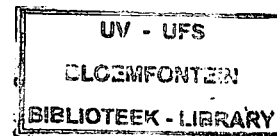


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Surface water - Groundwater interactions: Development of methodologies suitable for South African conditions

By

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Submitted in fulfilment of the requirements for the degree of Doctor of Philosophy in the Faculty of Natural Sciences and Agriculture, Institute for Groundwater Studies, University of the Free State, Bloemfontein, South Africa.

Declaration

I declare that the dissertation hereby handed in for qualification PhD Geohydrology at the University of the Free State, is my own independent work and that I have not previously submitted the same work for a qualification in another university or faculty.



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Acknowledgements

Firstly and most importantly I thank the Lord who made it all possible and blessed me with wonderful and supportive people.

I would like to express my gratitude and appreciation to my supervisor Dr Danie Vermeulen for his selfless and unreserved support and guidance during the research investigation. I also wish to specially thank my co-supervisor, Professor Ingrid Dennis who believed in me even when I was down, dragging my feet and almost out. Both of them played a very crucial role in ensuring that I complete this piece of work. A special word of appreciation also goes to Professor Kai Witthueser who selflessly allowed me to use the related work that he previously did together with Professor Dennis, as reference. I'm also quite appreciative of the entire Institute for Groundwater Studies staff for their unwavering support attitude. Of particular note is Professor Gerrit van Tonder for guidance on conceptual issues and data as well as Mr Eelco Lukas for assistance on programmes and technical hiccups. Both Dr Shafick Adams and Dr Modreck Gomo are acknowledged and thanked for the training they gave on how to draw stable isotope diagrams. A number of individuals who generously provided data such as Professor Simon Lorentz, Mr Fanie de Lange, Dr Rian Titus and staff of Water Geosciences Consulting are also acknowledged and thanked.

I also wish to thank my wife Monkie, who encouraged and supported me at all times.

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Abbreviations

Ca	:	Calcium
DAFLOW	:	Diffusion Analogy Surface-Water Flow model
EC	:	Electrical conductivity
EWR	:	Environmental Water Requirement
FC method	:	Flow Characteristics method
FEFLOW	:	Finite Element subsurface FLOW system
GSFLOW	:	Groundwater and Surface water FLOW
GMWL	:	Global Meteoric Water Line
HCO ₃	:	hydrogen carbonate
HRU	:	Hydrologic Response Unit
MAE	:	Mean annual potential evaporation
MAP	:	Mean annual precipitation
MODFLOW	:	Modular Groundwater Flow Model
Na	:	Sodium
PRMS	:	Precipitation-Runoff Modeling System
<i>sdf</i>	:	Stream depletion factor
SFR1	:	Streamflow Routing Package Number 1
SHETRAN	:	Système Hydrologique Européen TRANsport
SiO ₂	:	Silicon dioxide
TDS	:	Total Dissolved Solids
TSM	:	Transient Storage Models
USGS	:	United States Geological Survey
VSMOW	:	Vienna Standard Mean Ocean Water
WMA	:	Water Management Area

Units of Measurement

a	:	annum
cm	:	centimetre
°C	:	degree celcius
d	:	day
h	:	hour
km ²	:	square kilometre
l	:	litre
l/s	:	litres per second
m	:	metre
m ²	:	square metre
m ² /d	:	square metres per day
m ³	:	cubic metre
mamsl	:	metres above mean sea level
mbgl	:	metres below ground level
mm	:	millimetre
Ml/d	:	Mega litres per day
Mm ³	:	Million cubic metres
s	:	second
Ω	:	Ohm

1 Scope and Objectives

1.1 Introductory background and motivation

The South African National Water Act (Act No 36 of 1998) recognizes that water is a scarce and unevenly distributed national resource which occurs in many different forms which are part of a unitary, interdependent cycle and that as such should be managed in an integrated manner. However, surface water and groundwater systems have historically been managed and dealt with as though they were separate entities despite that they are both interlinked components of the water cycle. Ruehl *et al.*, (2006) and Weidong *et al.*, (2007) also contend that surface water and groundwater are undivided components of the hydrologic system, and that development or contamination of one component commonly affects the water quality of the other. Groundwater and surface water are not isolated components of the hydrologic system (Sophocleous, 2002), but instead interact in a variety of physiographic and climatic landscapes. According to Winter (1999) understanding the basic principles of the interaction of surface water and groundwater is needed for effective management of water resources. Conjunctive management and use of surface water and groundwater is crucial for sustainability of this precious and yet scarce commodity, water, especially in South Africa that is ranked the 30th driest country in the World. The water security related challenge that water resource managers and planners in the country often encounter besides economic and environmental concerns that are associated with building of dams, is that identification of suitable sites for construction of surface water storage facilities is seldom feasible, thus increasing dependency on groundwater to augment limited surface water resources.

Understanding surface water - groundwater interaction may also enable one to find solutions to some of the related problems such as potential water quantity or quality impact of one component by the other, quantification or estimation of environmental flows or even water allocation that ensures prevention of double counting. However, management decisions concerning the allocation of water resources in environments where surface water and groundwater are linked is quite challenging. McCallum *et al.*, (2009) contends that part of the difficulty lies in the lack of data and lack of knowledge about the processes

controlling the exchange of water between aquifers and rivers. Given this paucity of data (particularly but not exclusively in South Africa) and gap in knowledge, sustainable management of the quality and quantity of water resources particularly with regard to allocation thereof without the risk of double counting is unlikely. Currently, there are no methodologies that are formally accepted for use in South Africa besides tacit application of various methods, markedly but not exclusively hydrograph separation approaches, depending on availability of data, models and preferred approach by practitioners.

This research entails literature study on previous work on surface water - groundwater interaction, various techniques for quantification and assessment of the interaction based on case studies, currently used methodologies, as well as proposed approaches that are most likely to be suitable for South African conditions.

1.2 Hypothesis and objectives

1.2.1 Hypothesis

Although often evaluated as separate entities, groundwater and surface water interact to a varied extent or degree from no interaction to a highly interactive regime, and at various spatial and temporal scales. The nature of the interaction is complex, partly due to heterogeneous hydrological setting of the water regime, hydraulic properties of host rock strata, climatic factors, geology and structure. Understanding surface water - groundwater processes and dynamics of flow inform better water management especially regarding resource protection, assessment and allocation.

1.2.2 Objectives

This research study was aimed at developing appropriate methodologies for assessment and evaluation of the surface water - groundwater interaction and thus to quantitatively estimate groundwater contribution to surface water flows and the vice versa. However, during the course of the investigation it became clear that, in light of unforeseen challenges related to unavailability of adequate data and resources, development of methodologies would not be achievable. Hence the alternative approach taken was to investigate, identify and recommend methodologies that are suitable for South African conditions and to develop the relevant framework to guide the process of choosing suitable methodologies.

This research study consists of six chapters. Chapter 1 deals with scope, aims and objectives of the research investigation. Chapter 2 focuses on the literature review of various national and international techniques and methods that are aimed at evaluation or quantification of surface water - groundwater interaction. A historical account of the development and application of analytical and numerical models for characterisation of surface water - groundwater interaction is also covered in Chapter 2. Case studies used as part of the investigation and identification of the methodologies that are appropriate for South African conditions and associated lessons learnt are discussed in Chapter 3. Chapter 4 deals with the critical analysis of case studies and recommended changes in certain aspects that need to be re-looked at. Classification systems and frameworks for assessment of surface water - groundwater interaction are covered in Chapter 5. The final Chapter 6 entails knowledge contribution and conclusions.

In conclusion of this chapter, it should suffice to indicate that the outdated historical practice of treating groundwater and surface water as separate entities despite their inter-relation and linkages needs to be discarded. Additionally, the unintended consequences of South Africa's previous Water Act, 1956 (Act No. 54 of 1956) that had a riparian approach as a guiding principle, has since been adequately addressed under the current National Water Act, 1998 (Act No. 36 of 1998). The riparian principle perpetuated a system that kept use and to a large degree management of groundwater and surface water as separate entities, though exceptional cases, such as situations where Subterranean Government Water Control Areas had to be declared, included conjunctive use. Whereas current Act operates under the principle of integrated water resource management; thus ensuring equitable treatment of various phases of the water cycle. This research study's focus on seeking the appropriate methodologies for assessment of the interaction between the two components of the water cycle contributes to the challenge of effective conjunctive water use.

2 Literature Review

2.1 Introduction

Understanding the distribution and the dynamics of the interaction between surface water and groundwater is necessary and essential for assessment or quantification of the contribution of one component to another. Another essential element is conceptual knowledge of the structural and system controls that govern the occurrence and movement of water from the groundwater to the surface water component and vice versa. It is accordingly essential to begin this research by providing a general overview of system components (i.e. various aspects of the spatial distribution of surface water and subsurface water including groundwater) and the dynamics of flow across the interface between the two components. The overview of the system components is followed by a focus on the nature of linkages between these systems and a discussion on ways of identifying the interaction, and finally a literature review of methodologies for surface water - groundwater interaction.

2.1.1 Various components of surface water and groundwater

The surface water component comprises of water in the rivers, lakes, dams and overland flow, while the unsaturated zone component constitutes that part of the subsurface where the infiltrating water from rainfall or leakage from runoff does not completely fill the voids in between the soils and rocks. Although flow in the unsaturated zone is generally downwards in response to gravity, relatively impermeable rock layers often impede infiltration to layers below causing horizontal flow that could discharge as seepage to the surface or streams. Such flow is called interflow. The groundwater component comprises the saturated zone that is replenished or recharged by the infiltrating water from rainfall and overlying layers. Seepage from groundwater storage, particularly during extended drought periods sustains streams and such a contribution to surface water is called baseflow. Streams that are often observed flowing even long after it had rained are invariably fed by springs and groundwater leakages. Figure 1 conceptually depicts various components of both the surface and sub-surface water. However, in real hydrological

settings the system is more complex due to heterogeneity of the host rocks, where various factors such as climate, geohydrology, ecology and human induced impacts modify the process of interaction between surface water and groundwater.

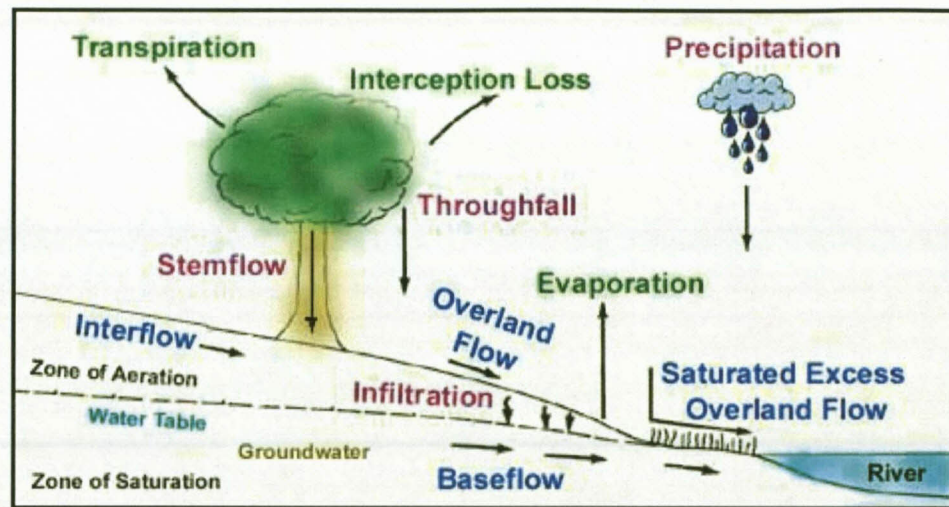


Figure 1: Various components of surface water and groundwater (Berner & Berner, 1987)

2.1.2 Surface water and groundwater are inter-linked components of the same system

Surface water and groundwater have long been considered separate entities, and have been investigated individually (Kalbus *et al.*, 2006). However, hydrologists have always recognized that groundwater and surface water are closely linked, yet studies have mostly been carried out largely by single disciplines. A number of authors (Winter, 1999; Sophocleous, 2002; and Weidong *et al.*, 2007) maintain that surface water and groundwater are undivided components of the hydrologic system, since development or contamination of one component commonly affects the response or water quality of the other. Historically, South Africa's Department of Water Affairs was structurally configured in a way that reflects differentiated units where the Hydrology division separately dealt with surface water issues, with minimal interaction with the Geohydrology directorate that looked after the groundwater component. However, those divisions have since been integrated through the restructuring process.

2.1.3 Identification or evidence of surface water - groundwater interaction

To identify areas of interaction between surface water and groundwater, various methods of assessing or measuring flux across the streambeds are necessary. These may include use

of air photos or map interpretation to locate geological features, electrical conductivity and temperature surveys, isotopic signatures and chemical analyses as well as measuring groundwater discharges and stream fluxes or seepage measurements.

It seems logical to define the area where the actual interaction between surface water and groundwater occurs, in order to understand the nature, mechanism and processes that govern that interaction. This interface, commonly known as the hyporheic zone was defined differently by various researchers. Figure 2 shows an example thereof. The Environment Agency (2005) states that the term hyporheic is derived from the Greek language – *hypo*, means under or beneath, while *rheos*, refers to a stream (*rheo* means to flow). In other words the hyporheic zone occurs below a stream. According to Janzen (2008) the term hyporheic zone is used to represent both the interface and the process where surface water and groundwater are exchanged through the channel bank and bed in the near stream environment, while Brown *et al.*, (2007) defines the hyporheic zone as a surface water - groundwater interface, that is important to solute transport and transformation in stream-aquifer systems. In this research investigation, the hyporheic zone is regarded as an interface between surface water and groundwater where the processes of interaction occur.

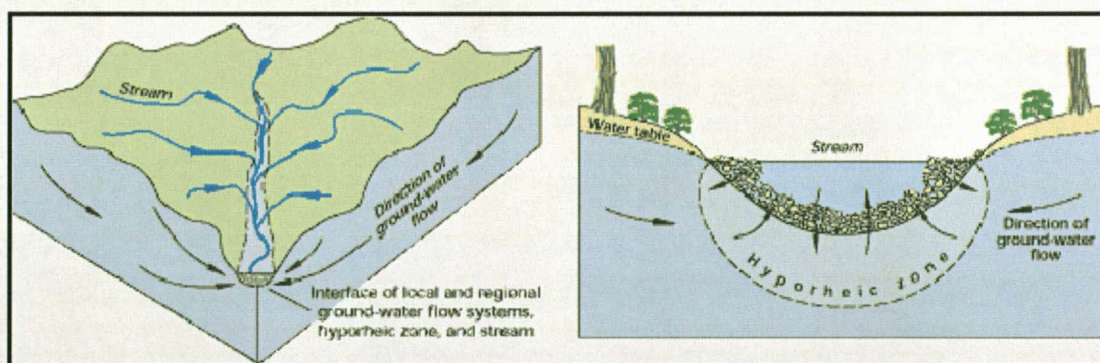


Figure 2: Conceptual hyporheic zone for a typical stream (Winter *et al*, 1998)

2.2 Detailed literature Review

Different methods of assessing the interaction between surface water and groundwater have over the years been developed by various investigators. The following is a brief

account of a number of approaches used ranging from Darcian flux based methods through chemistry approaches, isotopic hydrology, hydrograph separation methods, to analytical and numerical methods.

A number of authors (e.g. Sophocleous, 2002; Malcolm *et al.*, 2005; Kalbus *et al.*, 2006; and McCallum *et al.*, 2009) discuss various methods of investigating stream–aquifer interactions. Typical approaches often entail statistical analyses of hydrological data (i.e. rainfall, stream flow and hydrograph) in order to establish connectivity (i.e. whether the river is gaining water from or losing water to the aquifer); application of Darcy's Law, which states that water flux is a function of hydraulic gradient and conductivity; slug and pumping tests to determine hydraulic properties, and field measurements using seepage meters. However, conventional Darcian flux based methods may not be adequate to assess spatial and temporal dynamics of chemical loading to a river or an aquifer (Keery *et al.*, 2007). Hence tracers, chemical, and isotopic hydrology based methods are also considered in this research study, to; among other approaches confirm the interaction that would have volumetrically been evaluated.

Interaction between surface water and groundwater is likely to occur if these two components of the hydrological cycle are hydraulically connected. Winter *et al.*, (1998) recognizes that the interaction takes place in three basic ways: (1) streams gain water (Figure 3A) from inflow of groundwater through the streambed (gaining stream) such as the upper reaches of rivers on the eastern escarpment like Vaal, Olifants, Tugela, Blyde and Komati (Dennis and Witthueser, 2007); (2) streams may lose water as illustrated in Figure 3B (e.g. lower sections of Kuruman River and Molopo River) to groundwater by outflow through the streambed (losing stream), or (3) both gain in some reaches and lose in others (intermittent stream). On the other hand, Kalbus *et al.*, (2006) argues that the interactions between surface water and groundwater basically proceed in two ways; namely that groundwater flows through the streambed into the stream (gaining stream), or stream water infiltrates through the sediments into the groundwater (losing stream). The larger-scale (e.g. catchment) hydrologic exchange of surface water and groundwater in a landscape is controlled by (1) the distribution and magnitude of hydraulic conductivities, both within the channel and in the associated alluvial-plain sediments; (2) the relation of stream stage

to the adjacent groundwater level; and (3) the geometry and position of the stream channel within (Sophocleous, 2002) the alluvial plain. The direction of the exchange flow depends on the relative hydraulic heads. In gaining streams, the groundwater level would typically be higher than that of the stream. Conversely, in losing streams the groundwater level is lower than that of the stream. The fluxes occur over a range of time and length scales, and are influenced by channel geomorphology, lithologic variability, and hydrogeologic properties of the streambed and near-stream formations (Ruehl *et al.*, 2006).

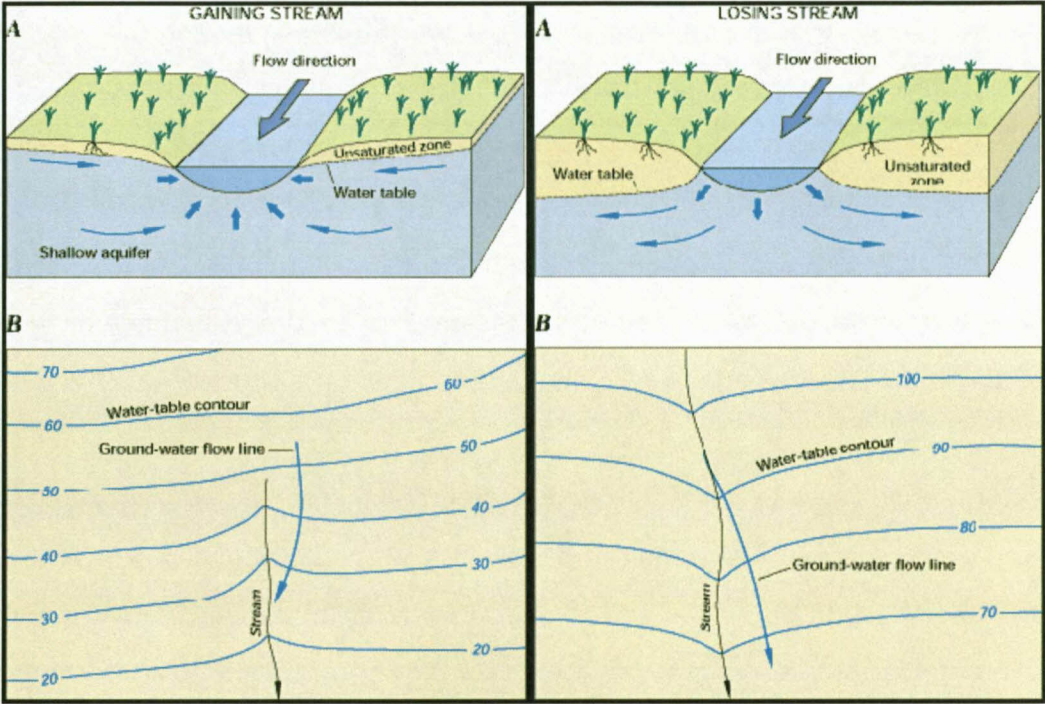


Figure 3A: Gaining stream (Winter *et al.*, 1998) **Figure 3B:** Losing stream (Winter *et al.*, 1998)

To understand the interaction between surface water and groundwater, one should first analyse the regional groundwater flow in relation to topographical characteristics and surface water bodies in order to determine the type of interaction that is likely to occur in the study area. Water-table contour maps provide information on the groundwater levels and the direction of groundwater flow; (see part B, or plan view of both Figures 3A and 3B). It can easily be deduced from the water table contour lines (Winter *et al.*, 1998) whether a stream is gaining (contour lines point in the upstream direction) or losing (contour lines point in the downstream direction).

The direction of local groundwater flow can be determined from the differences in hydraulic head between individual piezometers installed in groups (at least three in a triangular arrangement). In the case of horizontal flow, the hydraulic gradient can be calculated from the difference in hydraulic head and the horizontal distance (Kalbus *et al.*, 2006). For the vertical components of groundwater flow, which are particularly important to understand the interaction between surface water and groundwater, a piezometer nest may be installed, with two or more piezometers set in the same location at different depths. The hydraulic gradient can then be calculated from the difference in hydraulic head and the vertical distance. Furthermore, vertically distributed piezometer data can be used to draw lines of equal hydraulic head for the construction of a flow field map showing the groundwater flow behaviour in the vicinity of a surface water body. The former approach has for instance been used in Seekoei River (a non-perennial tributary of the Orange River in South Africa) while the latter approach was utilised in the experimental Weatherley Catchment in the Eastern Cape to determine the hydraulic gradient of groundwater flow.

Characterization of surface water - groundwater interaction can also be illustrated based on measurements of the concentration of dissolved oxygen. Malcolm *et al.*, (2005) in their research investigation on spatial and temporal variation of surface water - groundwater interaction undertook a spatial survey of stream flows, took samples from the hyporheic zone and analysed them for dissolved oxygen and electrical conductivity. The results indicated that the long residence groundwater is often typically characterized by low dissolved oxygen, while sites dominated by surface water had a higher concentration of dissolved oxygen content or near saturation level. The groundwater of low dissolved oxygen was also found to be of low water quality and hence to be detrimental to salmon survival. It is also clear from the study that hydro-chemical tracers are useful tools for assessment of the interaction between surface water and groundwater.

Hydraulic conductivity is a function of the density and viscosity of water which are temperature dependent. Hence, thermal effects cause changes in hydraulic conductivity and this relationship forms the basis for utilisation of a temperature series to assess the variability of hydraulic exchange between surface water and groundwater. Differences in streambed temperatures are used to identify areas of groundwater discharge or surface

water infiltration. The hydraulic conductivity or the capacity of a porous medium to transmit water is defined in terms of intrinsic permeability or ease with which fluid flows through a rock formation:

$$K = k \frac{\rho g}{\mu} \quad (2.1)$$

Where

- k = the intrinsic permeability,
- K = the hydraulic conductivity,
- μ = the dynamic viscosity,
- ρ = the fluid density
- g = the acceleration of gravity.

Conant Jr. (2004), in a research study on delineation and quantification of groundwater discharge using streambed temperature measurements, found that in winter higher groundwater discharge locations were associated with relatively cooler areas of the streambed. He also discovered that Darcian methods for calculation of vertical groundwater flux confirmed discharge inferred from temperature measurements. Temperature measurements can be analyzed for recharge and discharge rates, to detect infiltration of surface water into fractures and to solve the inverse problem by using temperature to estimate groundwater velocity and hydraulic conductivity (Anderson, 2005). Temperature measurements have also proved useful in estimating groundwater flux in wetland settings in lakes and in coastal aquifers including estimation of submarine groundwater discharge. For instance, O’Driscoll and DeWalle (2006) analyzed the differences in energy exchange processes occurring between a stream dominated by groundwater inputs and one with minimal groundwater inputs and illustrated how the stream versus air temperature relations are affected by groundwater inputs at 12 stations within the Spring Creek drainage basin in Pennsylvania in a karst setting. Stream temperature – air temperature relations were compared for the 12 stream locations using linear regressions between weekly average stream temperatures versus weekly average air temperatures. Hourly air and stream temperature were summarized as weekly moving average data for regression analyses. As a management tool, stream–air temperature

relations can reveal the importance of groundwater inputs, particularly in small watersheds where the travel time of surface waters is short enough to allow surface waters to carry thermal evidence of groundwater inputs. Records of stream–air temperature relations over time can be indicative of changing hydrological regimes. On the other hand, Keery *et al.*, (2007) developed an analytical method; a method of utilising temperature time series to calculate vertical water fluxes across riverbed sediments.

