1998). Physical and Chemical Hydrogeology, 2nd Edition: New York, John Wiley and Sons, Inc., 506 p.

Drever, J. I (1988). The geochemistry of natural waters. Second Edition. Prentice hall, New Jersey.

Drever, J.I. (1997). The geochemistry of natural waters: Surface and Groundwater Environments; Prentice-Hall, New Jersey, ISBN: 0 13 171790

Faure, G., 1998. Principles and Applications of Geochemistry, 2nd Ed., Prentice-Hall, Inc., New Jersey, 306 pp.

Fontes J. Ch. and J.M. Garnier (1979) Determination of the initial ¹⁴C activity of total dissolved carbon: A review of existing models and a new approach. *Water Resources Research*, 15: 399-413.

Frape S.K., Blyth A, Blomqvist R, McNutt R.H, and Gascoyne M (2004). Deep fluids in the continents: II. Crystalline rocks. In: Drver JI (eds) Treatise on Geochemistry 5 surface and groundwater, weathering, and soils. Elsevier, Amsterdam, pp 541-580.

Harris, C., Oom B.M., Diamond R.E., (1999). A preliminary investigation of the oxygen and hydrogen isotope hydrology of the greater Cape Town area and an assessment of the potential for using stable isotopes as tracers. *Water SA Vol. 25 No.1*, 15-24.

Harris, C. and Diamond R. E. (2002). The thermal springs of the Table Mountain Group: A stable isotope study. In: K. Pietersen and R. Parsons (eds.), A synthesis of the hydrogeology of the Table Mountain Group - formation of a research strategy. Water Research Commission (WRC) Report, TT 158/01, pp.230-235.

Harris, C. and Talma, A.S., (2002). The Application of Isotopes and Hydrochemistry as Exploration and management Tools, In: K. Pietersen and R. Parsons (eds.), A synthesis of the hydrogeology of the Table Mountain Group - formation of a research strategy. Water Research Commission (WRC) Report, TT 158/01, pp. 62-70.

Harter, T. (2004). Personal communications (e-mail). UCCE Groundwater Hydrology Program, University of California, Davis.

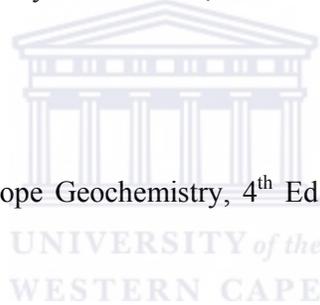
Hartnady, C.J.H., (1998). Record groundwater strike near Citrusdal. *Borehole Water Journal*, **39**, 1-4.

Hartnady, C.J.H. and Hay, E.R., (2002a). Boschkloof Groundwater Discovery, In: K. Pietersen and R. Parsons (eds.), A synthesis of the hydrogeology of the Table Mountain Group - formation of a research strategy. Water Research Commission (WRC) Report, TT 158/01.

Hartnady, C.J.H., and Hay, E.R., (2002b). Experimental Deep Drilling at Blikhuis, Olifants River Valley, Western Cape: Motivation, Setting and Current Progress, In: K. Pietersen and R. Parsons (eds.), A synthesis of the hydrogeology of the Table Mountain Group - formation of a research strategy. Water Research Commission (WRC) Report, TT 158/01.

Hinkle, S.R., and Snyder, D.T. (1993). Comparison of chlorofluorocarbon dating and particle-tracking results of a regional ground-water flow model-Portland Basin, Oregon and Washington [abs.]: Abstracts with Programs-Geological Society of America, in Geological Society of America, 1993 annual meeting, v. 25, no. 6, October 1993, p. 366.

Hoefs, J. (1996). Stable Isotope Geochemistry, 4th Ed. Springer-Verlag, Berlin, pp. 241.



Hoffmann, J.R.H., (1979). Die chemise samestelling van warmwaterbronne in Suidwes Afrika. CSIR, Pretoria, Special Report WAT 56A.

Hsieh, P.A. (2001). TopoDrive and ParticleFlow-Two computer models for simulation and visualization of ground-water flow and transport of fluid particles in two dimensions. U.S. Geological Survey Open-File Report 01-286, Menlo Park, California

Huyakorn, P.S., and Pinder, G.F., (1983). Computational Methods in Subsurface Flow: New York, Academic Press, 473 p.