Tracer tests are also powerful tools for assessment of surface water - groundwater interaction, although they were not utilised in this thesis. A series of tracer tests (i.e. injection of the tracers followed by periodic measurements of the concentrations, laboratory analyses and studying the breakthrough curve) to determine whether hydrologic exchange occurs within a strongly losing stream, were performed by Ruehl *et al.*, (2006). The study demonstrated that a thorough hydrologic balance obtained by repeated direct measurements of stream discharge, in combination with a transient storage models (TSM) approach can constrain seepage fluxes better than either method on its own. The results also showed that the TSM approach can be used to obtain meaningful stretch specific hydrologic parameters within a strongly losing stream.

Jones *et al.*, (2006) noted that separating event (precipitation) from pre-event (unsaturated & saturated sub-surface flow that existed prior to rainfall event) water using tracer-based separation techniques has been undertaken by a number of researchers. In that project the authors (Jones *et al.*, 2006) aimed to quantify the influence of dispersive/diffusive mixing of tracer signatures for rainfall, unsaturated and saturated waters, and to demonstrate how changing signatures along flow paths can influence tracer based estimates of pre-event subsurface contributions to streamflow using a fully coupled integrated hydrology model. In the first instance the so called “rapid mobilization of old water paradox” (i.e. rapid water level response as observed from boreholes immediately following rainfall despite slow movement of groundwater flow), which seems to suggest that pre-event water contributes more to streamflow during a rainfall event was then attributed to the capillary fringe effect. That is, since the capillary fringe extends from water table to land surface (especially near stream environments) particularly in shallow aquifers, addition of small amount of water relieves capillary tension and produces a rapid rise – increasing the hydraulic driving force

for rapid discharge. The Borden rainfall-runoff experiment was undertaken to assess the relative contribution of pre-event water to streamflow and a bromide tracer was added to artificial rainfall water, to ensure differentiation of event from pre-event water using a hydrograph separation approach (Jones *et al*). Three different sources were tagged that contribute to streamflow simulated using three (3) different tracers where one tracer concentration was assigned to rainfall to tag movement and transport of irrigation water, the second to tag water initially stored in the unsaturated zone while the third tagged water initially stored in the saturated zone. That enabled determination of relative contributions to the streamflow from rainfall (event), saturated zone (pre-event) and unsaturated zone (pre-event) waters as they migrate through the system. The results showed that the pre-event water contributions to the total discharge, estimated using tracer based separation methods, were found to be larger than hydraulically based values since an increase in the value of subsurface longitudinal dispersivity caused the estimate of the tracer based pre-event contribution to increase noticeably. When the longitudinal dispersivity is set at zero, thereby eliminating mechanical mixing between pre-event and event waters, the tracer based estimate was still significantly larger than that of hydraulically based estimate due to molecular diffusion on mixing. Regarding the influence of rainfall intensity or duration (shown by altering the rainfall event) it was found that increasing intensity while decreasing duration, led to a lower pre-event contribution (due to less time being available for hydrodynamic mixing by dispersion & diffusion).

A study on use of environmental isotope tracers to evaluate the contribution of surface water to a groundwater reservoir during the rainy season was undertaken by Bae *et al.*, (2000) in Korea. The investigation involved the collection of water samples from a stream, and from shallow and deep boreholes, daily, under conditions of continuous pumping of the deep borehole over a period of five months, followed by analysis of deuterium and tritium isotopes. All water samples from the stream and shallow aquifer were of the same chemical composition (i.e. Ca – HCO₃ type) while the water from the deep aquifer was comparatively more enriched in Ca, Na, SiO₂ and HCO₃ due to the long residence time and interaction with host rocks. The isotope signature was such that all water plotted closer to the World-wide meteoric line on the $\delta^{18}\text{O}$ - δD diagram. The results thus indicated that isotopic composition of stream and shallow boreholes was influenced by the amount of

rainfall since that varied with different rain events whereas deep groundwater maintained stable isotopic composition with time. The investigation also revealed that groundwater circulation was limited to shallow depths and that most of the recharged shallow boreholes discharged into streams through interflow and baseflow. Isotope analyses are relatively cost effective compared to conventional hydrological studies and useful in providing information on the origin, turnover and transit time of water in the system.

Other methodologies for assessment of surface water - groundwater interaction include the hydrograph separation approach. The hydrograph is the time series record that shows variation in discharge of a river over a time period during and after a rainfall event. It mainly and generally thought of as comprising two components; namely the quick flow which is the direct response to a rainfall event that can be resolved into overland flow (runoff), lateral movement in the soil profile or unsaturated zone (interflow) and direct rainfall onto the stream surface (direct precipitation); and baseflow which is the longer-term discharge derived from groundwater storage or from a shallow unconfined aquifer. Analysis of the hydrograph to separate out the baseflow component provides information on the characteristics of the natural storages feeding the stream.

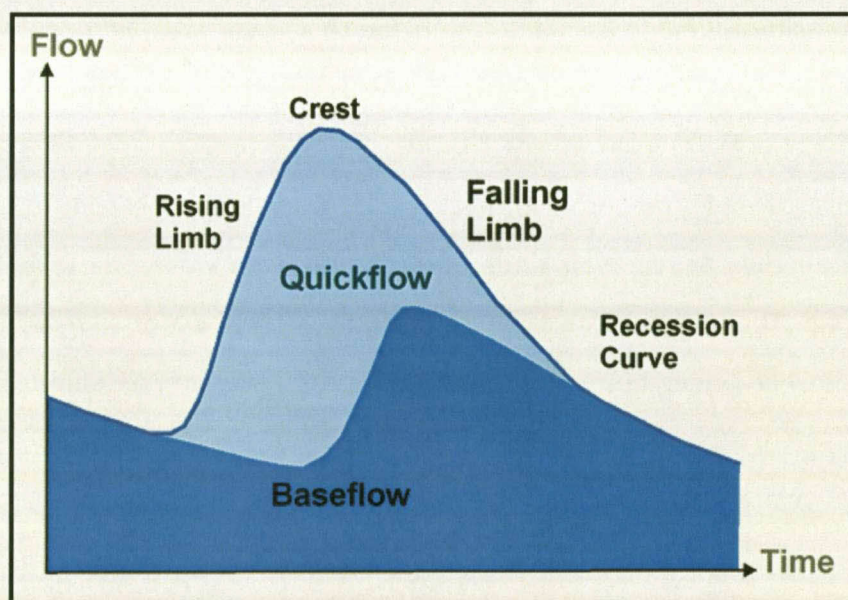


Figure 4: Components of a hydrograph (Australian Government, 2006)

Groundwater discharge from a shallow unconfined aquifer is commonly assumed to be the main contributor to baseflow. For this to be a significant process, the unconfined aquifer needs to be adequately recharged (typically on a seasonal basis), have a shallow water-table that is higher than the stream water level, and have adequate water storage and transmission properties to maintain flow to the stream (Smakhtin, 2001). For a gaining stream, where the underlying aquifer satisfies this criteria and groundwater contributes to stream flow, analysis of the stream hydrograph can indicate the magnitude and timing of this contribution.

During the dry weather periods, water is gradually removed from the catchment by evapotranspiration and groundwater/soil water discharge into a stream. A depletion of streamflow discharge during these periods is known as 'recession', and is reflected in streamflow hydrograph by a *recession curve* (Smakhtin, 2001). Different hydrological components including overland flow, interflow and groundwater flow have their own characteristic recession rates, but the ranges of these rates may overlap which reflect the fact that the distinctions between surface flow and interflow and between interflow and baseflow are not always clearly distinct. Hence a plot of logarithmic flow versus time enables categorisation into overland, runoff, interflow and baseflow where the gradient of the graph is the recession constant (Nathan and McMahon, 1990). Recession analyses undertaken by Nathan and McMahon (1990) based on 186 catchments in southern Australia indicated that the recession constants for surface water runoff, interflow and baseflow range as follows 0.2 to 0.8; 0.7 to 0.94; and 0.93 to 0.995 respectively. While on the other hand recession constants determined for the Ntuzi catchments (Kelbe and Germishuys, 2010) in South Africa, range from 0.51 to 0.66 for quick-flow or runoff; from 0.76 to 0.82 for through-flow or interflow and from 0.98 to 0.99 for baseflow. In other words the recession constants are area specific perhaps due to differences in geohydrological conditions.

Initially, the recession curve is typically steep as quickflow (made up of overland flow and/or interflow in topsoil) leaves the basin. The curve then flattens out with delayed flow supply from deeper subsurface stores and may eventually become nearly constant if sustained by outflow from a glacier or from groundwater storage (Smakhtin, 2001).

Several complex functions define recession flow but no single linear plot can be constructed for baseflow recession. Non-linearity is often attributed to carry-over storage, variation in aerial pattern of recharge, evapotranspiration, bank storage, etc. There are several methodologies or versions of baseflow separation approaches. Some assume that the hydrological response is concurrent with surface runoff, while others rely on effects of bank-storage or assume that recession continues long since it rained where the discharge follows an exponential decay function.

Baseflow separation method can either be based on the assumption that such a flow responds to a storm event concurrently with surface water runoff (Nathan and McMahon, 1994), or the flow recession continues after the time when surface runoff begins. In the latter case, the baseflow recession continues after the rise of the total hydrograph due to initial outflow from stream and adjacent banks and the baseflow peaks after the total hydrograph due to the storage routing effect of the subsurface stores. The separation techniques are either by graphical means or by filtering where the filter parameter or baseflow index is determined.

The digital filter method is used in baseflow separation because high frequency waves can be associated with the direct runoff, and low frequency waves can be associated with the baseflow (Lim *et al.*, 2005). Thus, filtering direct runoff from baseflow is similar to signal analysis and processing (Nathan and McMahon, 1994; Smakhtin, 2001 and Lim *et al.*, 2005). The following equation shows the digital filter used for baseflow separation:

$$f_k = \alpha_{f_{k-1}} + (1 + \alpha)(y_k + y_{k-1})/2 \quad (2.2)$$

Where,

- α = the filter parameter,
- $\alpha_{f_{k-1}}$ = the filter parameter for the previous month
- f_k = the filtered quick response,
- y_k = the original streamflow while the filtered baseflow is $y_k - f_k$.

Baseflow analysis should be undertaken in unregulated reaches, or at least the regulated catchment area should be no more than 10% of the catchment area of the streamflow gauge (Neal *et al.*, 2004). A more general form of a digital filter considering a digital filter parameter was suggested by Eckhardt (2005) and Lim *et al.*, (2005).

$$q_{b(i)} = \frac{(1-BFI_{max})aq_{b(i-1)}+(1-a)BFI_{max}q_i}{1-aBFI_{max}} \tag{2.3}$$

Where

- $q_{b(i)}$ = the baseflow at time step i ,
- $q_{b(i-1)}$ = the baseflow at the previous time step $(i - 1)$,
- q_i = the stream flow at the current time step i ,
- a = the recession constant
- BFI_{max} = the maximum value of the baseflow index that can be measured.

The Web based Hydrograph Analysis Tool (WHAT) system was developed by incorporating digital filter methods for baseflow separation with the iSep system. To utilize the filter methods for baseflow separation, users must provide parameters. For instance, Nathan and McMahon (1990) found that the filter parameter of 0.925 gave realistic results when compared to manual separation results. The baseflow values obtained using the digital filter methods do not overestimate baseflow when multiple high peaks occur in a short time period, compared with baseflow separation using a local minimum method that typically overestimates baseflow. For manual and other baseflow separation techniques, it is the user’s responsibility to check for data gaps when raw flow data are downloaded from the United States Geological Survey (USGS) web server. The WHAT system provides a series of flow datasets. Once the baseflow is separated from the streamflow, daily, monthly, and yearly direct runoff and baseflow output values are provided in tabular format as well as a graphical hydrograph. Depending on the separation method selected, the WHAT system performs baseflow separation and prepares the tabular data and hydrograph automatically. The baseflow separation or hydrograph analysis is frequently used to calibrate and validate

the direct runoff and baseflow components in the hydrologic model by comparing the simulated values with the measured values. The WHAT system incorporates baseflow separation modules entailing digital filter methods. The advantage of this automated system eliminates the inconsistency that characterises manual hydrograph separation approaches. The other positive aspect is that the WHAT system is relatively more efficient separating baseflow from quick flow in less than a minutes, and it can also be used to calibrate and validate hydrologic model.

Xu *et al.*, (2002) asserted that numerical simulation techniques that are often used to generate groundwater discharge time series are data intensive and instead proposed that in cases where data are lacking, use of an alternative approach based on the geomorphic features of streams and use of hydrograph separation methodologies is a better option. The approach entails classification of streams in terms of Upper Catchments that are characterized by steep gradients, deep incisions and large bed-loads; middle catchments by braided channels; lower catchments by deep bed deposits and thick terraces; and special cases with special geological structures. The groundwater interactions with streams are then categorized as braided streams without bank-storage occurring in upper mountainous areas, middle courses that are controlled by bed morphology and lower streams occurring in topographically flat areas near a regional base level and which is generally a discharge area for regional groundwater systems. The methodology relies on an understanding of the typical geomorphologic stage of a particular stream and relating that to geohydrological settings and then adopting an algorithm that proposes groundwater contribution to baseflow to be previous groundwater contribution and an rainfall induced flow increment. It is important to note that the groundwater contribution determined as such is conservative since evapotranspiration from the saturated zone along the stream valley is neglected. The method should be used with caution (Xu *et al.*, 2002) and in relation to chemical or isotopic methods to confirm quantification.

$$Q_{gi} = Q_{g0} * D^i + (\sum_{j=1}^i Q^{i-j} * D^{j-1}) * I \quad (2.4)$$

Where,

Q_{gi} = baseflow,

Q_{g0} = initial average groundwater contribution,

D and I are parameters between 0 and 1 whose values depend on the interaction type.

The advantage of the baseflow separation approaches, such as event-based graphical methods and automated filter-based algorithms referred to above is that they only require stream discharge data. On the other hand the weaknesses are that they are tedious to construct and lack physical basis and the filter parameter is arbitrarily chosen (Szilagyi *et al.*, 2006). Xu *et al.*, (2002) has an added advantage of taking account of geohydrological and geomorphological aspects into account, thus contextualising the conceptual model of the system.

A wide range of empirical, conceptual and physically based modelling approaches for application in aquifer–river interaction studies are available. Each modelling approach differs in terms of the degree to which physical processes are represented, in data requirements and associated data/computational costs, and in model capabilities and form of model outputs (Ivkovic *et al.*, 2009). Each modelling approach has its strengths and limitations. Often, there is no ‘best’ model for all applications, and the most appropriate model usually depend on the intended use and data availability.

2.2.1 Evaluation of surface water - groundwater interaction using analytical models

In the case of hydraulically connected stream–aquifer systems, the resulting exchange flow is a function of the difference between the river stage and aquifer head. This is based on Darcy’s law, where flow is a direct function of the hydraulic conductivity and head difference (Sophocleous, 2002) and can be expressed as

$$q = k\Delta h, \quad (2.5)$$

Where

$$\Delta = h_a - h_r, \text{ (} h_a \text{ is aquifer head, and } h_r \text{ is river head);}$$

- q = flow between the river and the aquifer (positive for baseflow or gaining streams; and negative for recharge from a river or for losing streams);
- k = a constant representing the streambed leakage coefficient or a conductance term (i.e. hydraulic conductivity of the semi-impervious streambed stratum divided by its thickness).

This equation may be non-linear as proposed by Rushton and Tomlinson (1979) (in Sophocleous, 2002):

$$q = -k_1\Delta h + k_2(1 - \exp(-k_3\Delta h)), \quad (2.6)$$

Where

k_1 ; k_2 and k_3 are constants

Non-linearity in the relationship regarding flow could be due to resistance and change in flow rate with time at the interface.

2.2.2 Analytical solutions

Conceptual models are typically linked to analytical solutions of differential equations. A brief account of analytical models for evaluation of surface water - groundwater interaction in historical order, assumptions, limitations and degree of applicability are discussed below. Much of this part of the work is based on an extensive and comprehensive unpublished technical report by Dennis and Witthueser (2007).

2.2.2.1 Theis (1941) and Glover & Balmer (1954)

Literature studies (Swamee *et al.*, 2000; Butler *et al.*, 2001; Dennis and Witthueser, 2007; and Intaraprasong and Zhan, 2009) indicate that Theis developed the earliest analytical model to evaluate surface water - groundwater interaction, by describing the influence of groundwater abstraction on a nearby river (acting as an infinite recharge boundary). He assumed

- an infinitely long river penetrating the entire homogeneous aquifer,
- isotropic semi-infinite aquifer,

- a fully penetrating borehole pumped at a constant rate with no drawdown in the river or aquifer, and that
- release of water from storage is instantaneous.

The analytical quantification of the rate of stream depletion was expressed by Theis (Swamee *et al.*, 2000) as a fraction of a stream contribution rate Q_s in the total uniform discharge Q (m³/s) from a borehole that is steadily discharging from a homogeneous, uniform, isotropic aquifer hydraulically connected to a fully penetrating stream,

$$Q_s = \frac{q}{Q} = \frac{2}{\pi} \int_0^{\pi/2} \exp(-\alpha \sec^2 u) du \tag{2.7}$$

Where

- q = discharge contributed by stream (m³/s); $\alpha = \frac{a^2 S}{4 T t}$ (dimensionless);
- a = effective distance between the pumping borehole and the recharging stream (m);
- S = aquifer storage coefficient (dimensionless);
- T = aquifer Transmissivity (m²/s);
- t = time since the start of pumping (s);
- u = dummy variable.

The conceptual model is illustrated in Figure 5.

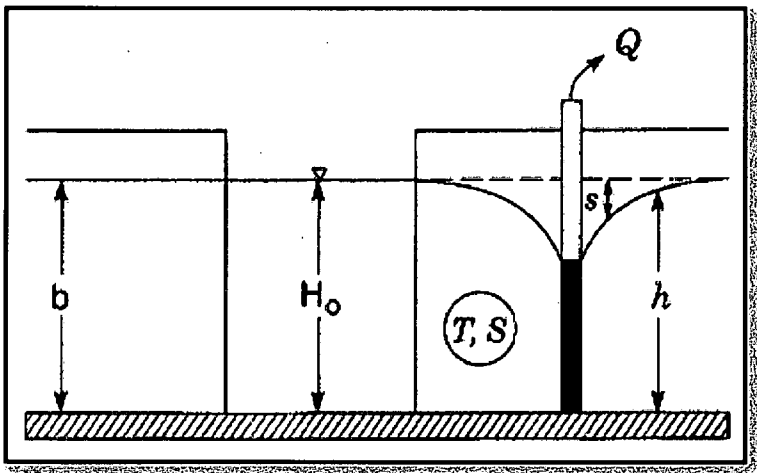


Figure 5: Conceptual model used by Theis (1941) to model streamflow depletion (Butler *et al.*, 2001 & 2007)

The solution to Theis's integral was then determined by Glover and Balmer (1954) using the complimentary error function (*erfc*):

$$\frac{\Delta Q}{Q_{abs}} = \text{erfc} \left[\sqrt{\frac{Sd^2}{4Tt}} \right] \quad (2.8)$$

Where

ΔQ = the stream depletion rate or leakage,

Q_{abs} = the constant (from $t = 0$ to $t = \infty$, i.e. continuous) abstraction rate at the borehole,

S = the aquifer storage coefficient or specific yield,

T = the aquifer transmissivity,

t = the time and

d = the shortest distance between the borehole and the stream.

The complementary error function is given by the dummy variable u

$$\text{erfc}(z) = 1 - \frac{2}{\sqrt{\pi}} \int_0^z e^{-u^2} du \quad (2.9)$$

There is no clogging layer in this simple model hence the solution provides conservative estimates, and at steady state the induced leakage equals the abstraction rate of the borehole.

2.2.2.2 Hantush

Hantush (1965) improved on this work by providing an analytical solution for a fully penetrating stream and identical set of conditions considered by Theis (1941) and Glover & Balmer (1954) with the addition of a clogging layer (i.e. a vertical layer of semi-permeable material) lining the stream bed of thickness b' and permeability k' . Figure 6 depicts the conceptual model with aquifer thickness b , rest water level in the stream H_o , depth to water level in the aquifer, s and hydraulic head, h .

2.2.3 Jenkins and Wallace

Jenkins (1968) used the solution of Theis (1941) and Glover & Balmer (1954) superposition and time translation to estimate residual impacts after pumping stopped to evaluate the effects of limited pumping schedules. While Wallace *et al.*, (1990) investigated the impacts of cyclic pumping using these solutions. Impacts develop over several annual cycles with maximum depletion in later years due to superposition of impacts. The volume of stream depletion factor as a function of time for continuous pumping is

$$V = Q_{abs} t 4i^2 \operatorname{erfc} \left[\sqrt{\frac{Sd^2}{4Tt}} \right] \quad (2.13)$$

Where

$i^2 \operatorname{erfc}$ = the second repeated integral of the complementary function.

Jenkins proposed the concept of stream depletion factor sdf which is a function of *aquifer diffusivity* (T/S) and the shortest distance d between the borehole and the stream and describes the response time. The stream depletion factor (Pattle Delamore Partners Ltd; 2000) is given by

$$sdf = \frac{d^2 S}{T} \quad (2.14)$$

2.2.4 Grigoryev and Bochever

Grigoryev (1957, cited by Butler *et al.*, 1999; as well as Dennis and Witthueser, 2007) developed a steady-state solution for a partially penetrating stream. The model is based on the assumptions that the river penetrates only a small proportion of the full aquifer thickness and that the stream and aquifer are separated by a thin zone of relatively low hydraulic conductivity (clogging layer). The equation is not given. Bochever (1966, in Butler *et al.*, 1999 and Dennis and Witthueser, 2007) used a more realistic conceptual model of a partially penetrating stream, i.e. a stream with shallow penetration depth compared to aquifer thickness. A thin semi-pervious layer (clogging layer) separates the river from the aquifer. T_i and S_i are Transmissivity and Storativity of the aquifer, ω is the width of the stream while 1, 2 and 3 are various zones as indicated in the Figure 7.

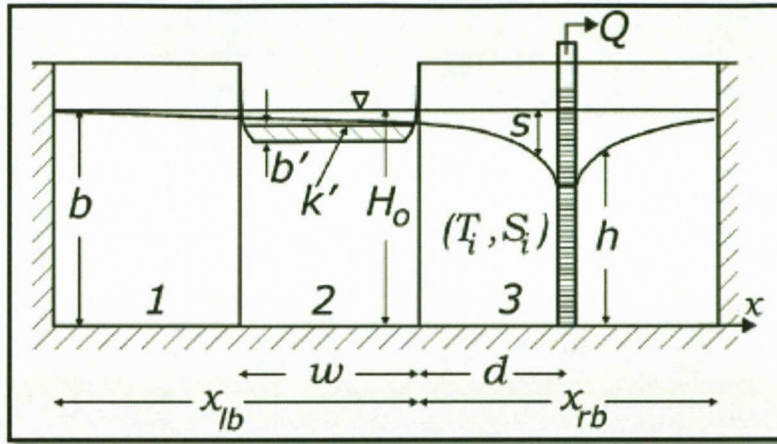


Figure 7: Conceptual model used by Grigoryev (1957) and Bochever (1966)

2.2.5 Glover

Glover (1974), in Dennis and Witthueser, (2007) improved on his earlier equation (2.8) by introducing an analytical solution for a finite stream reach (from $-x$ to $+x$),

$$\frac{\Delta Q}{Q_{abs}} = \frac{2}{\pi} \int_0^x \frac{de^{-x^2}}{x^2 + d^2} dx \quad (2.15)$$

$$z^2 = \frac{(x^2 + d^2)S}{4tT}$$

This, at a steady state reduces to

$$\frac{\Delta Q}{Q_{abs}} = \frac{2}{\pi} \arctan\left(\frac{x}{d}\right) \quad (2.16)$$

This equation can be used to show the relationship between streamflow depletion and normalised distance from the stream to the borehole. A plot of showing such a relationship

$\left(\frac{\Delta Q}{Q_{abs}} \text{ versus } \frac{x}{d}\right)$; is on Figure 8.

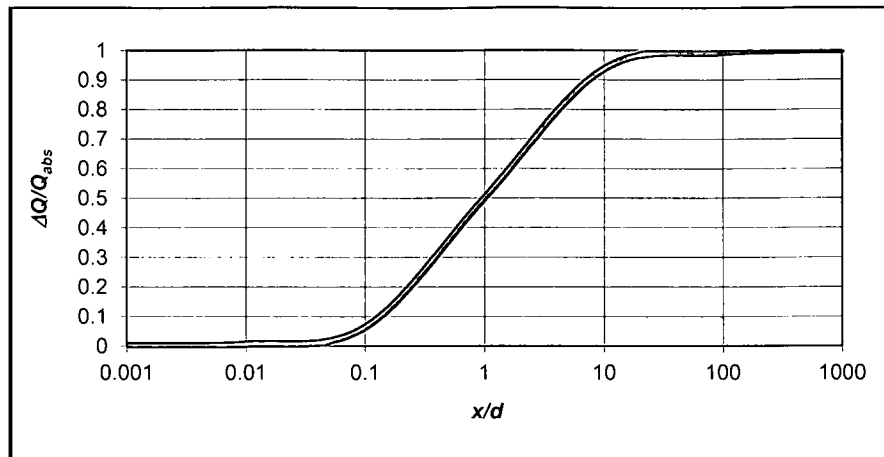


Figure 8: Streamflow depletion versus normalized distance to the borehole (Dennis and Witthueser, 2007)

The figure indicates that streamflow depletion equals abstraction at steady state only if the distance to the borehole is approximately less than 1 % of the length of the stream reach ($\frac{x}{a} = 100$). If the distance to the borehole equals the length of the stream reach ($\frac{x}{a} = 1$), the abstraction borehole receives equal fluxes from the river and the aquifer.

Most importantly is that Glover (1974) developed transition curves from groundwater abstractions to surface water depletion for 2-dimensional homogeneous isotropic aquifer. Sophocleous *et al.*, (1995) evaluated the applicability of Glover's model for more complex geohydrological settings by comparing the analytical results to the numerical model results and ranked the relative importance of the assumptions of the analytical solutions as:

1. Streambed clogging
2. Degree of stream partial penetration
3. Aquifer heterogeneity
4. Non-equilibrium losing stream
5. Aquifer storativity
6. Aquifer hydraulic conductivity, gaining stream, partial borehole penetration.

Streambed clogging, degree of stream partial penetration and aquifer heterogeneity are the most significant factors causing overestimation of stream leakage by analytical solution.

2.2.6 Stang and Hunt

The analytical solution for a conceptual model by Stang (1980) and also proposed by Hunt (1999) is equivalent to that of Grigoryev and Bochever (i.e. equation 2.12). The assumptions are that the aquifer is homogeneous, isotropic and of infinite extent with dominant lateral flow, constant transmissivity and overlain by a thin stream with impervious layer. That is

$$\frac{\Delta Q}{Q_{abs}} = \operatorname{erfc} \left[\sqrt{\frac{Sd^2}{4Tt}} \right] - \exp \left[\frac{\lambda^2 t}{4ST} + \frac{\lambda d}{2T} \right] \operatorname{erfc} \left[\sqrt{\frac{\lambda^2 t}{4ST}} + \sqrt{\frac{Sd^2}{4Tt}} \right];$$

This equation is equivalent to Hantush's solution (i.e. equation 2.10), where λ (a constant of proportionality between the seepage rate per unit distance through the streambed and the difference between river and groundwater levels), is defined as:

$$\lambda = 2 \frac{T}{L} \quad (2.17)$$

That means the solution for a fully penetrating streambed (2.12) can be applied to a partially penetrating streambed (2.10) by expressing the stream leakance L in terms of λ and T using (2.17) instead of modifying distance d between the stream and the borehole.

2.2.7 Wilson

Wilson (1993) developed a steady-state solution (2.18) for a fully penetrating stream fed by an aquifer (Figure 9), where q_u is inflow from the aquifer. Wilson also regards streamflow depletion as induced infiltration. The amount of induced infiltration is a function of many factors, including aquifer hydraulic conductivity, aquifer geometry, borehole pumping rate, the strength of the hydraulic connection between the aquifer and stream due to stream penetration and clogging layer, and the presence of other sources of water supplying the borehole.

$$\frac{\Delta Q}{Q_{abs}} = \frac{2}{\pi} \left[\frac{-(\beta-1)^{0.5}}{\beta} \right] + \tan^{-1}(\beta-1)^{0.5} \quad (2.18)$$

Where;

$$\beta = \frac{Q_{abs}}{\pi q_u d}$$

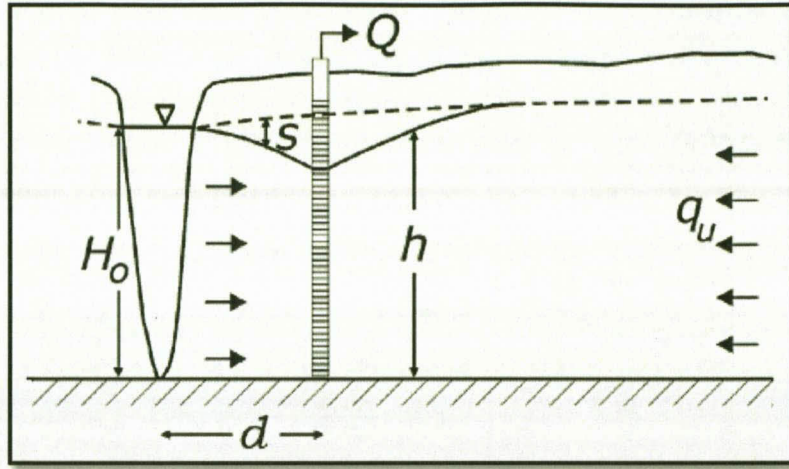


Figure 9: Conceptual model by Wilson (in Dennis and Witthueser, 2007)

2.2.8 Zlotnik

Zlotnik *et al.*, (1999) developed a transient solution to Grigoryev and Bochever model (Figure 7) assuming a stream of finite width, shallow penetration, separated from the aquifer by a low permeability streambed with stationary flow. They present analytically inverted solutions for two simplified cases. The first is the special case where the storativity of the confined aquifer under the streambed is neglected (zone 2 of Figure 7) and the stream depletion is then given by:

$$\frac{\Delta Q(\bar{t})}{Q_{abs}} = \operatorname{erfc} \left[\frac{d}{w\sqrt{\bar{t}}} \right] - \exp(b(b\bar{t} + 2d/w)) \operatorname{erfc} \left[b\sqrt{\bar{t}} + \frac{d}{w\sqrt{\bar{t}}} \right] \quad (2.19)$$

$$b = \omega \beta cth(2\omega) \left[1 - \frac{1}{csh(2\omega)} \right]$$

$$\omega = \frac{k'(2w)^2}{T_{zone2}b'}$$

$$\bar{t} = \frac{T_{zone3}t}{S_w^2}$$

The second case is where in addition to zero storativity the width of the stream is assumed to be infinite and in such a case the streamflow depletion is given by:

$$\frac{\Delta Q(\bar{t})}{Q_{abs}} = \operatorname{erfc}\left[\frac{d}{w\sqrt{\bar{t}}}\right] - \exp\left(\lambda\left(\lambda\bar{t} + \frac{2d}{w}\right)\right) \operatorname{erfc}\left[\lambda\sqrt{\bar{t}} + \frac{d}{w\sqrt{\bar{t}}}\right] \quad (2.20)$$

Where,

$$\lambda = \frac{k'(2w)^2}{T_{zone3}b'}$$

$$\bar{t} = \frac{T_{zone3}t}{S_w^2}$$

Note that,

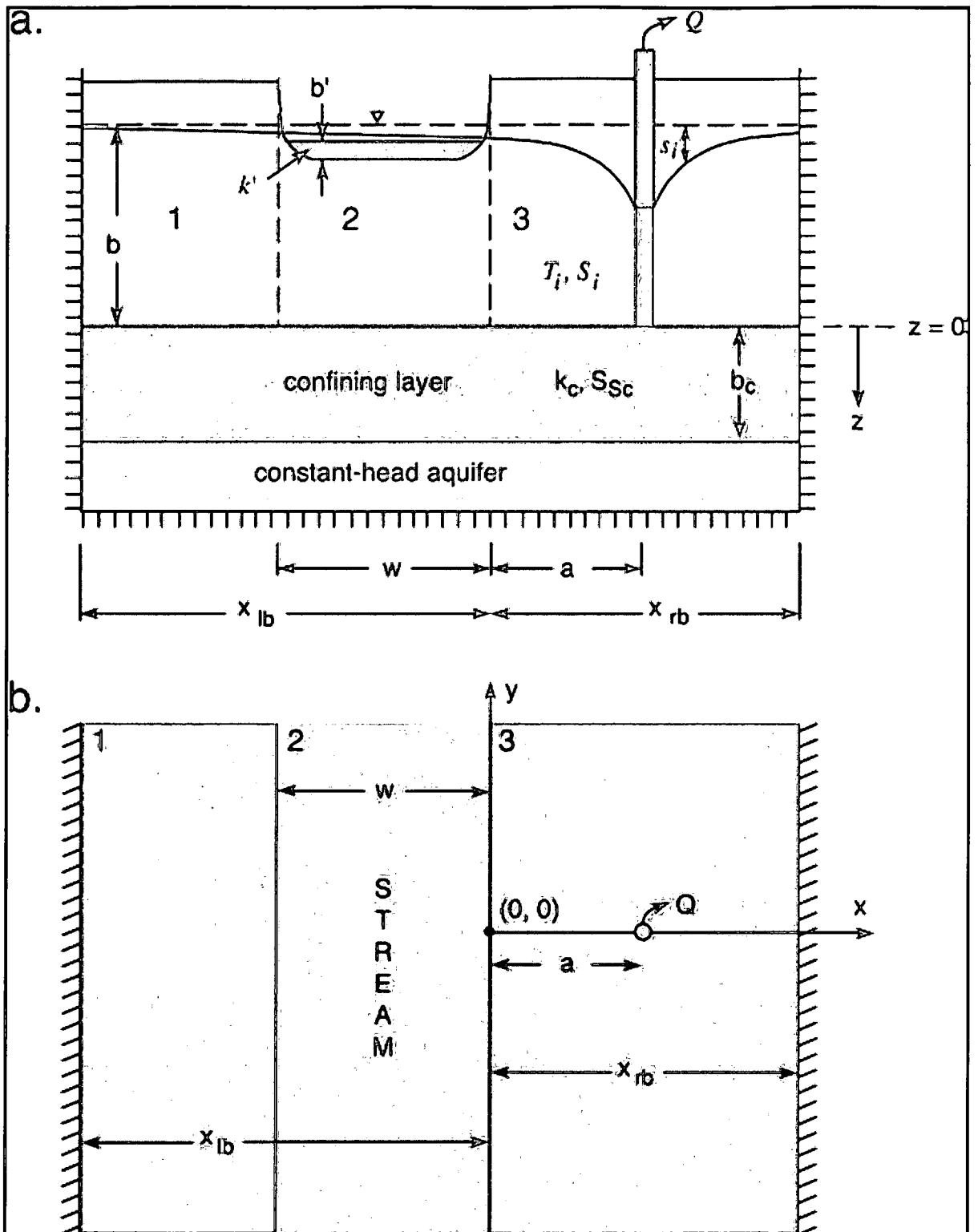
w = width of the stream, while ω is a derived value (as in Figure 2.19).

However, Zlotnik *et al.* (1999) showed that, while the storativity of the aquifer below the streambed could for all practical purposes be neglected; the leakance parameter influences the stream depletion significantly and must be considered.

2.2.9 Butler

In their paper Butler *et al.*, (2001) used the steady state Grigoryev-Bochever model as a starting point for the development of a general analytical solution for transient drawdown and stream depletion. Their analytical solution also considered, in addition to a partially penetrating stream underlain by a semi-pervious layer, an aquifer of finite lateral extent (Figure 10a & b). It is worth noting that $\underline{a}$ is the distance to the borehole in the figure, instead of $\underline{d}$ used in previous figures.

Their model assumes that a vertical flow is negligible, the aquifer is isotropic, the stream level is not affected by pumping, the borehole is screened and fully penetrating and the aquifer heads remain above the stream bottom.



2.2.10 Damara

Damara determined the stream depletion by cyclic pumping of boreholes in a semi-infinite aquifer, bounded by a partially penetrating stream with a semi-pervious bed, using Hantush's solution (1965, in Dennis and Witthueser, 2007) and superposition principles. The total stream depletion as a result of the superposition of stream depletions of cyclic pumping schedules is given by Figure 11.

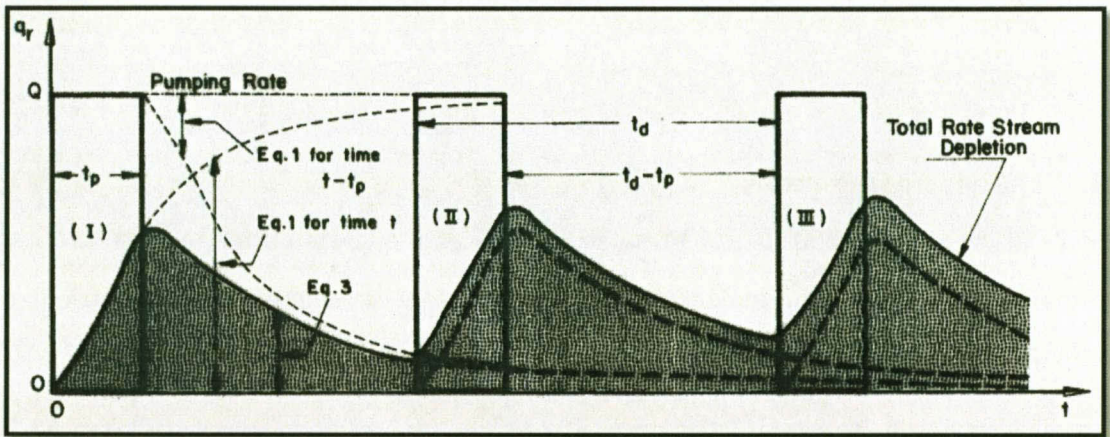


Figure 11: Stream depletion resulting from a series of cyclic pumping (Damara, 2001; in Dennis & Witthueser, 2007)

The stream depletion during pumping is given by

$$\frac{\Delta Q}{Q_{abs}} = \operatorname{erfc}(U) - \exp(-U^2 + (U + w)^2 \operatorname{erfc}(U + w) - (\operatorname{erfc}(U_p) - \exp(-U_p^2 + (U_p + w_p)^2 \operatorname{erfc}(U_p + w_p))) \quad (2.21)$$

$$t_p \leq t \leq \infty$$

$$U = \sqrt{\frac{t_a}{4t}}$$

$$t_a = \frac{d_{eff}^2 S}{T}$$

$$w = \frac{R}{2U}$$

$$R = \frac{d_{eff}k'}{Kb'}$$

$$U_p = \sqrt{\frac{t_a}{4(t - t_p)}}$$

$$w_p = \frac{R}{2\sqrt{\frac{t_a}{4(t - t_p)}}}$$

2.2.11 Bakker and Anderson

Bakker and Anderson (2003) derived a steady-state analytic solution in the form of a complex potential for borehole extraction adjacent to a partially penetrating narrow stream (Figure 12) with clogging layer in a semi-infinite aquifer.

$$\Omega = \frac{Q}{2\pi} \ln \frac{z - z_d}{z - \bar{z}_d} - \frac{Q}{2\pi} \exp(Z_1) E_1(Z_1) - (U - iV_1)z + A; \quad y \leq 0$$

$$\Omega = -\frac{Q}{2\pi} \exp(Z_2) E_1(Z_2) - (U - iV_2)z + A \quad y \geq 0 \quad (2.22)$$

where

$$Z_1 = \frac{iB}{cT} (z - \bar{z}_d)$$

$$Z_2 = -\frac{iB}{cT} (z - z_d)$$

where an overbar denotes the complex conjugate, and where A is a real constant given by

$$A = \frac{cT}{2B} (V_1 - V_2) + \Phi_0$$

$E_1(z)$ is the exponential integral of z as defined Abramowitz and Stegun (1965):

$$E_1(z) = \int_z^\infty \frac{\exp(-t)}{t} dt$$

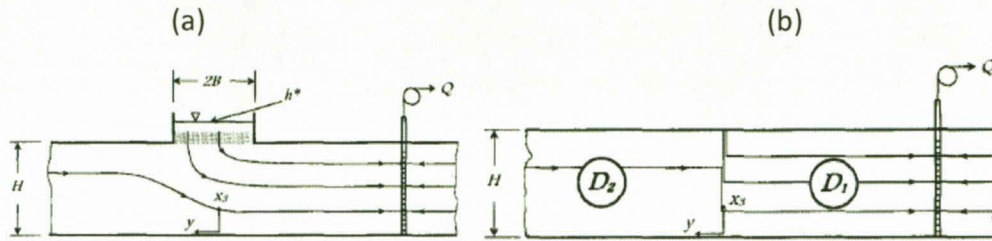


Figure 12: Pathlines in a vertical cross section through the stream and the well (Bakker and Anderson, 2003):

(a) the exact behaviour with distributed leakage; and (b) the approximate behaviour when leakage is lumped at the centre of the stream

The solution allows for different head gradients towards the river on either side of the stream as well as for a surface water gradient. They showed that if river water is captured by the borehole then the capture zone will always extend beyond the far side of the stream.

2.2.12 Chen and Yin

The semi-analytical solution of Chen and Yin (2004) is an extension of Hunt's (1999) solution for a partially penetrating stream, adding a gradient between the aquifer and the gaining stream. The total stream depletion ΔQ is subdivided into induced stream infiltration Q_s and baseflow reduction Q_b (interception of some or all of the regional groundwater flow, which otherwise would have discharged into the river), Figure 13.

Equation (2.24) indicates that the stream infiltration and baseflow reduction in a gaining stream are affected by Δh besides aquifer properties and stream leakage.

$$\Delta Q = \lambda \int_{-\infty}^{+\infty} \phi(0, y, t) dy$$

$$\phi(0, y, t) = \frac{Q}{4\pi T} \left\{ W \left(\frac{d^2 + y^2}{4Tt/S} \right) - \int_0^\infty e^{-\theta} W \left(\frac{(d + 2T\theta/\lambda)^2 + y^2}{4Tt/S} \right) d\theta \right\} \quad (2.23)$$

$$W(u) = \int_u^\infty \frac{e^{-u}}{u} du \quad (\text{Theis well})$$

$$\lambda = \lim_{\substack{w \rightarrow 0 \\ k'/b' \rightarrow \infty}} \frac{w}{b'} K' \quad (\text{Streambed leakage})$$

per unit length of stream reach)

$$Q_s = \lambda \int_{-y}^{+y} (\phi(0, y, t) - \Delta h) dy$$

$$Q_b = \Delta Q - Q_s \quad (2.24)$$

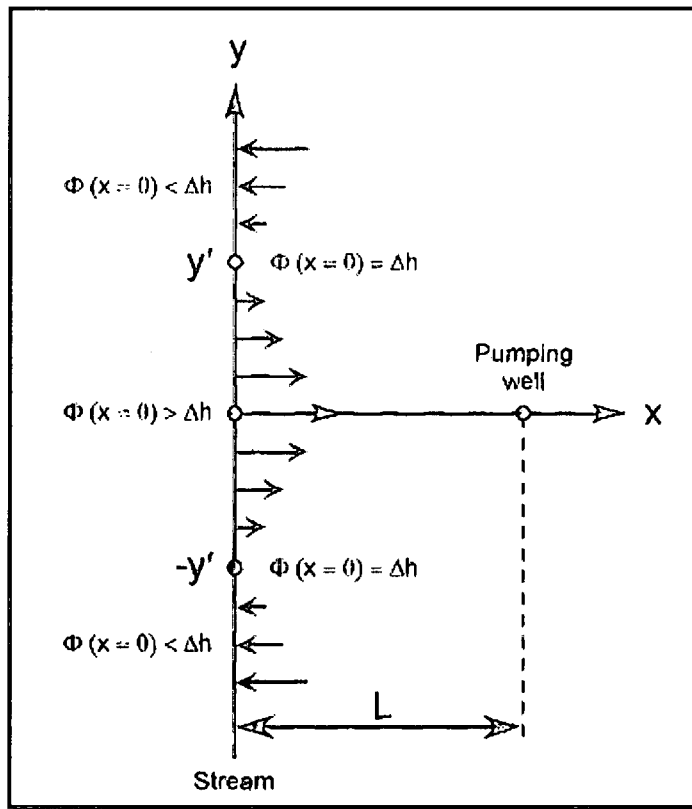


Figure 13: Schematic diagram showing the dividing points (y' & $-y'$). Stream infiltration has been induced by pumping well for the reach between $-y'$ and y' (Chen & Yin, 2004)

Based on this subdivision, they could show that baseflow reduction accounts for a large, percentage of total stream depletion if the ratio of $\frac{\lambda L}{T}$ is large, whereas induced stream infiltration becomes of more importance if the ratio is small. If the ratio is larger than 9, almost no stream infiltration is induced at all (zero for a ratio of 500). The authors propose this limit to avoid groundwater contamination by surface water.

The total depletion calculated (Chen & Yen, 2004) using the Theis (1941) model is plotted in Figure 14 for comparison, and indicates that the Theis model can provide an overestimated depletion when it is used for a partially penetrating stream.

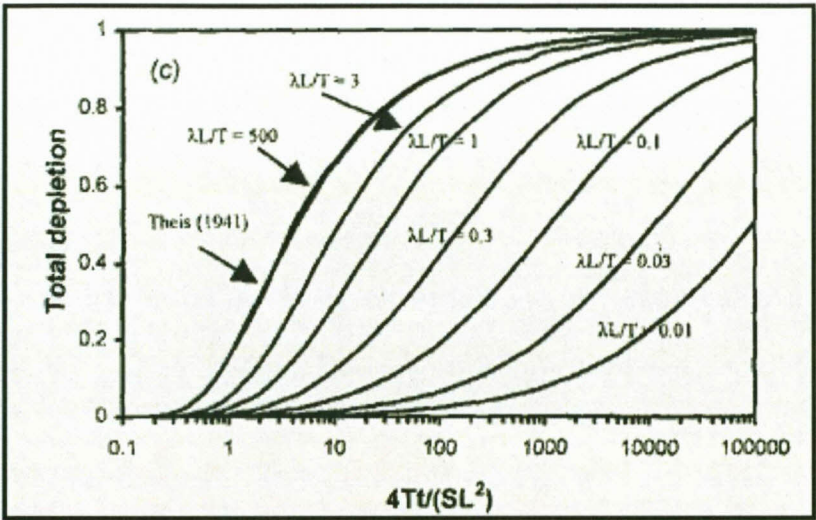


Figure 14: Dimensionless depletion curve for a total depletion given by equation (2.24) (Dennis and Witthueser, 2007)

Their analytical results verified the numerical simulation results obtained by MODFLOW for both pumping and post pumping periods.

2.2.13 Di Matteo & Dragoni

Di Matteo and Dragoni (2005) undertook a study to investigate an empirical relationship, indicating under steady-state conditions, stream depletion for a stream partially penetrating the aquifer and for a borehole that penetrate the full thickness of the aquifer and screened at the bottom. In this case the vertical component of flow close to the stream and to the borehole is relevant. The solution was also based on MODFLOW model runs using the conceptual and numerical model given in Figure 15. Parameters as indicated in the figure are as follows: h is simulated head, S is drawdown while h_r is the river depth.