International Atomic Energy Agency (IAEA), (1981). Statistical treatment of environmental isotope data in precipitation. Tech. Report 206, Vienna.

Imbach, T. (1997). Deep groundwater circulation in the tectonically active area of Bursa, Northwest Anatolia, Turkey. *Geothermic, Vol. 26, pp. 251-278, Elsevier.*

James E.R., Manga M., Rose T.P. and Hudson G.B, (2000). The use of temperature and the isotopes of O, H, C, and noble gases to determine the pattern and spatial extent of groundwater flow. *Journal of Hydrology 237, 100-112.*

Jones M.Q.W. (1992). Heat flow in South Africa. *Geol. Surv. S. Afr. Handbook, 14, 174pp.*

Kalin R. (1999). ^{14}C in groundwater. In: P.G. Cook and A.L. Herczeg (eds) Environmental tracers in subsurface hydrology. Kluwer, Netherlands.

Kent L.C., (1949). Thermal waters of the Union of South Africa and South West Africa *Trans. Geol. Soc. S. Afr., 52, 231-264.*

Kotze, J.C. (2002) Towards a Management Tool for Groundwater Exploitation in the Table Mountain Sandstone Fractured Aquifer. WRC Report No. 729/1/02

Lin, L. (2003). Hydrostratigraphic subareas of the TMG Sandstone. A poster prepared for the WRC project No. K5-1419, University of the Western Cape, Bellville.

Lloyd, J. W and J. A Heathcote (1985). Natural inorganic hydrochemistry in relation to groundwater, an introduction. Clarendon Press, Oxford.

Maloszewski P. and Zuber A. (1991) Influence matrix diffusion and exchange reactions on radiocarbon ages in fissured carbonate aquifers. *Water Resources Research*, 27: 1937-1945.

Manning C.E, Ingebritsen S.E (1999). Permeability of the continental crust: implications of geothermal data and metamorphic systems. *Rev. Geophysics* **37**: 127-150.

Mazor, E. (1991). Applied chemical and isotopic groundwater hydrology. Open University Press, Buckingham.

Mazor, E. and Verhagen, B.T. (1983). Dissolved ions, stable and radioactive isotopes and noble gases in thermal waters of South Africa. *Journal of Hydrogeology* **63**, 315-319.

McLennan J. and Williams S., (2005). Deep water Africa reaches turning point. *Oil and Gas Journal* **103** (6), 18-25.

Meyer P.S. (1999). An Explanation of the 1:500 000 General Hydrogeological Map, Oudtshoorn 3320. Department of Water Affairs and Forestry. Pretoria.

Meyer P.S. (2001). An Explanation of the 1:500 000 General Hydrogeological Map, Cape Town 3317. Department of Water Affairs and Forestry. Pretoria.

Meyer, P.S. (2002). Springs in the Table Mountain Group, With Special Reference To the Fault Controlled Springs, In: A synthesis of the Hydrogeology of the Table Mountain Group-Formation of a Research Strategy, Pietersen K. and Parsons R. (Eds.). WRC Report No. TT 158/01, Pretoria.

Ninham Shand (2004). TMG Aquifer Feasibility Study and Pilot Project: Draft Scoping Report. Report Number 400396/3715

Neretnieks, I. (1993): Solute transport in fractured rock- Applications to radionuclide waste repositories, Inflow and contaminant transport in fractured rock, Chapter 2, pp.39-128, Academic Press Inc.

Okamoto, R., Ogata, M. and Kojima, K. (1988): A new framework for civil engineering Japan Society of Civil Engineers, Vol.14, pp.65-71

Pietersen K. and Parsons R. (eds.) (2002). A synthesis of the Hydrogeology of the Table Mountain Group-Formation of a Research Strategy, WRC Report No. TT158/01, Pretoria, South Africa.

Rosewarne P. (2001). Hydrogeological Characteristics of the Table Mountain Group Aquifers, In: A synthesis of the Hydrogeology of the Table Mountain Group-Formation of a Research Strategy, Pietersen K. and Parsons R. (Eds.). WRC Report No. TT 158/01, Pretoria.