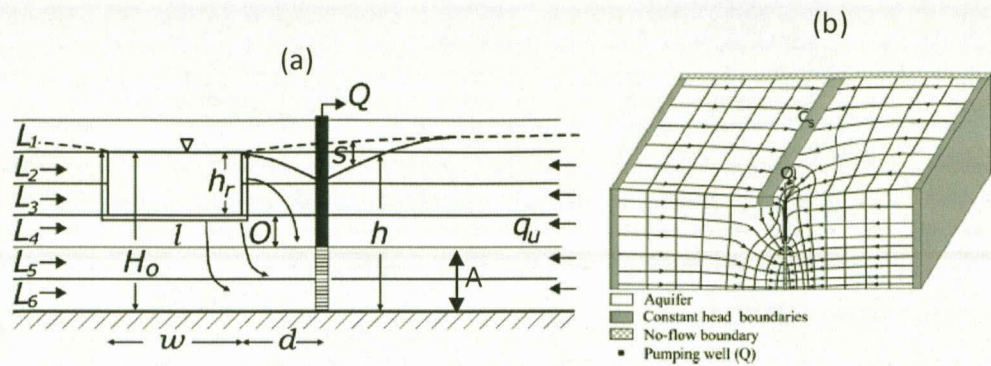


Figure 15: Conceptual model (a) and numerical model (b), used by Di Matteo and Dragoni (2005)

Simulations were obtained by placing the constant head boundaries at 1 000m from the river (Di Matteo & Dragoni, 2005). The authors used a trial-and-error method to analyse the numerical model results $\left(\frac{\Delta Q}{Q_{abs}}\right)$ with linear regression models for streamflow depletion. They give the best relationship as:

$$\frac{\Delta Q}{Q_{abs}} = 1 + (15.53M^{2.165} - 6.92M)$$

$$M = 0.42 \sqrt{\frac{q_u H_0}{Q}} \left[\frac{d}{H_0}\right]^{0.277} \left[\frac{A}{H_0}\right]^{0.137} \left[\frac{l_w}{H_0}\right]^{-0.41} \left[1 \pm \frac{O}{H_0}\right]^{-0.53} \quad (2.25)$$

d = river-borehole distance,

q_u = unit inflow from aquifer into the river,

Q = pumping rate,

H_0 = aquifer thickness,

A = length of screen

l = wetted perimeter of river,

O = overlap vertical distance between the top of the well screen and riverbed.

2.2.14 Comparison of Analytical Methods

Analytical methods are based on various assumptions. Common assumptions among analytical methods discussed here are:

- Isotropic aquifers of uniform thickness

- There are no perched rivers or groundwater table is always above streambed
- Horizontal groundwater flow and no borehole losses
- Transmissivity is independent of head
- Stream water level is not affected by pumping

However, limiting factors are that:

- In real situation, the groundwater table is not always above the streambed
- Aquifers are not really isotropic especially for South African conditions
- Modelling disconnected rivers can only be done in consideration of the clogging layer

Table 1 compares various models with respect to assumptions upon which their based.

Table 1: Comparison of analytical models (after Dennis & Witthueser, 2007)

Assumptions	Theis/Glover & Balmer	Hantush	Jenkins – Wallace et al	Grigorev & Bochever	Stang/ Hunt	Wilson	Zlotnik et al	Butler et al	Darama	Bakker & Anderson	Chen & Yin	Di Matteo & Dragoni
Homogeneous aquifer	X	x	x			x			x	x	x	
Semi-infinite areal extent	X	x	x	x	x		x		x	x	x	
Horizontal rest water level	X	x	x	x	x		x	x	x			
Fully penetrating well	X	x	x	X	x	x	x	x	x	x	X	
Steady state				X		x				x		x
Clogging layer		x		x	x	x	x	x	x	x	x	
Fully penetrating stream	X	x	x			x			x			
Cyclical pumping			x						x			
Semi-analytical							x	x			x	
Empirical												x

The Table is handy for deciding on the model to use under particular conditions. For instance, in case of a fully penetrating borehole, any of the listed analytical models with the exception of DiMatteo & Dragoni would be the appropriate for consideration.

Besides system components including the surface runoff, unsaturated zone and groundwater this chapter also covered the literature review on the historical evolution and application of various analytical models. Although analytical models are based on some assumptions and hence have inherent limitations, some are relevant and useful to South African hydrologic conditions subject to the conceptual model of the concerned system.

2.3 Evaluation of surface water - groundwater interaction using numerical methods

There are two main categories of numerical modelling methods, namely the gridded or discretized method that entails grids or a mesh of small elements, and the non-gridded or mesh free method, also called the boundary element method which is only discretized at boundaries or along flow elements. In this study only gridded methods will for illustrative purposes be discussed to develop surface water - groundwater interaction methodologies.

The gridded methods such as the finite element or finite difference methods solve the groundwater flow equation by breaking the problem domain into small elements such as squares and blocks. The flow equation is then solved for each element where all material properties are assumed constant or linearly variable within an element, and then linking together all elements using conservation of mass across boundaries between elements.

A finite difference is a mathematical expression of a general form:

$$\Delta_h^\mu[f](x) = \sum_{k=0}^N \mu_k f(x + kh), \quad (2.26)$$

Where

$\mu = (\mu_0, \dots, \mu_N)$ is its coefficient vector.

The finite difference method is based on the representation of continuous differential operators derived from Taylor series using discrete intervals Δx and Δt . For instance the

first-order time derivative is usually approximated using a forward finite difference, where the subscripts indicate a discrete location,

$$\frac{\partial h}{\partial t} = h'(t_i) \approx \frac{h_i - h_{i-1}}{\Delta t} \tag{2.27}$$

MODFLOW, developed by the United States Geological Survey (USGS) as a modular and extensible simulation tool, is an example of a general finite difference groundwater flow model.

Water is considered incompressible in many cases and, as such, the density thereof does not depend on pressure. Hence the mass fluxes across boundaries become volume fluxes, and the groundwater flow equation describing a flow regime is characterised by a mathematical statement that indicates that a change in hydraulic head (Δh) with time, is proportional to the hydraulic diffusivity (α) and the hydraulic head (h).

$$\frac{\partial h}{\partial t} = \alpha \nabla^2 h - G \tag{2.28}$$

$$\frac{\partial h}{\partial t} = \alpha \left[\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} \right] - G \tag{2.29}$$

where hydraulic diffusivity is the ratio of transmissivity to storativity ($\alpha = \frac{T}{S}$); G is the sink/source term and Laplace operator, Δ is:

$$\frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} + \frac{\partial^2}{\partial z^2} = \nabla \cdot \nabla = \nabla^2 \tag{2.30}$$

MODFLOW code discretizes and simulates an orthogonal form of the governing flow equation. Figure 16 shows a 3-D finite difference grid used in MODFLOW

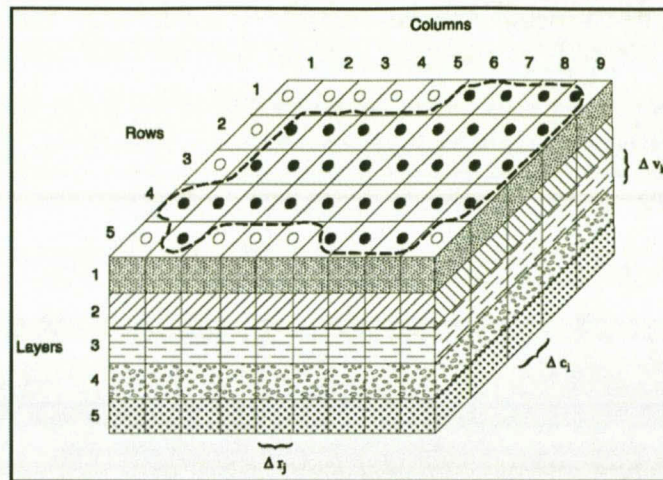


Figure 16: A 3-D Finite Difference grid used in MODFLOW (McDonald & Harbaugh, 1988)

The finite element method, on the other hand is more flexible in design (triangular elements) and also commercially available (e.g. FEFLOW). For illustrative purposes, the finite difference method will be discussed in the next sections.

2.3.1 Finite difference method

Choice of an appropriate numerical method for simulation of surface water - groundwater interaction is informed by the conceptual model of the system in the study area. In other words the first approach should be the development of the physical and the mathematical conceptual model of the system. Any numerical groundwater flow simulation requires a good conceptual geohydrological model and the development of a systematic database accounting for all model input parameters (Ayenew and Tilahun, 2008). This task may require collection of field data and information, as well as reference to published information products such as maps and databases. The conceptual model is a descriptive representation of our geohydrological understanding of how water flows into, through and out of a groundwater system (Water Corporation, 2005). This is usually presented in the forms of diagrams and maps of the physical characteristics of the geological formation geometry and flow systems/directions. The mathematical model is the assembly of numerical data sets that reflect the conceptual model, and the computer coded package of equations that enable calculation of changes in water level or pressures in the various geological strata to be calculated. Then with all input parameters, meteorological and

hydrological data considered and available, one may identify appropriate numerical models to use for simulation, calibration and prediction or for use as required.

Among a number of widely used and available surface water-groundwater interaction models is the MODFLOW in combination with the RIVER, DAFLOW, BRANCH or STREAM add-on packages. Numerical models of surface water - groundwater interaction couple numerical solutions of equations for surface water routing and groundwater flow using mostly a Darcian approach to model vertical exchanges between the riverbed and the aquifer. Few numerical codes (e.g. SHETRAN, ICMM) model actual three-dimensional surface-groundwater coupling (Environment Agency, 2002b).

2.3.2 Coupling surface water - groundwater models

Full coupling of surface and subsurface flow regimes is accomplished implicitly by simultaneously solving one system of nonlinear discrete equations describing flow and transport in both flow regimes. Obviously these full-fledged and well-packaged models can offer powerful modelling capabilities for sophisticated simulation of watershed hydrological processes (Chen et al, 2007). However, their applicability in a large watershed is sometimes seriously limited because physical parameters of basin characteristics are simply not available.

Coupled hydrologic models for simulation of surface water - groundwater interaction may generally be based on either of the two options. One is referred to as a fully integrated approach whereby equations governing surface and subsurface flows are solved simultaneously (Markstrom *et al.*, 2008). The second approach partitions the surface and subsurface systems into separate regions and the governing equations that describe flow in each region are integrated (coupled) using iterative solution methods. The fully integrated approach uses the three-dimensional form of Richards' equation to simulate unsaturated and saturated flow. This approach requires much finer spatial grids and smaller time steps than typically are used to simulate saturated flow, which limits their applicability for simulating flow through regional hydrologic systems that encompass hundreds to thousands of square kilometres and simulation time periods of months to decades. Hence for the purpose of this study, the focus will be on coupled rather than fully integrated models.

An example of coupled surface water - groundwater interaction model is the GSFLOW model that was developed by USGS to simulate coupled groundwater surface water flow in one or more watersheds by simultaneously simulating flow across the land surface and within subsurface saturated and unsaturated materials. GSFLOW can be used to evaluate the effects of such factors as land-use change, climate variability, and ground-water withdrawals on surface and subsurface flow. It was designed to simulate the most important processes affecting surface-water and groundwater flow using a numerically efficient algorithm. The model incorporates well documented methods for simulating runoff and infiltration from precipitation, as well as the interaction of surface water with groundwater in large scale watersheds.

The GSFLOW model is an integration of the U.S. Geological Survey (USGS) Precipitation-Runoff Modeling System (PRMS) with the 2005 version of the USGS Modular Ground-Water Flow Model (MODFLOW-2005). PRMS and MODFLOW have similar modular programming methods, which allow for their integration while retaining independence that permits substitution of additional PRMS modules and MODFLOW packages (Markstrom *et al.*, 2008).

2.3.3 Setting up a finite difference numerical model

In constructing a finite difference numerical model to represent the spatial variability in the system being modelled, the area is overlain by a rectangular grid, with each cell in the grid representing a point at the centre of that cell (block centred grid) or at the intersection of the grid lines (mesh-centred grid). These are node points at which the solution of the unknown values, such as the water fluxes is sought or determined. The choice of the type of the grid mainly depends on boundary conditions. For instance it's more convenient to use block centred grids in cases where the flux is specified while mesh centred types could be more useful where values of the head are specified. The grid can be applied to a number of layers if the hydraulic characteristics of the geological section vary with depth. Each layer can represent the various formations that occur within the model domain. Each individual cell, within each layer is then assigned a range of properties determined from the investigation work, such as the ability to store water, the ability to transmit water (horizontally and vertically), the vertical thickness of the formation at that point, and water

level or pressure (Water Corporation, 2005). Hence the general assumption is that all discharge from or recharge to the nodal area occurs at the node point and that water levels in the entire nodal area are the same as at that node point. In fact each cell is a hydrologic response unit (HRU) and the discretization is based on hydrologic and physical characteristics such as drainage boundaries, land-surface altitude, slope, and aspect; plant type and cover; land use; distribution of precipitation, temperature, and solar radiation; soil morphology and geology; and flow direction. Each HRU is assumed to be homogeneous with respect to these hydrologic and physical characteristics and to its hydrologic response (Markstrom *et al.*, 2008).

As part of the discretization process, rivers or streams are divided into reaches and segments where a reach is defined as a section of a stream that is associated with a particular finite-difference cell. More than one stream reach can be assigned to a particular finite-difference cell, but only one finite-difference cell can be assigned to a single reach. Reaches are grouped into segments that represent lengths of the stream between connections with another stream or tributary, a lake, or a watershed boundary. User-specified inflows to a stream that are external to inflows calculated by the model are added to the stream at the upstream end of a segment. Specified outflow at the upstream or downstream end of a segment can be used to divert water from a stream to a pipeline or lined canal; the water that is diverted in this way is removed from the modelled area without surface water - groundwater interaction.

2.3.4 Some considerations in the design of grids

In areas where more detail or higher level of accuracy on groundwater responses is required, a closer spacing is applied to the grid. This is often the case closer to the river or recharge boundary, where more detailed information is required and the grid is accordingly more tightly spaced than at a distance away from the river. Boreholes' nodes are placed near the physical location of pumping boreholes or centre of the borehole field. Grids should be oriented with major trends as much as possible.

To reduce the possibility of mathematical instabilities, the ratio of length to width of cells should be kept as close to one as possible. Otherwise, long linear cells are likely to lead to numerical errors.

2.3.5 The initial conditions, boundary conditions and model input parameters

Initial conditions for changes in hydraulic head or drawdown are often assumed zero everywhere since these are caused usually by pumping and computed drawdown can be superimposed on natural flow system. In water-table conditions, a head distribution should be specified as an initial condition. If available observation data are limited, interpolation techniques using the relationship between water levels and altitude may be used to estimate initial hydraulic head values. On the other hand a transient simulation should start from a steady state position.

The nature or type of boundary conditions can be deduced from field evidence or a geohydrological conceptual model of the system. Cells used for simulation of boundary conditions may either be specified head (i.e. constant head) such as is the case in the boundary between the aquifer and the river; or no flow cells (where flow into cells or flow out of cells is not allowed). The rest of the cells are variable-heads cells where groundwater head, vary with time and are thus computed. Time-variant inflow and (or) outflow boundary conditions can be assigned to variable-head cells using MODFLOW stress packages such as the Well Package, where the stress rates only can a change at the beginning of a stress period. Another type of possible boundary condition applied to variable-head cells is the head-dependent flow boundary condition. Interaction of groundwater with streams and lakes is an example of a head-dependent flow boundary condition (Markstrom et al., 2008). The input data is often estimated from pump testing results, previous studies, and observation measurements or adjusted through calibration.

2.3.6 Model simulation of water fluxes

The key for integrating the two models is to simulate the water fluxes, which link the two storage reservoirs together. The entire hydrological cycle involves water fluxes at three major interfaces: 1) evapotranspiration and surface runoff at the interface between atmosphere and soils; 2) groundwater recharge and loss at the interface between unsaturated and saturated zones; and 3) water exchanges (baseflow or groundwater recharge) at the interface between aquifer and stream channel (Chen et al, 2007). Parameter values are assigned for each cell in the entire model prior to steady state simulation.

2.3.7 The RIVER package and the STREAM package

The RIVER package and the STREAM package are the two MODFLOW packages that can be used to specify the cells in which the stream occurs, the stage height of water in the stream, the height of the bottom of the streambed and the conductance of the streambed (Pattle Delamore Partners Ltd; 2000). The flow between the stream and the aquifer Q_{riv} is calculated as:

$$Q_{riv} = C_{riv}(H_{riv} - h) \text{ for } h > R_{bot} \quad (2.31)$$

Where

- C_{riv} = the hydraulic conductance of the stream aquifer interaction;
- H_{riv} = the water level in the stream or river;
- h = the hydraulic head
- R_{bot} = the height of the bottom of the streambed.

The flow to and from the river varies depending on the head in the aquifer.

$$Q_{riv} = C_{riv}(H_{riv} - R_{bot}) \text{ for } h \leq R_{bot} \quad (2.32)$$

The hydraulic conductance, using Darcy's law can be calculated as:

$$C_{riv} = \frac{k' L_c w}{b'} \quad (2.33)$$

Where,

- k' = hydraulic conductivity of the river material;
- L_c = the length of a river within a cell,
- w = the width of the river
- b' = the thickness of the riverbed.

Equation 2.33 is used in MODFLOW version 5.3, in the RIVER package (Pattle Delamore Partners Ltd; 2000) where a constant head is assumed in the river for each stress period in each river cell, (i.e. no river flow or change in water level).

Since most of these parameters are often not known, C_{riv} is usually adjusted during the model calibration. As for the other add-on packages, fluxes between surface and groundwater can be organized into tables. Due to the fact that the MODFLOW RIVER package uses Darcy's law instead of the Richards equation to describe the vertical exchange, it cannot accurately model vertical seepage through the unsaturated zone (relative permeability-saturation relationship) as is necessary for perched rivers (Dennis and Witthueser, 2007).

2.3.8 DAFLOW-MODFLOW

The DAFLOW (i.e. Diffusion Analogy Surface-Water Flow model) employs a one dimensional diffusive wave approximation for in-channel flow that is an add-on package to MODFLOW and allows for flow routing, and contains an iterative time stepping approach for coupling the surface and subsurface interactions. The model links surface and subsurface domains using a hydraulic gradient driven flux and assume a saturated subsurface domain (Valerio, 2008). This model simulates one-dimensional flow through a system of interconnected channels subdivided into branches and sub-reaches (which might extend beyond the groundwater model domain, (Figure 17) by solving the diffusive-wave form of the flow equations. The accuracy of the solution increases with the slope and is sensitive to the chosen temporal discretisation (time-step size) in relation to the streambed slope (Dennis and Witthueser, 2007).

Leakage Q_l from a sub-reach into a single specified aquifer cell requires careful discretisation, and is calculated for each surface water time step as:

$$Q_1 = \frac{k'lw(H_a - Y - B_e)}{b'} \quad (2.34)$$

Where,

- k' = the hydraulic conductivity of the streambed,
- l = length of the sub-reach in hydraulic connection with the aquifer cell,
- w = the average width of the stream in the cell,
- H_a = the head of the aquifer in the cell,
- Y = the average depth of the stream in the sub-reach,

B_e = the average elevation of the streambed

b' = the thickness of the streambed.

Jobson & Harbaugh (1999) recommend the model application, especially for upland streams with a unique relation between river stage and discharge, and without flow reversal or backwater conditions.

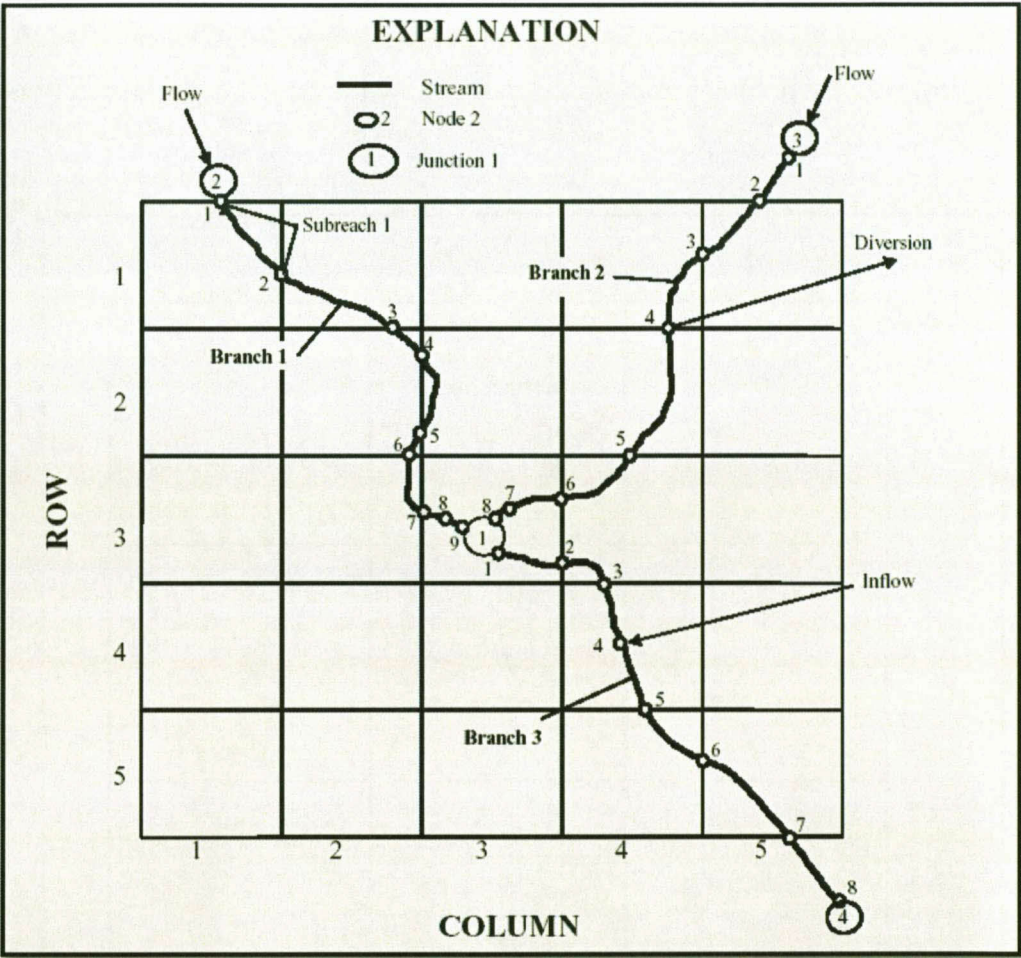


Figure 17: DAFLOW stream network expanding beyond a finite difference model grid (Johnson & Harbough, 1999; in Dennis & Witthuiser, 2007)

2.3.9 MODBRANCH

BRANCH simulates unsteady, non-uniform flow in open channels using an implicit, weighted four point finite difference approximation for the dynamic wave equations. It is referred to as MODBRANCH when incorporated into MODFLOW (Valerio, 2008). MODBRANCH couples the BRANCH module and MODFLOW, and can simulate unsteady flow in a network

(dendritic or looped) of single open-channel reaches or branches, (Figure 18). River flow is solved by the one-dimensional continuity equation; in- and outflow rates are equal at all times and no water is added or removed from storage (Dennis and Witthuiser, 2007). Groundwater interactions are modeled as one-dimensional vertical leakage through a clogging/confining layer in the riverbed. Beyond the data requirements to solve groundwater flow using MODFLOW, the BRANCH module needs the channel geometry, initial flow conditions at all cross-sections, as well as boundary conditions at channel edges. Stationary flow values between surface and groundwater can be directly entered into tables. As for the RIVER package, leakage is modeled for saturated conditions.

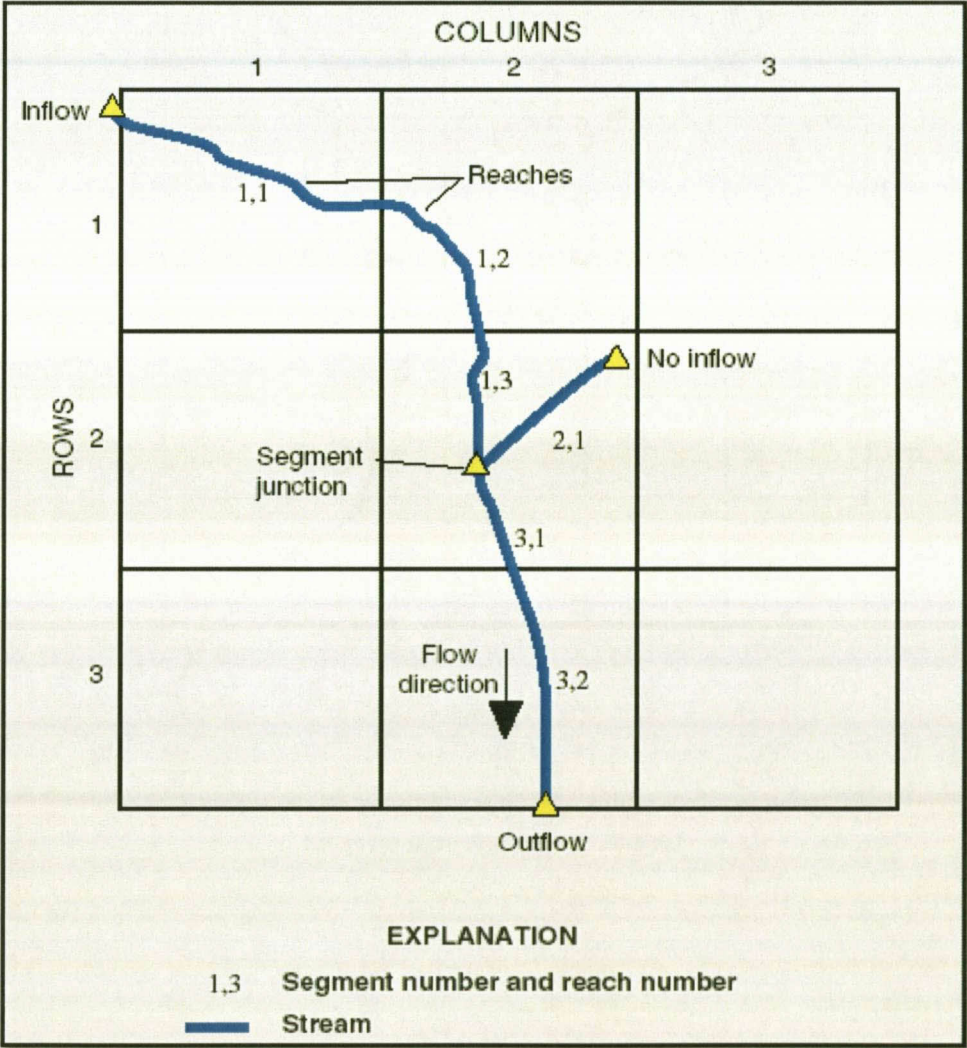


Figure 18: A stream network in a finite-difference model grid (Prudic *et al.*, 2004; in Dennis and Witthuiser, 2007)

2.3.10 MODFLOW-SFR1 (Streamflow routing)

The STREAMFLOW ROUTING package SFR1, coupled with MODFLOW 2000, replaces the earlier STREAM package, which initially overcame the limitations of a constant head in the river by using instantaneous flow routing. The SFR1 approach uses one channel cross section for a stream segment that is divided into three parts on the basis of eight paired horizontal and vertical locations (Figure 19). Eight horizontal distances relative to the left edge of the cross section (viewed in downstream direction) and the corresponding eight vertical altitudes relative to the specified top of streambed (lowest point or thalweg in the channel) are used for computing stream depth, top width, and wetted perimeter (Markstrom *et al.*, 2005). Stream depth, width, and wetted perimeter also are dependent on the slope of the streambed, and two roughness coefficients - one for the centre part of the cross section and another for the two outer parts that may represent overbank flow. The ends of the cross section are assumed to have vertical walls. A mixed bisection-secant method is used to compute stream depth in relation to streamflow since the stream area and hydraulic radius for this option result in implicit functions of depth.

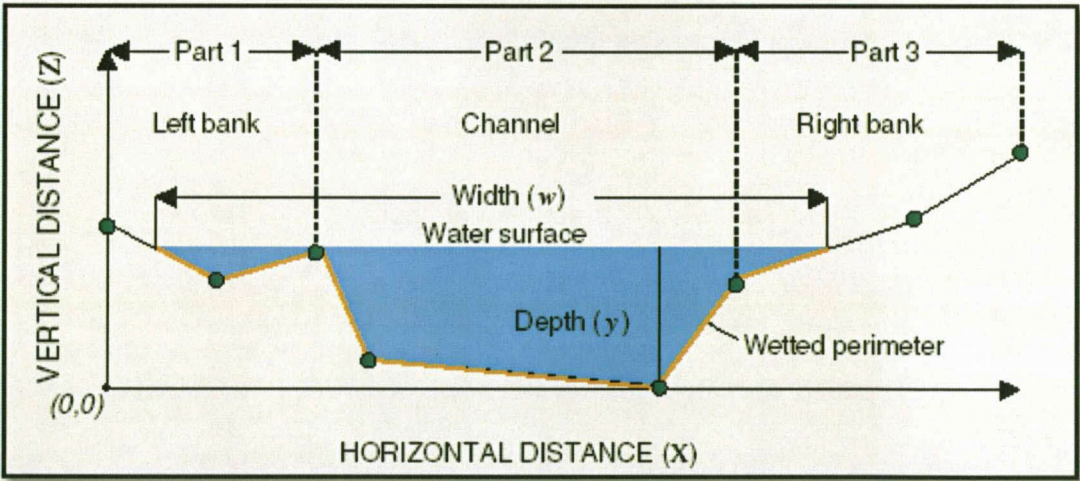


Figure 19: An 8-point cross section for calculation of depth, width and wetted perimeter for a stream segment (Prudic *et al.*, 2004; in Markstrom *et al.*, 2005)

For SFR1, seepage is limited by streambed conductance and the head difference between the stream and aquifer, or when the stream is hydraulically disconnected from the water table, by the head difference between the stream and the bottom of the streambed (Niswonger and Prudic, 2005).

2.3.11 MODFLOW-SFR2

The SFR2 package is the modification of SFR1 since the latter cannot simulate the unsaturated flow beneath streams (Niswonger and Prudic, 2005). SFR2 includes all capabilities of SFR1 and also, unlike SFR1; seepage loss from streams may be restricted by the hydraulic conductivity of the unsaturated zone. Surface water - groundwater interaction is often simulated as a continuum where time delay for water flow in the unsaturated zone is negligible, yet in cases where depth to groundwater is very large, consideration of storage and flow in the unsaturated zone may be important, especially for arid regions such as South Africa. SFR2 caters for this by simulating flow in the unsaturated zone.

Vertical seepage through a homogeneous unsaturated zone is approximated with kinematic waves and that approximation is made by simplifying Richards' equation:

$$\frac{\partial \theta}{\partial t} = \frac{\partial q}{\partial z} = \frac{\partial}{\partial z} \left[D(\theta) \frac{\partial \theta}{\partial z} - K(\theta) \right] \quad (2.35)$$

Where,

- θ = the volumetric water content;
- z = the elevation in the vertical direction;
- $D(\theta)$ = the hydraulic diffusivity;
- $K(\theta)$ = the unsaturated hydraulic conductivity;
- t = time.

Equation 2.35 is simplified such that vertical flux (q) is only driven by gravitational forces. Vertical flux is given by:

$$q = -K(\theta) \quad (2.36)$$

The characteristic solution for the kinematic-wave equation for the unsaturated flow can easily be determined by substituting 2.36 into 2.37 and then expressing θ partially in terms of t and z :

$$\frac{\partial z}{\partial t} = \frac{\partial K(\theta)}{\partial \theta} = v(\theta) \quad (2.36)$$

Where,

$v(\theta)$ = the characteristic velocity,

The characteristic velocity, which has positive and negative characteristics in the case of open channel flow, is restricted to the downward direction for the movement of a wetting front because the force of gravity can only cause the wetting front to move downward (Niswonger and Prudic, 2005).

In gravity-dominated flow, the flux in the unsaturated zone is equal to the unsaturated vertical hydraulic conductivity and cannot increase above the value of the saturated hydraulic conductivity. Consequently, seepage across the streambed in SFR2 is limited by the underlying vertical hydraulic conductivity in the unsaturated zone. The volume of water that seeps from a stream is calculated by multiplying the infiltration rate by the wetted area of the stream. The wetted area of the stream may be held constant or determined on the basis of the stream cross-sectional dimensions, discharge, and stage. The relation between stage and discharge is calculated using Manning's equation. The option to simulate unsaturated flow between streams and aquifers may be used for a stream of constant wetted area and width or for an eight-point cross-section and leakage is modelled for each of the compartments (Figure 20).

The capability to model unsaturated flow is seen as a key feature of modeling perched rivers more realistically.

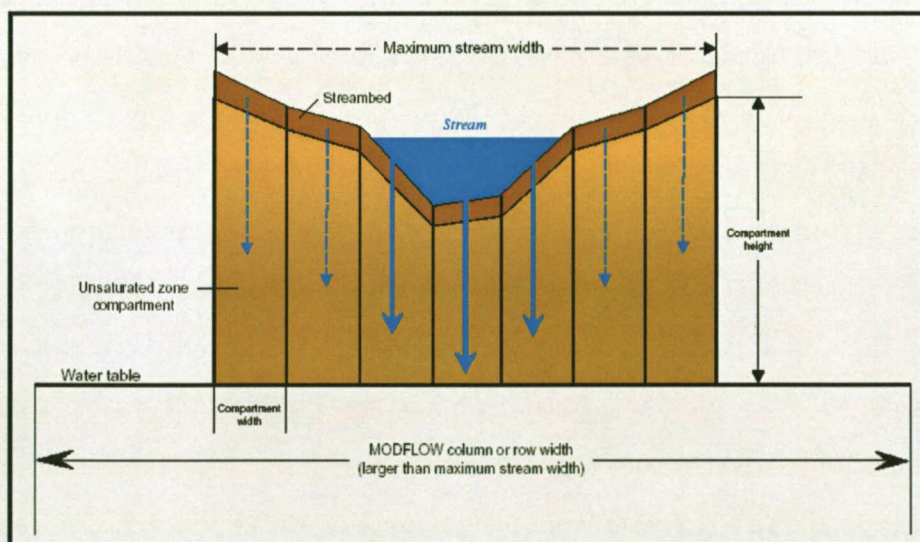


Figure 20: The discretization of the unsaturated zone under a stream of variable cross section in a single MODFLOW cell

2.3.12 MIKE SHE and SHETRAN

MIKE SHE is an integrated hydrological modelling system for simulating surface water flow and groundwater flow, while SHETRAN is also a physically based distributed model but one that provides three dimensional coupling of water flow and contaminant transport. The model assumes an unconfined aquifer underlain by an aquiclude (impermeable layer). The catchment is discretized by a horizontal orthogonal grid and vertical columns of horizontal layers representing different hydrological compartments.

The hydrological processes are modeled by finite-difference solutions of the equations of mass, energy and momentum conservation, or by empirical equations. Although physically based, the discretization requires the estimation of effective hydrological and geohydrological parameters for the grid. The surface water-groundwater interaction is a function of the head difference between the river and the aquifer and can account for clogging layers on the streambed as well as disconnected streams.

2.3.13 FEFLOW and MIKE 11

FEFLOW, a three-dimensional finite element in which saturated and unsaturated flow, and groundwater modeling software developed by WASY are coupled via an interface manager (only on Windows platform) with the MIKE11 surface water software developed by the Danish Hydraulic Institute (Dennis and Witthuiser, 2007). MIKE11 models one-dimensional unsteady flow in surface water (branched or looped) networks and quasi two-dimensional flow on floodplains with a finite-difference scheme. Several structures in the river such as weirs, bridges and pumps, as well as user-defined structures, can be incorporated. A non-iterative coupling with FEFLOW is achieved by exporting discharges (positive or negative) calculated by FEFLOW at coupled boundary nodes (3rd type or Cauchy boundary conditions) to MIKE11 H-points, as an additional baseflow boundary condition (Figure 21). Once MIKE11 has calculated its time steps to coincide with the FEFLOW time steps, the water levels of the H-points are exported to the FEFLOW coupling boundary nodes, and FEFLOW calculates its next time step.

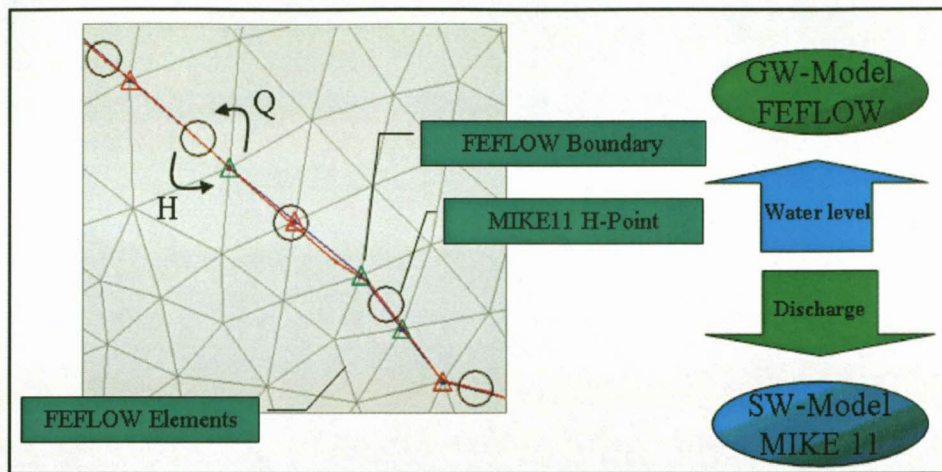


Figure 21: Coupling of FEFLOW and MIKE11 (WASY, 2005 in Dennis and Witthuiser, 2007)

MIKE11 requires information for each branch on the cross-section, initial (water level) and boundary conditions (water levels, inflows or stage-discharge relation) as well as the bed (surface water flow) resistance (Manning or Chezy equation). If values are provided for the Chezy equation, MIKE11 uses global values, except for the up- and downstream end of a branch (WASY, 1995).

Manning's equation is based on Chezy's equation calculated as:

$$C = k[R^{(0.125)}/n] \quad (2.38)$$

Where,

C = Chezy's roughness coefficient;

R = Hydraulic radius;

n = Manning's roughness

k = Constant.

Substituting the calculated roughness into Chezy's equation yields Manning's equation:

$$Q = \left[\frac{k}{n} \right] AR^{2/3} S^{1/2} \quad (2.39)$$

Where,

Q	=	Discharge;
k	=	Constant;
n	=	Manning's roughness coefficient;
A	=	Flow area;
R	=	Hydraulic Radius;
S	=	Slope Friction

The coupled software system is the most comprehensive modelling approach to surface-groundwater interaction so far. However, the data requirements as well as software skills are quite comprehensive if the software is to be used to its full capability (Dennis and Witthueser, 2007).

2.4 Comparisons of numerical models

The intended purpose and the conceptual geohydrological model inform the type of numerical model that would be suitable to use under a particular situation. It is also important to note that the adequacy of the conceptual model determines the performance of the model. However, data requirements and availability as well as knowledge and skill requirements and availability are also important to consider in numerical modelling. Numerical models are based on basic assumptions and conditions under which modelling simulation occurs.

Given the nature and typical hydrologic conditions of the surface water and groundwater systems in South Africa which are characterised by semi-arid and perched groundwater aquifers, and available resources any of the following coupled models are recommended for use in quantification of surface water - groundwater interaction.

- MODFLOW-SFR2: MODFLOW is widely used Worldwide including South Africa where aquifer systems are characterise by flow in the unsaturated zone can be handled by SFR2. It can model unsaturated flows between aquifers and streams (especially perched rivers).

- FEFLOW & MIKE 11: Though data intensive, this coupled model may also be applicable to South African conditions
- SHETRAN: This model can also account for clogging layers on the streambed.

Table 2A and Table 2B representing the comparisons of various numerical model can also be used for identifying suitable models as these tables link model assumptions to particular attributes of aquifer and river systems.

2.5 Comparisons of groundwater surface water interaction
evaluation methods with respect to costs and data requirements

Table 3 depicts various assessment methods with respect to associated costs and data requirements. It is also indicated whether the particular method is quantitative (i.e. quantify or estimate amount of seepage flux across interface) or qualitative (in other words only indicates whether there is interaction regard less of how much water flows across the interface between surface water and groundwater).

Table 2: Comparisons of numerical methods

Models → Attributes ↓		Finite difference methods									
	FEFLOW (coupled to MIKE-11)	MODFLOW							MIKE-SHE	SHETRAN	GSFLOW
		Main program	Modules or subroutines								
			River	DafLOW	Branch	SFR1	SFR2	Stream			
Simulation of flow in the unsaturated zone (use Richards)	x						x		x	x	X
Model solute transport		X (GWT)								x	X
Account for lateral or interflow & model preferential flow											X
Instantaneous leakage between stream & aquifer		X				X		x			
Use manning’s equation to compute river stage	x					X	x	x			x
Use Darcy to model vertical river0aquifer exchanges		x	x			X	x	x			x
Applied upstream to unidirectional flows				x							
Stream-aquifer exchange & route through surface flow						x					

Models  Attributes 	Finite element methods	Finite difference methods									
	FEFLOW - (coupled to MIKE-11)	MODFLOW							MIKE-SHE	SHETRAN	GSFLOW
		Main programme	Modules or subroutines								
			River	Daflow	Branch	SFR1	SFR2	Stream			
Stream-aquifer exchange & route through surface flow						x					
Simulate unsteady flow in a network of single open channel branches					x						
One dimensional vertical leakage					x						
Assume constant head			x					x			
Can model perched rivers							x		x	x	
Account for clogging layer					x				x	x	
Steady, uniform & constant density streamflow (i.e. no water added or removed)						x					

Table 3: Comparisons of various methods for assessment of groundwater surface water interaction with respect to data requirements and associated costs

Type of Method	Short description of the method	Comparison of methods		Additional comments
		In terms of data requirements	In terms of costs	
Hydrograph separation / analysis (quantitative method)	Stream flow analysis to estimate baseflow component (i.e. groundwater discharge). Separate quick flow from interflow and baseflow.	Stream flow data often collected regularly – available in most cases	Low	Tools available for use (e.g. excel spreadsheet). Provides info through time and commonly used but method assumes gaining stream only. However, method is subjective
Field observation of seepage flux (qualitative method)	Visual indication of springs and tell-tale mineral precipitates.	Requires understanding of field indicators (as substitute to data)	Low	Method can help identify seepage hotspots and can be complimented by field mapping.
Seepage flux measurements (quantitative)	Seepage meters are used to measure flux at the interface between groundwater and surface water	Several observation points are required for spatial spread. Hence method is time consuming	Medium	Meters are simple, easy to construct and inexpensive. However, meters may not be appropriate for high flows
Analytical models (quantitative)	Assumption based analytical solutions where stream depletion rate is estimated	Can be computationally challenging; moderate in data requirements	Medium	Conceptual model is a prerequisite, and relevant hydrological parameters are used to estimate stream depletion rate
Numerical models (quantitative)	Numerical simulation of flow using mathematical equations	Data intensive & data is not always available	High & can be highly complex	Conceptual model is a prerequisite
Environmental tracers & chemistry techniques (generally qualitative)	Use chemical composition and isotopic signature of water	Good results can still be attained with limited data	Low-medium analyses may be expensive	Relatively robust and less data intensive. Can give information on source or evolution of water

3 Case studies

3.1 Background

The following case studies are used to demonstrate and test surface water - groundwater interaction methodologies.

- The Richmond Farm area case study builds on previous work done wherein groundwater assessment for water supply was undertaken. That project also included estimation of baseflow to the Klein Dwers River which prompted the choice of this area for research.
- The Weatherley Catchment case study is based on hillslope hydrological research that has been going on for a number years in that area. Data and information available have potential to assist in understanding flow mechanisms including possible interaction between surface water and groundwater.
- The work previously done in the Seekoei River catchment including focus on the interaction between groundwater and the pools along the riparian zone encouraged inclusion of this area as a case study.
- In the Mokolo River catchment the previous work done by Department of Water Affairs and other researchers as part of the determination of the reserve including focus on the interaction between the alluvial aquifer and the river made this area an obvious choice among the list of candidate areas for the research.
- The Institute of Groundwater Studies initiated a Water Research Commission project aimed at enhancing the understanding of the interaction between surface water and groundwater in the Krugersdrift Catchment. That also made this area a good choice for a case study on the interaction between the river and the alluvial aquifer

3.2 Case Study: Richmond farm area

3.2.1 Location

This case study is based on feasibility assessment studies on potential water supply for the emine, previously undertaken by various investigators (EGS and GMS, 2003; ERM, 2004; van Tonder and Dennis, 2005; and Kotze *et al.*, 2006), from boreholes in the Richmond farm within the Klein Dwers River catchment. Richmond boreholes are located on the farm

known as Richmond, upstream of the confluence of Groot and Klein Dwars River, about 30 km south southwest of Steelpoort town within the Eastern Bushveld complex in the Mpumalanga Province (Figure 22). The other borehole field that was investigated for water supply is Helena to the south west of Richmond.

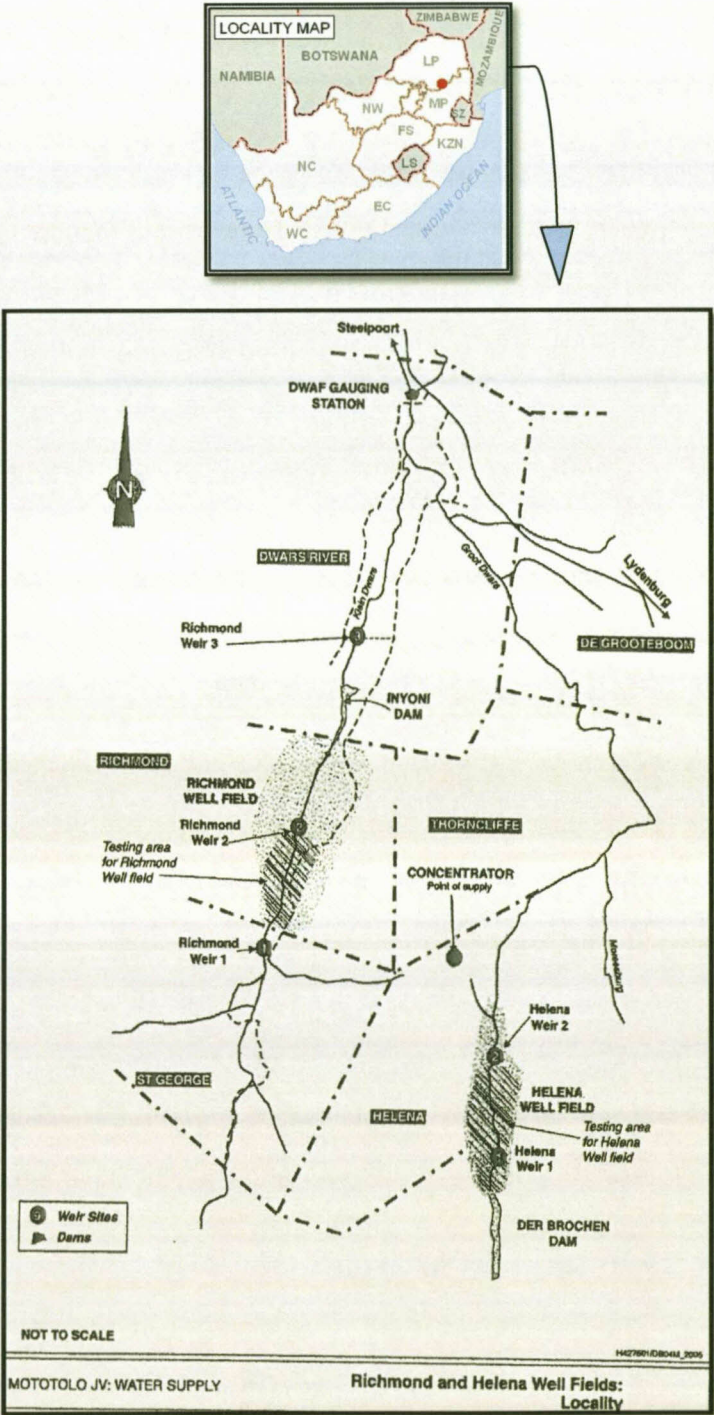


Figure 22: Locality map of the Richmond farm area (not to scale) borehole field sites and rivers (adapted from BKS, 2005)

3.2.2 Major rivers and topography

The perennial Klein Dwars River flows northwards from its source along a steep sided;

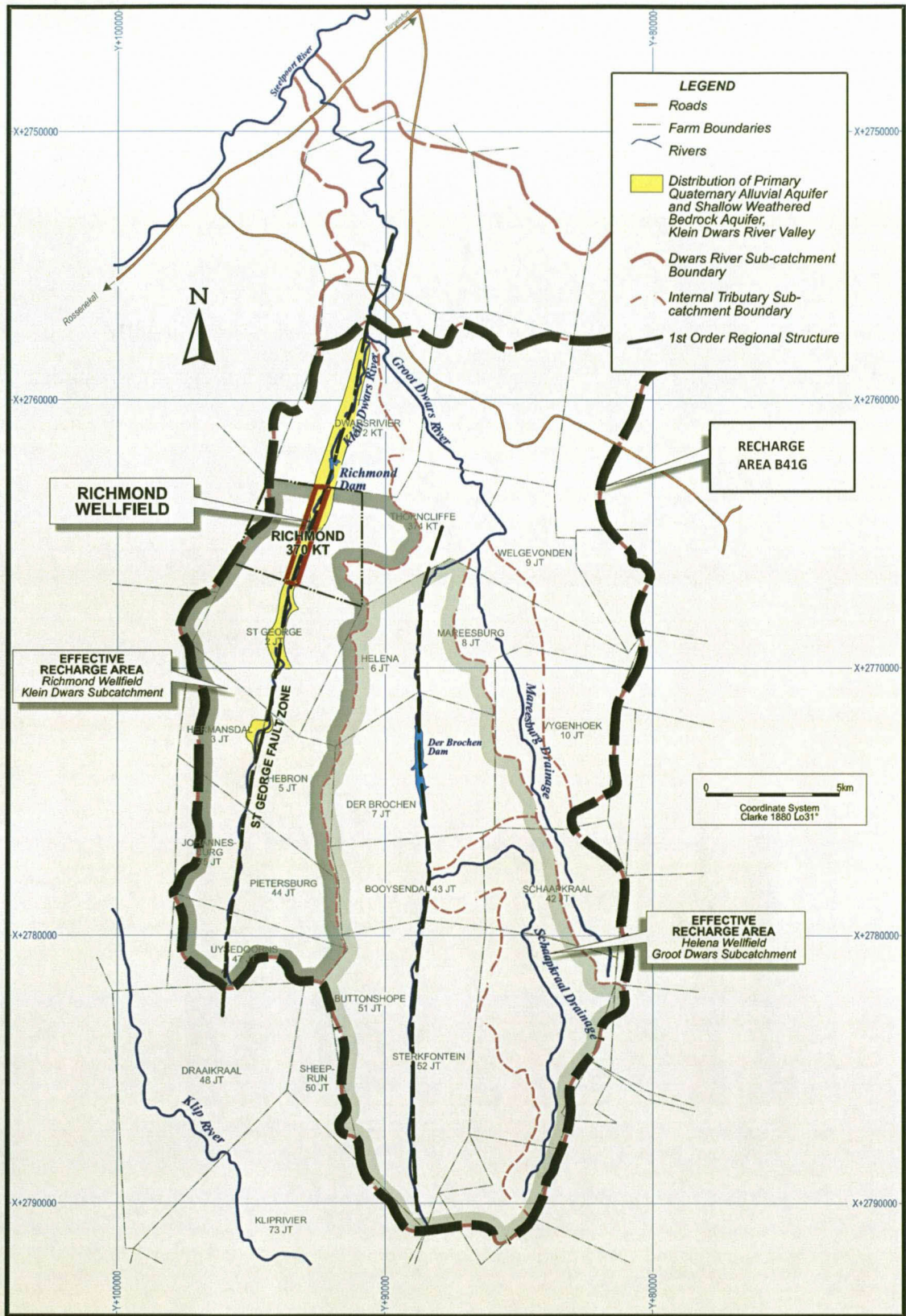


Figure 23: Richmond map showing recharge areas, rivers and borehole positions (Kotze et al, 2006)

deeply eroded valley in mountainous terrain, to the northern boundary of Richmond Farm for 16 km. Beyond Richmond Farm the valley opens into flatter undulating country where the Klein Dwars River joins the Groot Dwars River, a total length of 25 km (Kotze *et al.*, 2006). The elevation at the head of the Klein Dwars catchment is 1812 mamsl and drops to 900 mamsl at the confluence with the Groot Dwars River (Figure 23). The average width of the catchment is about 6 km.