----- and J.M.C Weaver, (2002). Identification of Targets for Drilling in the TMG Aquifer. In: A synthesis of the Hydrogeology of the Table Mountain Group-Formation of a Research Strategy, Pietersen K. and Parsons R. (Eds.). WRC Report No. TT 158/01, Pretoria. Pp 85-88.

Saeze, H.A. (2005). Delineation of deep groundwater regimes in the Table Mountain Group Syntaxis, unpublished PhD progress report (July 2005), University of the Western Cape.

Scholl, M.A., Ingebritsen, S.E., Janik, C.J., Kauahikaua, J.P. (1996). Use of precipitation and groundwater isotopes to interpret regional hydrology on a tropical volcanic island: Kilauea Volcano area, Hawaii. *Water Resour. Res.* 32, 3525–3537.

Seiler K. and Lindner W. (1995). Near-surface and deep groundwaters. *Journal of Hydrology*, 165, 33-44.

Snow, D.T. (1968): Rock fracture spacings, openings and porosities, In: Proc. of ASCE 96, pp.73-91.

Stober I and Bucher K (2005) Hydrogeologic properties of a deep fractured basement aquifer. *Geofluids*, 5, pp.8–19.

Stumm, W. and Morgan, J.J.(1996): Aquatic Chemistry, Chemical Equilibria and Rates in Natural Waters, 3rd ed. John Wiley & Sons, Inc., New York, 1022p

Talma A.S., Vogel, J.C. and Heaton, T.H.E. (1983). The geochemistry of the Uitenhage artesian aquifer: Carbonate solution in a closed system. In: Isotopes in Hydrology, Int. Symp, Vienna: IAEA.

Tamers, M.A. (1975). The validity of radiocarbon dates on groundwater. *Geophysical Survey*, 2:217-239.

Tanaka, K. and Inoue, D. (1995): Fault -definition and usage-, Tsuchi-to-kiso, The Japanese Society of Soil Mechanics and Foundation Engineering, Vol.43-3, pp.11-14.

TMG Deep Aquifer Feasibility and Pilot Project. Available at <http://www.tmg-aquifer.co.za/>

Verhagen, B.Th. and M.J. Butler (2004). Isotope hydrology-An African Experience. Geoscience Africa 2004, Extended Abstracts, Vol. 2. Geological Society of South Africa, Johannesburg. pp 671-672.

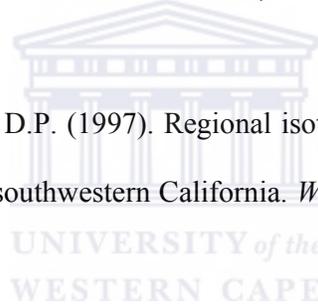
Vogel, J.C. 1967. Investigation of groundwater flow with radiocarbon. Isotopes in Hydrology, IAEA, Vienna

Vogel J.C. (1970) Carbon-14 dating of groundwater. In: Isotope hydrology, IAEA Symposium 129, pp.225-239.

Weaver, JMC and Talma AS, (1999). Field Studies of Chlorofluorocarbons (CFC'S) as a Groundwater Dating Tool in Fractured Rock Aquifers. WRC Report No. 731/1/99.

Weaver, J.M.C, Talma A.S and Cavé L.C (1999). Geochemistry and Isotopes for Resource Evaluation in the Fractured Rock Aquifers of the Table Mountain Group. Report No. 481/1/99, Water Research Commission, Pretoria.

Williams, A. E. and Radoni, D.P. (1997). Regional isotope effects and application to hydrologic investigations in southwestern California. *Water Resour. Res. Vol. 33, No. 7, p. 1721 (97WR01035)*.



Wu, Y. (2005). Groundwater recharge estimation in TMG aquifers systems with a case study of Kammanassie area. Unpublished PhD thesis. University of the Western Cape. Bellville.

Wu, Y. and Xu, Y (2005). Snow impact on groundwater recharge in Table Mountain Group aquifer systems with a case study of the Kommissiekraal River catchment South Africa. *Water SA Vol. 31 No.3. ISSN 0378-4738*.

Yurtsever, Y., 1999. Use of environmental tritium to study catchment dynamics: case study from the Danube River basin. In: Leibundgut, C., McDonnell, J., Schultz, G. (Eds.), *Integrated Methods in Catchment Hydrology – Tracer, Remote Sensing and New Hydrometric Techniques*. IAHS Publication No. 258: 167-174.