3.2.3 Climate

The catchment lies within the summer rainfall area with rainfall predominantly occurring between October and March with a maximum in January. The mean annual rainfall is approximately 540 mm, while evaporation is 1 061 mm/year. Figure 24 depicts the average rainfall and evaporation in Richmond over a period of 70 years. Annexure 1 shows typical rainfall during November through January in the Richmond area.

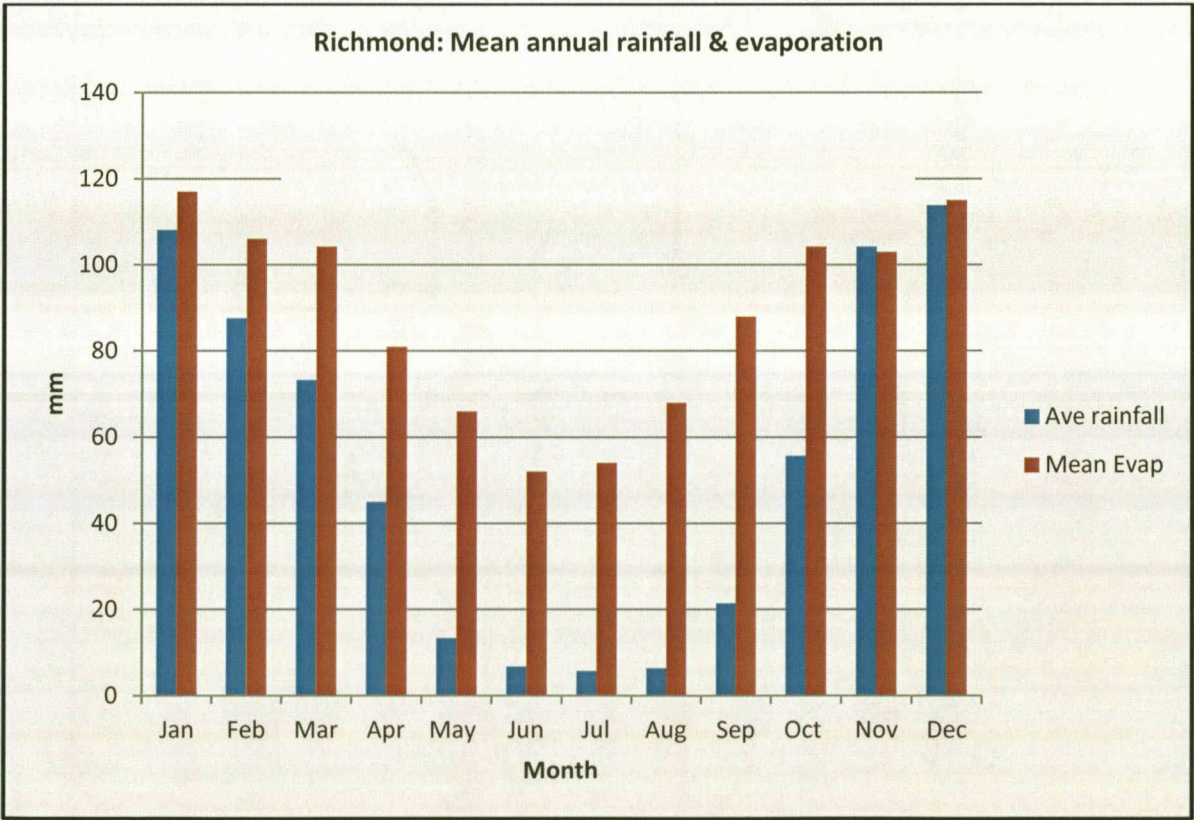


Figure 24: Average rainfall and evaporation in Richmond from 1924 to 1994 (data from BKS, 2004)

3.2.4 Geology and structure

The Klein Dwars catchment is underlain by mafic and ultramafic igneous rocks of the Bushveld Complex. The Richmond Farm is situated on norites with subordinate anorthosite and pyroxenite layers of the Critical Zone in the northern and eastern portions of the farm as shown on Figure 25. These are overlain by norites, gabbro-norites and leuconorites of the Main Zone in the southern and western part of the farm.

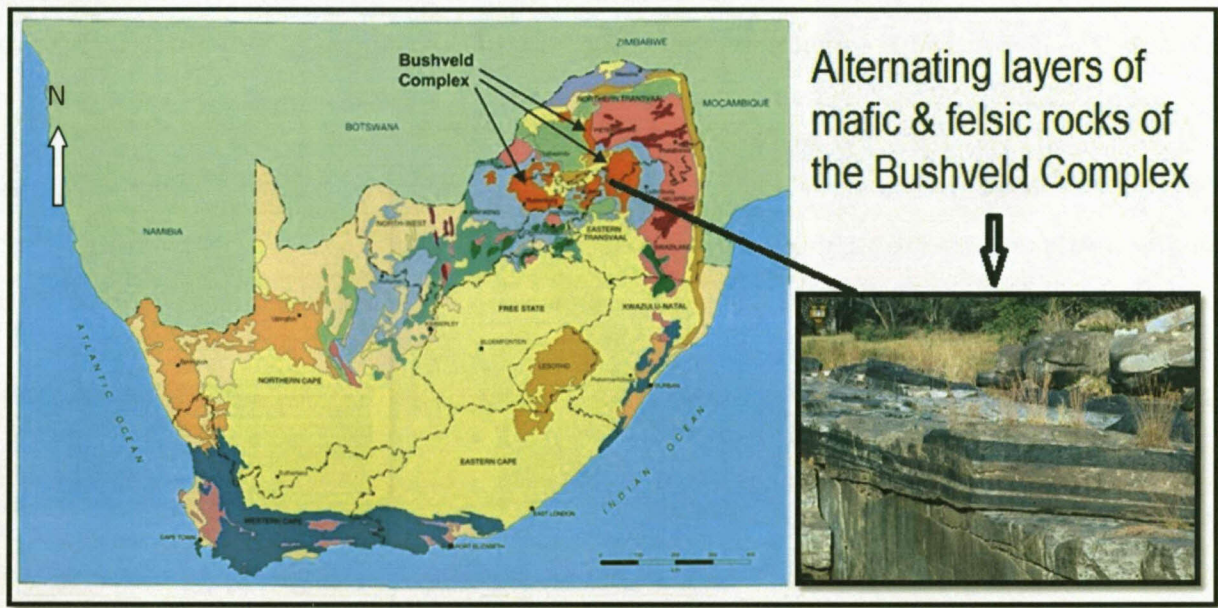


Figure 25: Alternating layers of mafic (dark grey – mainly pyroxenite) rocks and felsic (lighter bands of norite & anorthosite); picture of rocks adapted from pictures by David Waters, Oxford University

3.2.5 Methodology and approach to groundwater exploration and development

Previous investigations had postulated that efficient groundwater abstraction could be achieved along the valley bottom of Richmond/St George Farms from a shallow weathered bedrock aquifer system at a depth of 20 - 60 m underlying the primary alluvial aquifer of the Klein Dwars River (Kotze et al, 2006). However, in the follow-on investigation, the subsequent researchers developed and adopted a comprehensive approach to determine potential groundwater supply at a higher level of assurance. Their methodology entailed conducting long term pump testing to estimate hydraulic properties as well as:

- Rainfall recharge using the chloride method

- Recharge from other catchments to the aquifer
- Baseflow to the river using hydrograph separation methods
- Groundwater abstraction from the aquifer
- Surface water - groundwater interaction between Klein Dwars river and the aquifer,
- Consumption and evapotranspiration by the riparian vegetation

This approach enabled the development of a conceptual site model for setting up a numerical model for assessment of the groundwater flow dynamics. One of the findings was that the shallow weathered aquifer and the overlying alluvial aquifer were interconnected.

The Klein Dwars River valley is underlain by a regional north north-east trending lineament which is believed to be a first order sub-vertical fault zone (St George Fault). A series of thin dolerite dykes intrudes along the Klein Dwars valley parallel to the fault. A series of local east-west, south-east and south south-east trending fractures, characterised by zones up to 20 m wide of closely spaced joints (5-50 cm joint spacing), occurs throughout the farm.

The depth of bedrock weathering progressively increases from near surface in the mid- and upper slopes of the valley side (coinciding with bedrock outcrop) to greater than 40 m underlying the Klein Dwars River channel as indicated in Figure 26 and Figure 27.

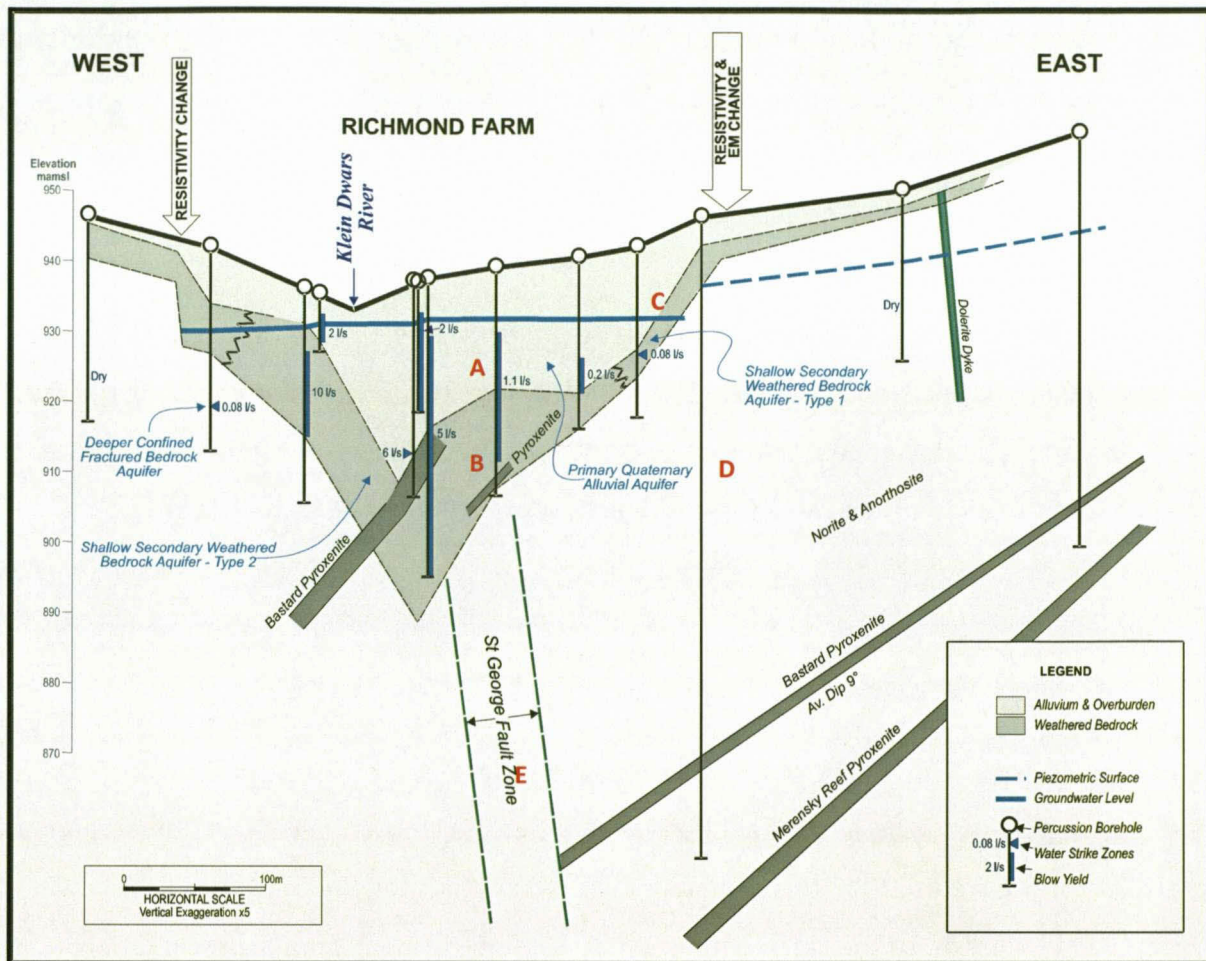


Figure 26: A cross section showing geology and conceptual site model of Richmond (Kotze *et al.*, 2006)

3.2.6 Conceptual site model and the aquifer system

The cross section in Figure 26 depicts the conceptual site model describing the aquifer dynamics of the Richmond pilot boreholes. It also represents an east-west cross-section across the Klein Dwaars River Valley based on geophysical resistivity and electromagnetic (EM) profiles and direct drilling data. The following aquifer systems were identified by Kotze *et al.*, (2006):

- **A:** Alluvial aquifer represents up to 28 m thick of which 24 m is saturated unconsolidated alluvial sand, silt and clay sediments. It underlies the active channel of the Klein Dwaars River in a zone that is approximately 250 m wide.
- **B:** Type 2 shallow weathered bedrock aquifer, underlying the primary alluvial aquifer, where weathering extends to depths greater than 42 m.
- **A and B:** Shallow, leaky unconfined primary alluvial aquifer in direct hydraulic connection with Type 2 shallow weathered bedrock aquifer. These aquifers are continuously

developed along the valley bottom of the Klein Dwars River on Dwars Rivier, Richmond and the northern part of St George Farms.

- **C**: Shallow semi-confined secondary weathered bedrock aquifer (Type 1) is developed along the Klein Dwars River valley in the weathered bedrock zone of the lower and mid-slopes of the valley sides.
- **D**: Deeper confined fractured bedrock aquifer, a low yielding aquifer present in the mid- and upper valley slopes of the river valley extending to the peripheral watershed boundary. It is developed below the normal weathered zone in tightly fractured and jointed Bushveld norites and anorthosite bedrock (and locally dolerite dykes).
- **E**: Regional steep structural aquifer, postulated to occur along the St George Fault, which underlies the shallow primary and secondary bedrock aquifers in the central part of the Klein Dwars River valley.

3.2.7 Recharge, long-term pumping test and estimation of the sustainable yield

Recharge estimate to the alluvial aquifer was relatively higher at 15% of the mean annual precipitation than the 3% of the mean annual precipitation for the bedrock aquifer, while the average recharge for the catchment is about 3.2% of the mean annual precipitation.

A total of 4 boreholes were pumped over a period of 4.5 months and 20 boreholes observed during the pump testing. Pump test results are summarised on Annexure 2. Figure 27 shows the structural geology, location of pumping and observation boreholes, and two weirs that were installed upstream and downstream of the river. In addition to monitoring change in drawdown with time, rainfall and stream flow were recorded continuously during the controlled pumping test to determine the sustainable yield of the borehole field.

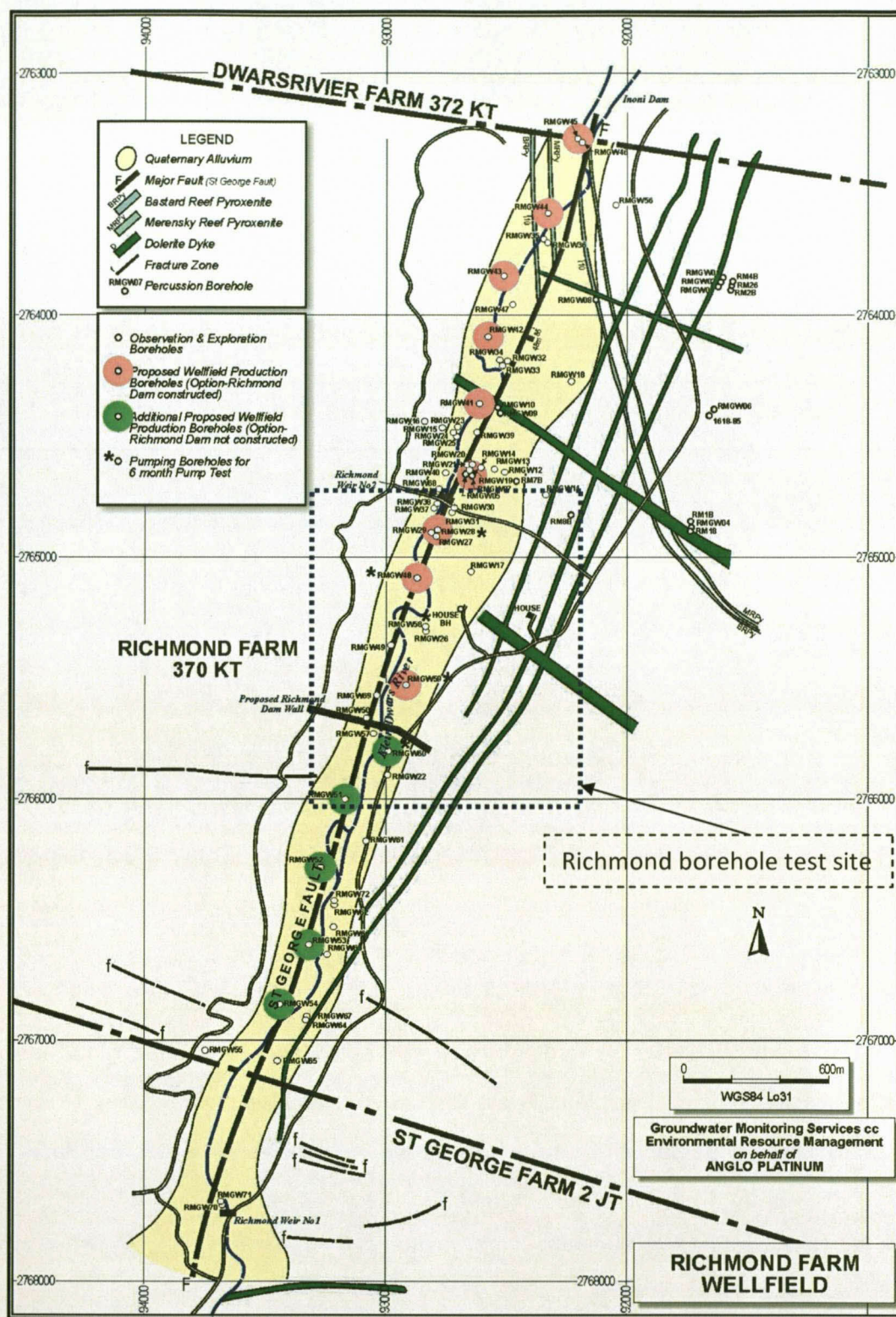


Figure 27: Location of Richmond farm pumping and monitoring boreholes and weirs (Kotze *et al.*, 2006)

The Flow Characteristics (FC) method was used to analyse the long-term pumping test data, and Table 4 represents the pumping test results.

Table 4: Richmond pumping test results (source data – van Tonder and Dennis, 2005)

Borehole No.	Abstraction rate (l/s)	T-late (l/s)	Min. Sustainable yield (l/s)	Max. Sustainable yield (l/s)
RMGW28	2.50	28	2.50	3.0
RMGW48	4.00	25	1.50	2.2
RMGW58	1.00	8	0.58	2.3
RMGW59	4.00	15	0.86	1.5

3.2.8 Baseflow estimation based on flow measurements

The winter baseflow rate of the Klein Dwars River was estimated to be 60 l/s in August 2002, while the baseflow rate of the Groot Dwars River measured with a square notch at the same time was 85.2 l/s (Kotze *et al.*, 2006).

3.2.8.1 Flow measurements, relative groundwater and surface water level, and environmental isotope hydrology and water sample chemistry

Kotze *et al.*, (2006) took surface water flow measurements at weirs 1 located 3 km upstream of weir 2 as well as rainfall, and found that the abstraction had no influence on flow and no increase in flow at weir 2 even after pumping had stopped as indicated in Figure 28. The environmental isotope values (Oxygen-18) of pumped water was found to be -3.2‰ while that of the river was relatively more enriched at +0.5‰, thus indicating that there was neither loss of water from the Klein Dwars river to the aquifer nor recharge to the river downstream (i.e. no perceptible interaction between the river and the aquifer) in the vicinity of the pumped borehole field. However, the investigators inferred that interflow seepages from the weathered shallow aquifer mainly occur upstream, near weir 1.

Possibly there was no interaction in this part of the investigation area. Exchange between surface water and groundwater usually does not occur throughout the entire river reach.

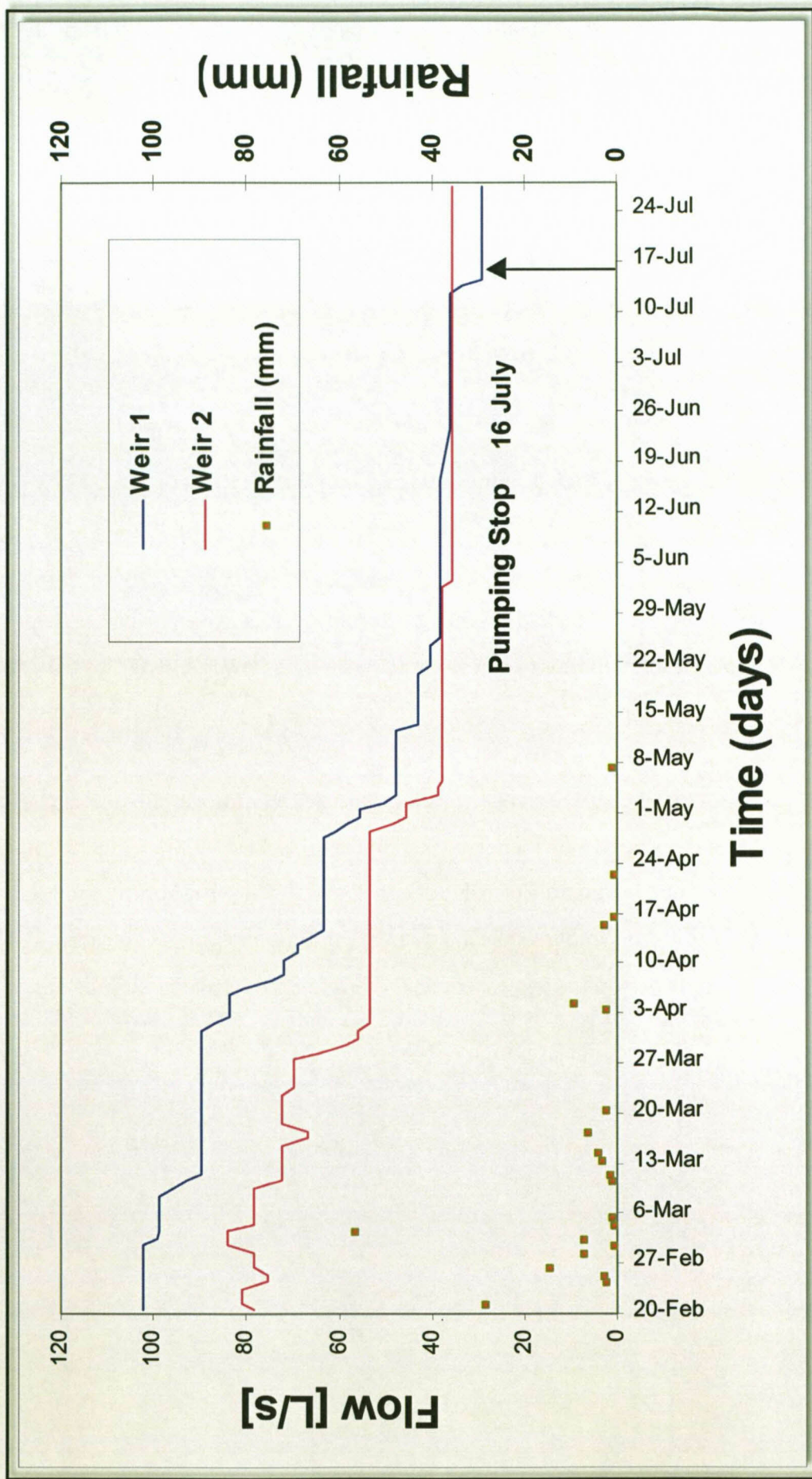


Figure 28: Flow measurements at weirs 1 and 2, as well as rainfall (Kotze *et al.*, 2006)

3.2.8.1.1 Interconnectivity between the alluvial aquifer and weathered bedrock aquifer
To conclude that the alluvial aquifer and the weathered bedrock aquifer are hydraulically connected, ERM (2004) encountered similarity in water level in the two boreholes that were each drilled in these different aquifer systems. Kotze *et al.*, (2006) also concluded from their work done in the area that the two aquifer systems are connected. In particular, Figure 29 shows that borehole RMGW05 that was drilled and open only in the alluvial aquifer's water level was compared and found to be similar to the water level in borehole RMGW07 that was drilled and open only in the weathered bedrock aquifer.

3.2.8.1.2 Shallow weathered bedrock aquifer and the Klein Dwars River

Within the Type 1 (i.e. zone C in Figure 26) shallow weathered bedrock aquifer is expected to contribute to baseflow of the Klein Dwars River as groundwater flow is most likely to follow the topographic gradient towards the valley bottom. However, given a relatively low transmissivity of bedrock aquifer and barrier by dolerite dykes, the flow could be sluggish.

3.2.8.1.3 Relative water level stage between the river and adjacent boreholes

Figure 30 shows that the groundwater elevation in boreholes RMGW30 (933.375) and RMGW31 (933.773) is higher in relation to the surface water level elevation at the Klein Dwars River Bridge (933.05). This implies that the river has potential to gain water from the alluvial aquifer. However, that does not guarantee the interaction between surface water and groundwater.

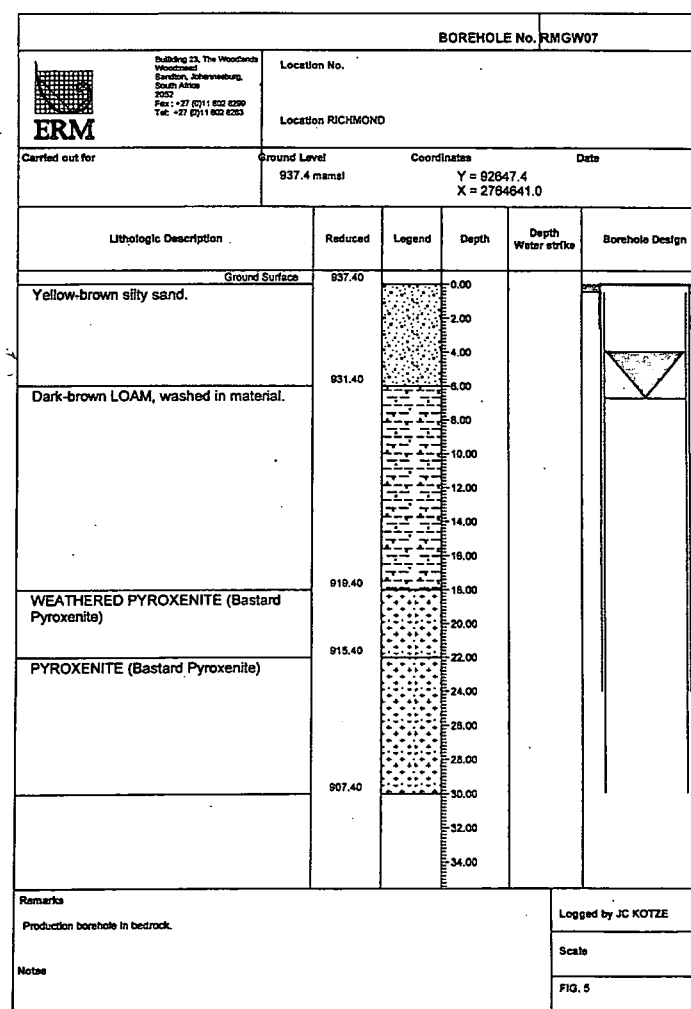
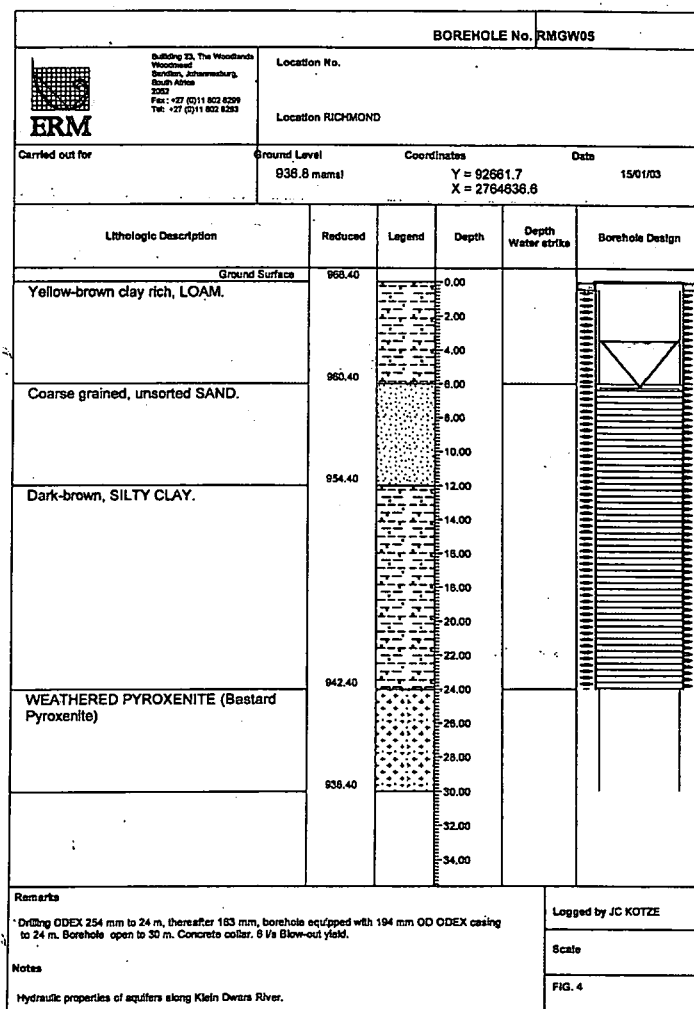


Figure 29: Richmond - comparison of water levels between borehole RMGW05 and RMGW07 (ERM, 2004)

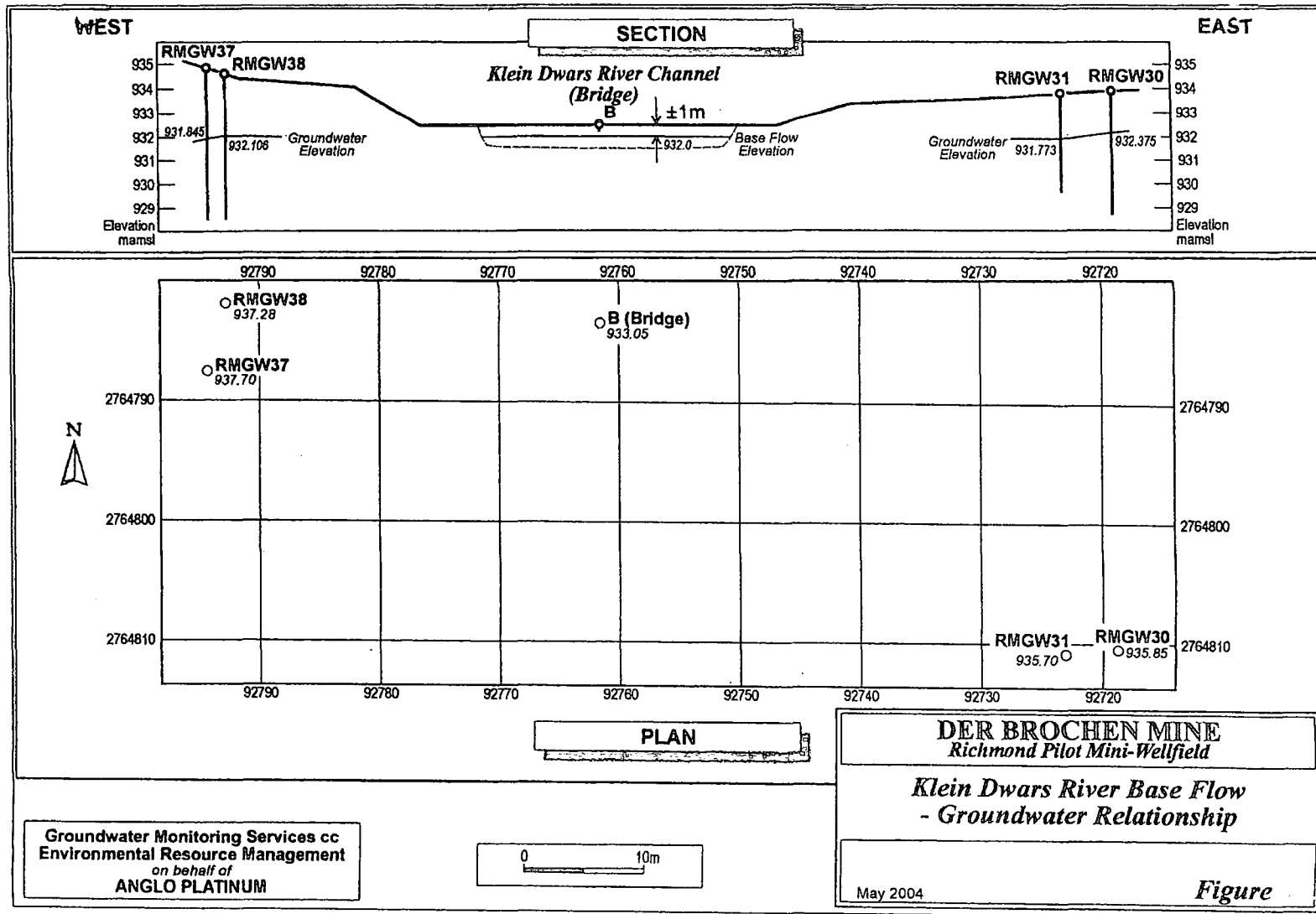


Figure 30: Comparison of groundwater levels in the alluvial and underlying weathered aquifer (ERM, 2004)

3.2.8.1.4 Pump testing of RMGW19 and environmental isotope data including for other boreholes

The RMGW19 borehole was pump tested over a period of 7 days and Figure 31 shows the drawdown recorded where water is drawn from the alluvial aquifer.

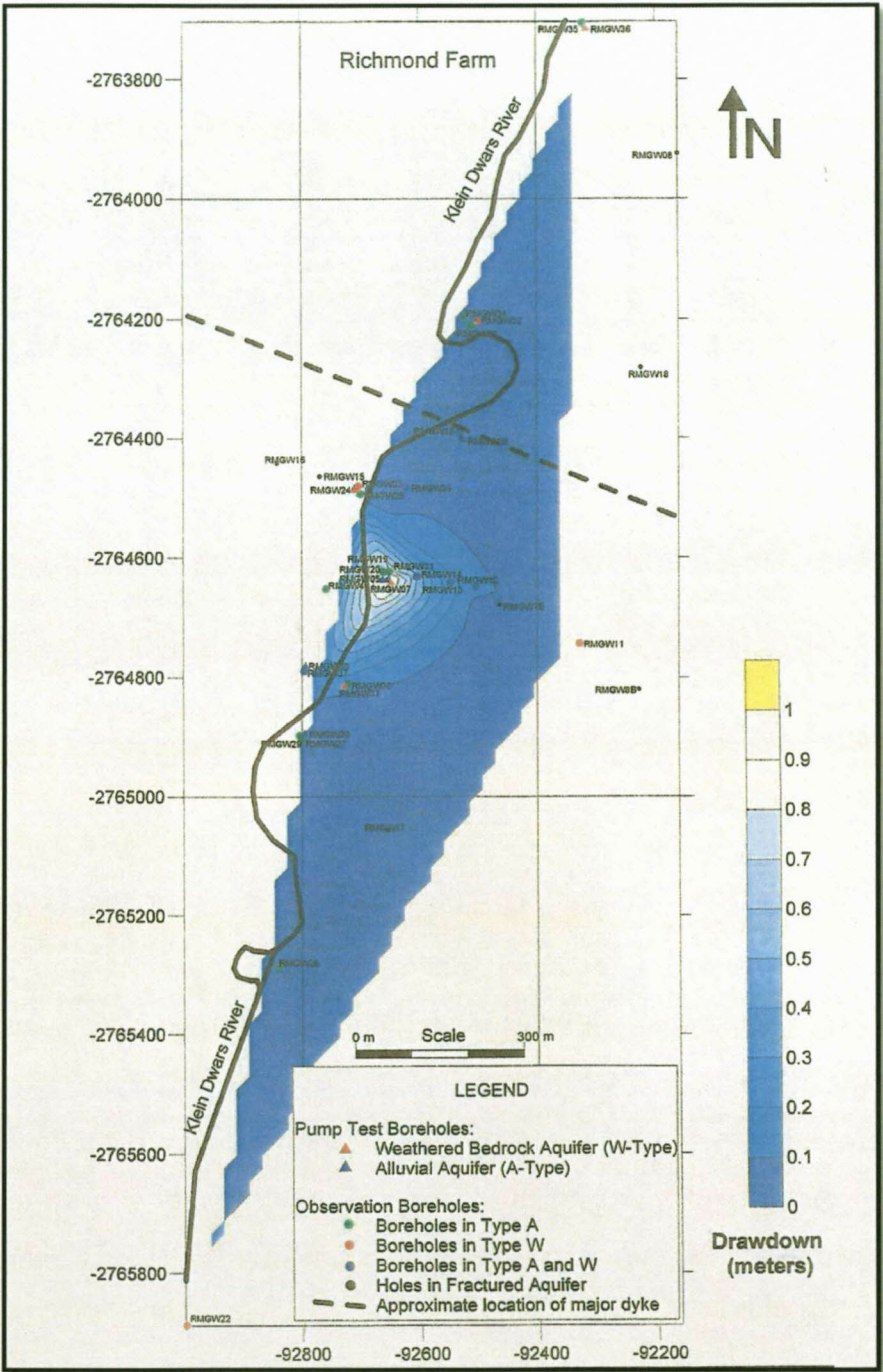


Figure 31: Drawdown record during pump testing of RMGW19 (ERM, 2004)

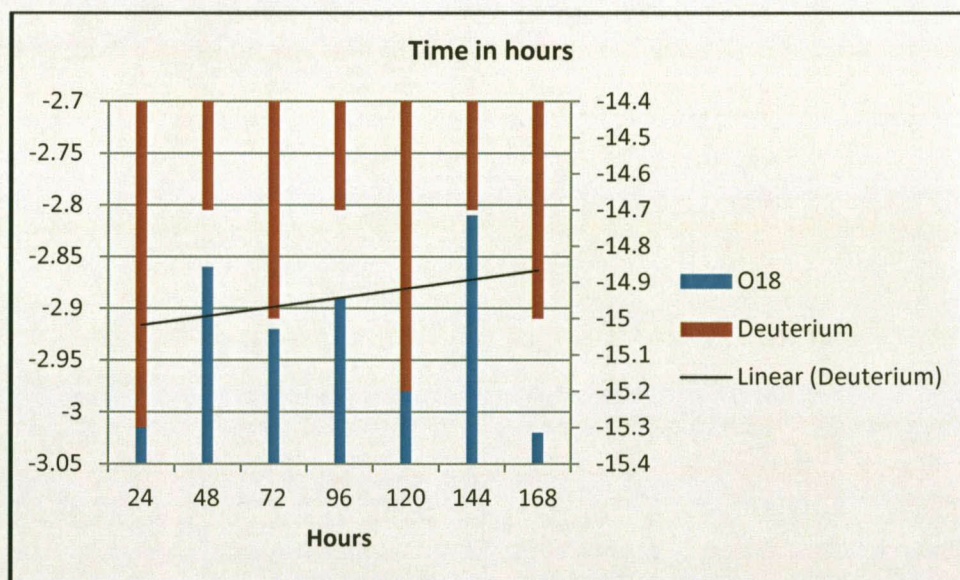
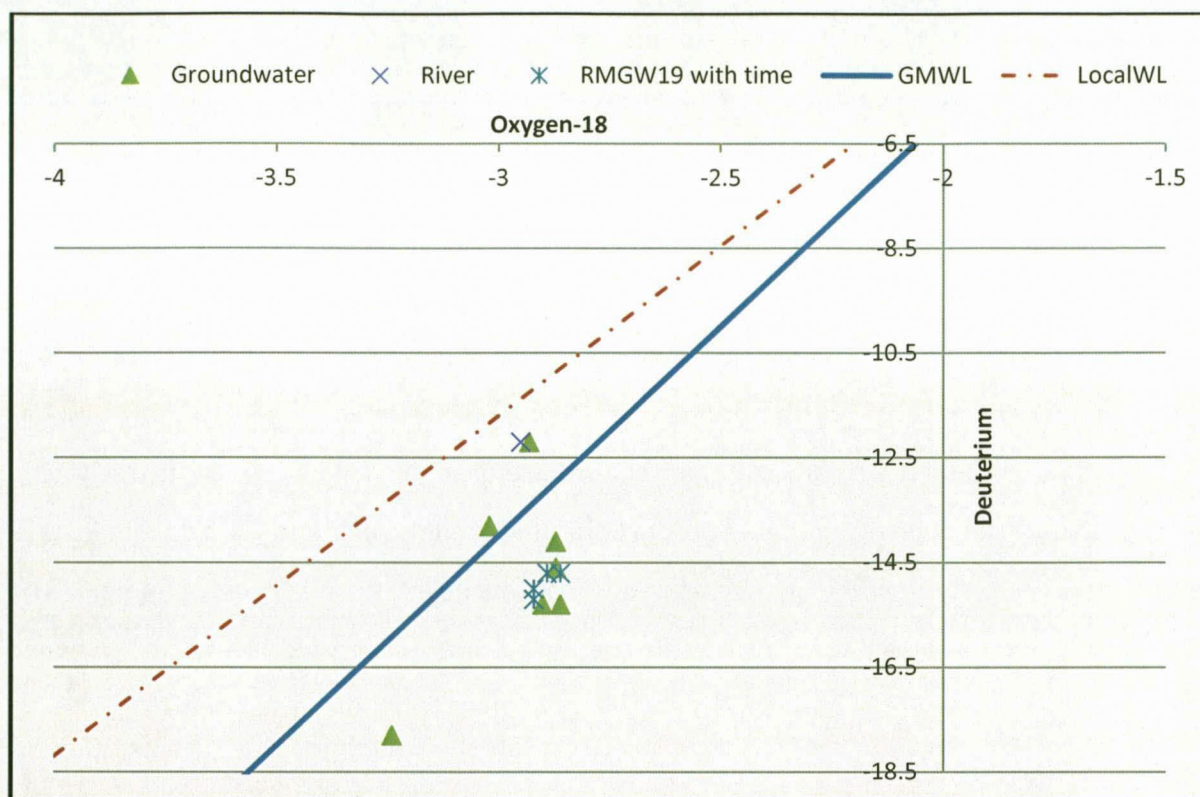
The isotopes data was also captured from observation boreholes in the vicinity of the pumping borehole with the data collected from the latter (i.e. pumping borehole RMGW19) over time as shown on Table 5.

Table 5: Environmental isotope data for Richmond (ERM, 2004)

Borehole	Time (hr)	δD (‰)	$\delta^{18}O$ (‰)
RMGW05	1	-14.1	-2.87
RMGW05	72	-13.5	-2.97
RMGW07	1	-13.8	-3.02
RMGW07	72	-13.9	-3.02
RMGW08	2	-12.2	-2.93
RMGW19	24	-15.3	-2.9
RMGW19	48	-14.7	-2.86
RMGW19	72	-15	-2.92
RMGW19	96	-14.7	-2.89
RMGW19	120	-15.2	-2.92
RMGW19	144	-14.7	-2.81
RMGW19	168	-15	-3.02
RMGW24		-14.6	-2.87
RMGW32		-15.3	-2.86
RMGW36		-17.8	-3.24
Dwars River		-12.2	-2.95

The isotopic composition of water, determined by mass spectrometry, is expressed in per mil (‰) deviations from the Standard Mean Ocean Water (SMOW). These deviations are expressed as δ^2H or δD for the deuterium (heavier isotope of hydrogen), and $\delta^{18}O$ for a heavy isotope of oxygen. Water with less deuterium than SMOW has a negative δD ; water with more deuterium than SMOW has a positive δD .

Figure 32 represents δ^2H and δD isotopic signature which shows a slight depletion on the river water sample probably due to high evaporation of river water prior to recharging the weathered bedrock aquifer. Data for all boreholes plot nearer the World Meteoric Line thus reinforcing the possibility of mixing of alluvial aquifer and weathered bedrock aquifer water. The RMGW19 water sample indicates fluctuation in isotopic signature with pumping in which case deuterium gets enriched with time, probably due to ingress of or mixing with river water(Figure 33).



3.2.8.1.5 The water chemistry and groundwater surface water exchange

The Piper diagram is quite useful in graphically showing the type or nature of a given water sample, as well as the inter-relationships to other samples. The Piper diagram can be used to classify and identify geologic units with chemically similar water, and define the evolution in water chemistry along the flow path.

Clearly the groundwater from sampled boreholes and the Klein Dwars River water (the latter sample was taken from the bridge near Weir 2) have similar chemistry (Ca-HCO_3) as shown on Figure 34. The bicarbonate dominance is typical of shallow or recently recharged water, while the seemingly equitable proportions of Mg and Ca elements could be indicative of the source rocks (i.e. the pyroxenite and anorthosite of the Bushveld complex). The Dwars river sample also plots in the same area which may suggest some water exchange between the river and the aquifer. Thus chemistry confirms possible interaction.

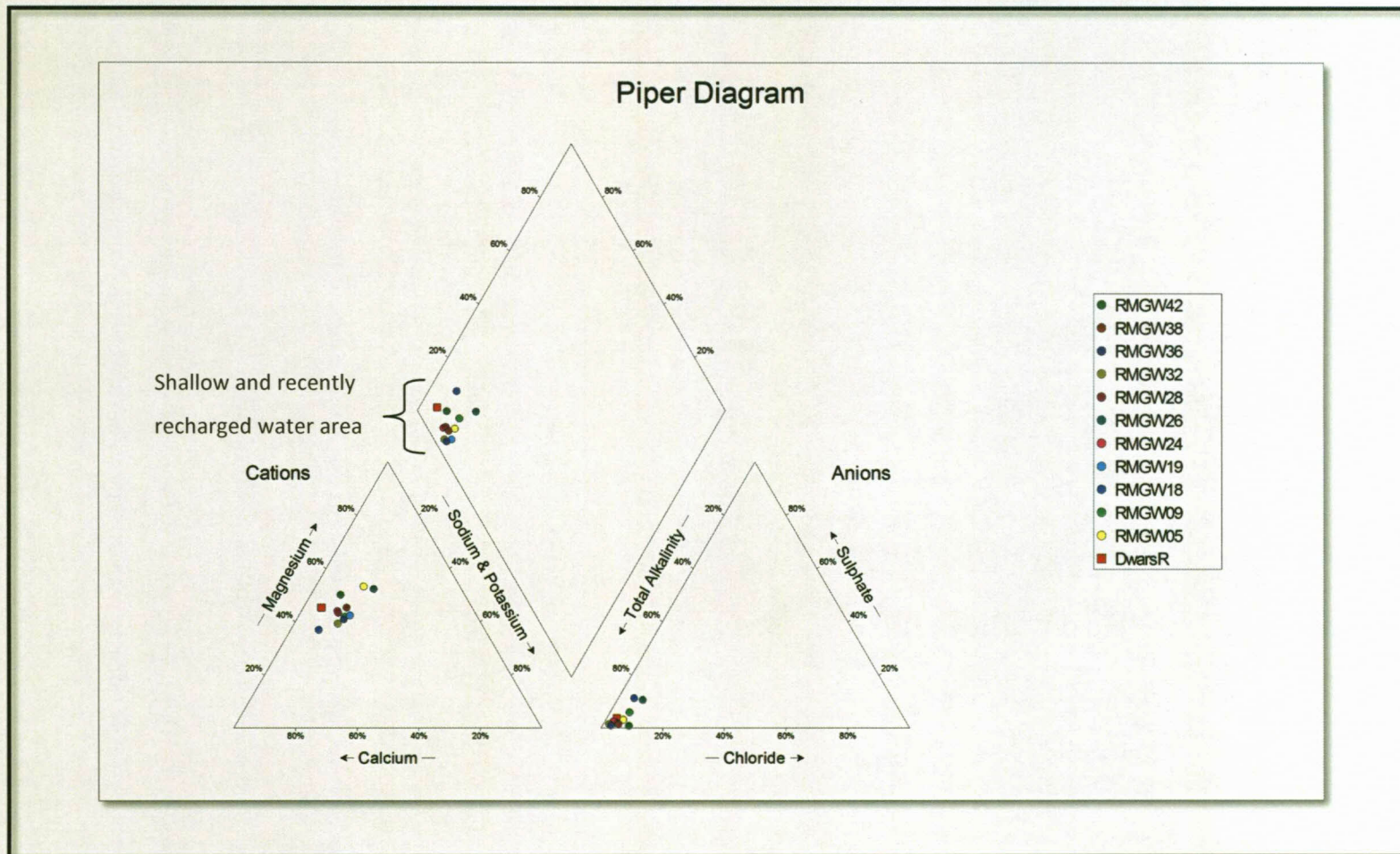


Figure 34: Richmond - groundwater and Klein Dwars River water chemistry plots in the recently recharged region. (Source data – ERM and GMS, 2004)

3.2.8.2 Analytical methods of estimating groundwater surface water exchange

Lack or unavailability of adequate and relevant data poses challenges with regard to use or application of the numerical methods. However, some data are available to at least test one of the appropriate analytical methods to estimate the exchange. Assuming constant transmissivity of a homogeneous, isotropic aquifer of an infinite extent with dominant lateral flow overlain by a thin stream with impervious layer the Stang-Hunt method is applicable for this exercise (equation 2.12, 2.17 & Table 1):

$$\frac{\Delta Q}{Q_{abs}} = \operatorname{erfc} \left[\sqrt{\frac{Sd^2}{4Tt}} \right] - \exp \left[\frac{\lambda^2 t}{4ST} + \frac{\lambda d}{2T} \right] \operatorname{erfc} \left[\sqrt{\frac{\lambda^2 t}{4ST}} + \sqrt{\frac{Sd^2}{4Tt}} \right]$$

Borehole RMGW28 that was pump tested for 72 hours is used in this test (data are as follows):-

Q (pumping rate) = 12 l/s (535.68 m³/d)

D (distance to river) = 200 m

S (specific yield) = 0.08

T (Transmissivity) = 80 m²/d

T (time pumping) = 3 days (for 72 hours)

k' (permeability) = 0.01

b' (thickness) = 0.5 m

w (stream width) = 250 m

L (stream leakance is calculated from equation 2.17)

A programme developed by Dennis (2011) is used for this purpose (Figure 35). The river gains water at the rate of 15.8 m³/d; with the net gain despite pumping from borehole RMGW28 for 72 hours at 2.5 l/s within the cone of influence. Figure 36 indicates the depletion rate with time if pumping was to continue on up to 120 days (4 months). By 120th day the river would then be losing water to the aquifer at the rate of 22 m³/d.

Input			
Q	[m ³ /d]	535.68	Abstraction
d	[m]	200	Shortest distance between well and stream
S		0.08	Storage coefficient or specific yield
T	[m ² /d]	80	Transmissivity
t	[days]	3	Time
k'		0.01	Permeability of semi-pervious streambed material
b'	[m]	0.5	Thickness of semi-pervious streambed
w	[m]	200	Stream width
L	[m ² /d]	4320.0	Stream leakance
Lamda		0.0	Constant of proportionality
Calcs			
A		0.058925565	
B		0.000160751	
C		0.04629628	
Output			
DeltaQ	[m ³ /d]	-15.788	Stream depletion rate
DeltaQ	[l/s]	-0.183	Stream depletion rate

Figure 35: Stang-Hunt method (programme developed by Dennis, 2011)

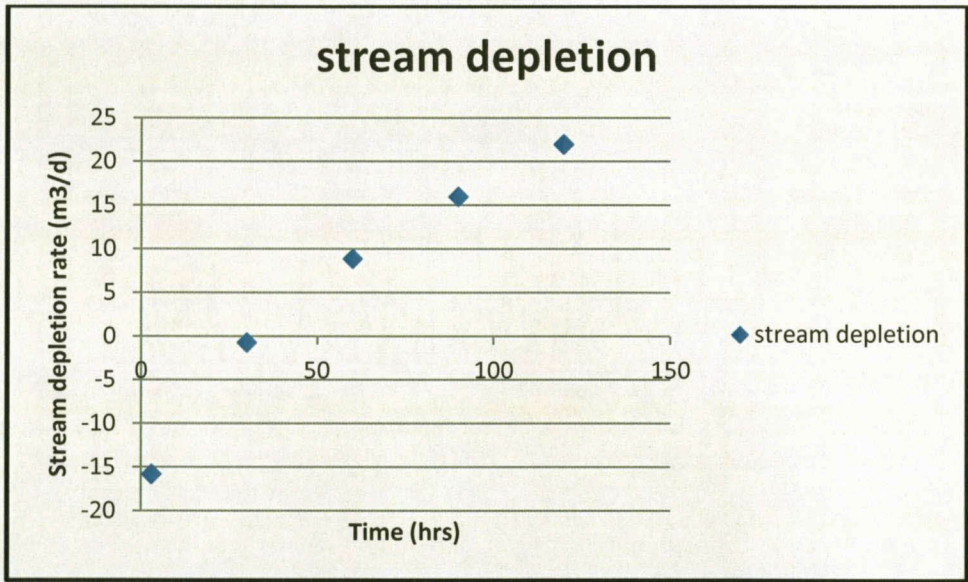


Figure 36: Richmond: Estimated stream depletion rate using Stang-Hunt analytical method

In the foregoing sections (3.2.8.1 & 3.2.8.2) both water chemistry and environmental isotope data analyses indicate that there has at least been some exchange between groundwater and surface water, while the analytical method, though seems to contradict the other methods could be used to quantitatively estimate the exchange. Relative water levels between groundwater and surface water on the other hand indicate the probability of whether the river would gain or lose in the case of interaction between the 2 systems.

3.2.8.3 Estimation of baseflow using hydrograph separation method for the Klein Dwars river sub-catchment

Kotze *et al.*, (2006) estimated baseflow of the entire Klein Dwars river catchment by hydrograph separation method. The flow data was obtained from the flow gauge B4H009 of the Department of Water Affairs. A straight line was drawn using the minimum monthly flow values to obtain an amount of $4.5 \text{ Mm}^3/\text{a}$ (approximately 142 l/s) accounting for a total baseflow component of quaternary catchment B41G (Figure 37). A statistical percentile analysis carried out on flow data yielded an 80% confidence level for the estimated catchment baseflow of this amount. The baseflow component for the Klein Dwars sub-catchment is calculated as $1.21 \text{ Mm}^3/\text{a}$ (38 l/s), obtained by multiplying the baseflow calculated for the catchment with a factor expressing the proportional sub-catchment size (118 km^2) in relation to the total catchment size (442 km^2).

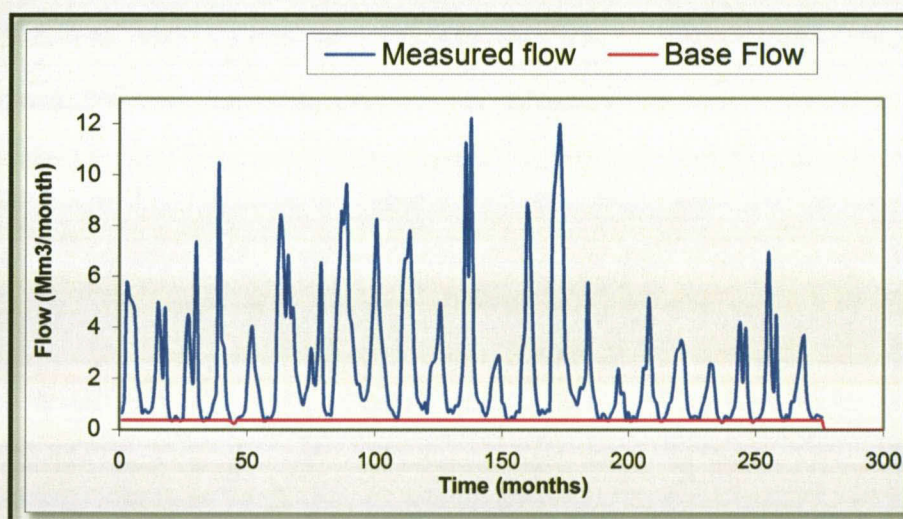


Figure 37: Baseflow estimation using hydrograph separation method (Kotze *et al.*, 2006)

Groundwater Monitoring Services (2001), through numerical simulation or modelling of the catchment obtained a relatively large estimate of $28\,173 \text{ m}^3/\text{d}$ (326 l/s) as net recharge of groundwater to the river.

3.2.9 Water balance in the Richmond borehole field

The flux towards the alluvium aquifer from the hard rock aquifer is estimated at 0.41 Ml/day assuming a groundwater gradient of 0.05 towards the river, while the flux entering the

alluvium aquifer from the upper reaches of the river is 1.1 Ml/day, assuming a gradient in the river of 0.01 (van Tonder and Dennis, 2005). Therefore the total amount of water entering the alluvium aquifer is 1.8 Ml/day. The riparian vegetation in the vicinity of the river could use up to 0.66 Ml/d which is derived by multiplying the riparian area with the potential evaporation rate of 1.7 m/year. The total volume of water stored in the alluvium is approximately 1.53 Mm³ (1 500 Ml). The minimum safe yield of the four abstraction boreholes at Richmond used during the pumping test, is 5.4 l/s along a stretch of about 1 km for a period of 2 years. Therefore the sustainable borehole field yield of the 14 production boreholes in Richmond Borehole field is 1.8 Ml/d, as opposed to the initial estimated exploitation potential of 5.15 Ml/d.

Surely vegetation often intercepts and takes up the lion's share of water that would otherwise, contribute to baseflow.

3.2.10 Lessons learnt from the Richmond case study

- Preliminary investigations that had not taken groundwater and surface water interaction into account, had indicated that much of groundwater supply required could be extracted along the valley bottom of Richmond/St George Farms from a shallow weathered bedrock aquifer system underlying the primary alluvial aquifer of the Klein Dwars River. However, it was later established that the shallow weathered aquifer and alluvial aquifer are interconnected. In fact it became clear to the investigators that overestimation of sustainable yield occurred due to short term pumping test interpretations and water balance that were used without taking interaction between the two aquifers.
- The impact by riparian vegetation needs to be taken into account as well since vegetation evapo-transpires most of the water from the system.
- Understanding of the conceptual site model and water systems dynamics informs appropriate methodological approaches in assessment of water exchange.
- Geology and structure including depth to weathering has a significant influence and control on connectivity and potential interaction between groundwater and surface water.
- Chemistry and environmental isotope data analyses are quite useful in establishing whether there is exchange between groundwater and surface water.

- Either analytical or numerical methods are useful in quantitative estimation of the exchange between groundwater and surface water. However, unavailability of data often limits the use of numerical methods.
- The analytical method can be used to demonstrate that continued pumping of a borehole near a gaining river can reverse the hydraulic gradients thus leading to a losing river.
- In light of the high confidence level (80%) based on statistical analysis of flow data in the estimation of catchment baseflow using hydrograph separation method, it could be concluded that despite its limitation, this approach also has important role in surface water - groundwater interaction.
- Use of a combination of various methods (at least more than one), is useful to confirm whether interaction between groundwater and surface water resources exists, and to enhance the level of confidence in preliminary results.

3.3 Case Study: Weatherley Catchment

3.3.1 Location and of the Weatherley Catchment

The Weatherley research catchment is situated in the Eastern Cape, 4 km south-west of Maclear, South Africa (Lorentz *et al.*, 2004; Lorentz *et al.*, 2008; van Tol *et al.*, 2010; and Le Roux *et al.*, 2010). The catchment covers approximately 160 ha and is one of many small tributaries of the Mooi River (Figure 38). The catchment forms a valley-like structure surrounded by hills in very short distances (Figure 39), hence this study area has received much attention in the past with respect to hill slope hydrology.

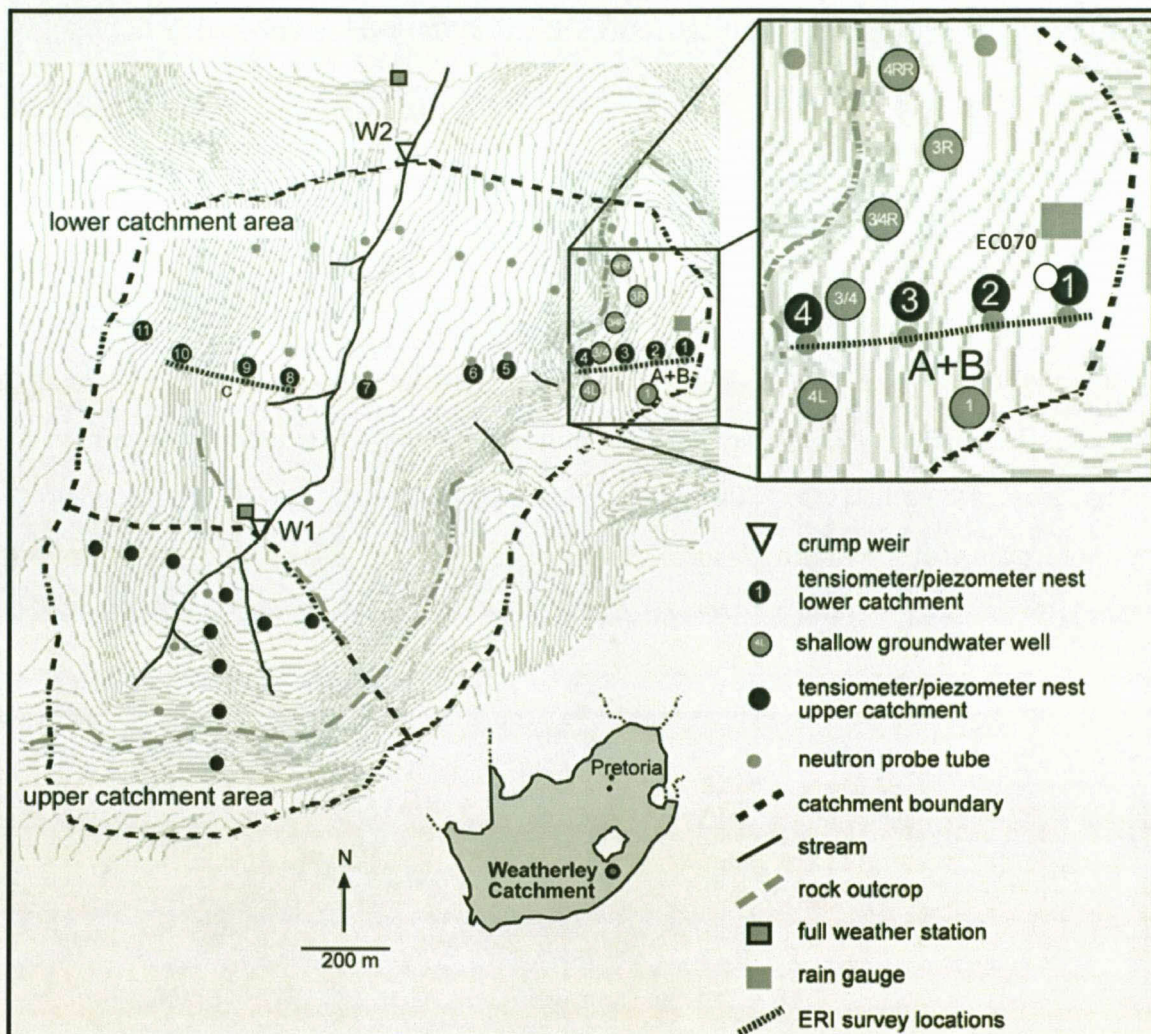


Figure 38: The location of the Weatherley Catchment (after Wenninger *et al.*, 2008; adapted from Lorentz *et al.*, 2004)

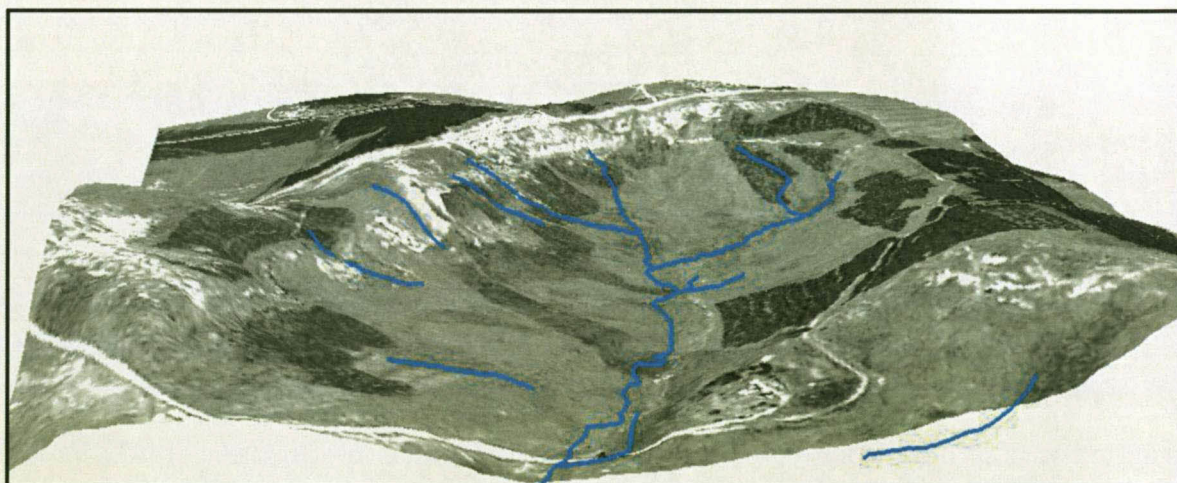


Figure 39: topographical features of the Weatherley catchment viewed from downstream (Bouwer, 2012)

3.3.2 Geology, topography and structure of the Weatherley Catchment

The geology consists of sandstone and mudstone of the Elliot Formation predominantly underlain by the Molteno Formation as shown on Figure 40. The Molteno Formation forms a prominent ‘shelf’ around most of the catchment at an altitude of approximately 1 320 m, which has a dominating influence on the hydrology of the catchment, especially on the eastern and southern sides. Two small dolerite dykes with a north-south strike occur in the catchment. Wetland conditions exist along the entire reach of the stream and range in width from 100 to 400 m, being widest where marshy conditions are associated with lines of seepage from the contributing hillslopes (Le Roux et al, 2010). The highest point in the catchment occurs in the south western corner at 1 352 m above mean sea level, and the stream exits the catchment at 1 254 m above mean sea level.

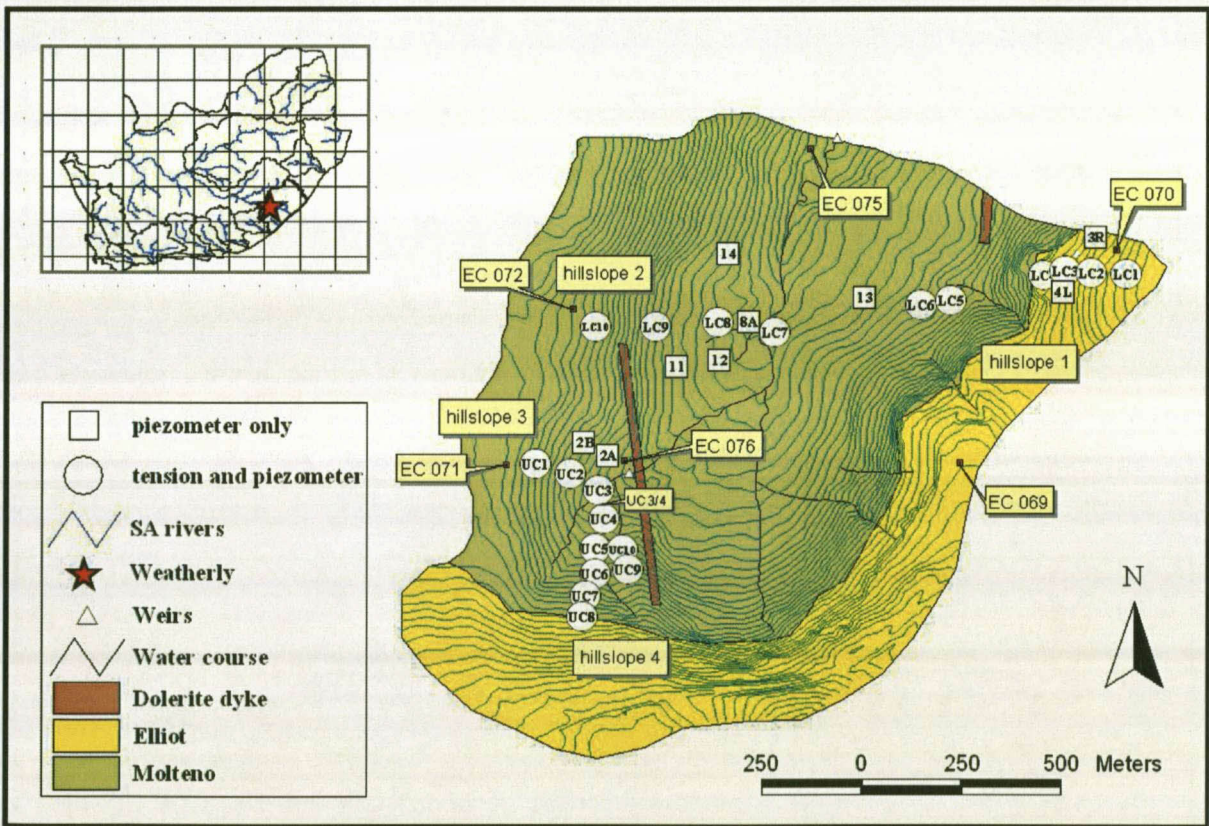


Figure 40: The geology of the Weatherley Catchment (Freeze et al, 2011)

3.3.3 Climate

Based on rainfall data from 1998 to 2007, the mean annual precipitation (MAP) for the Weatherley catchment is about 1 016 mm (Figure 41). The precipitation occurs in summer from October - March with May - September receiving least amount of rainfall (Figure 42).

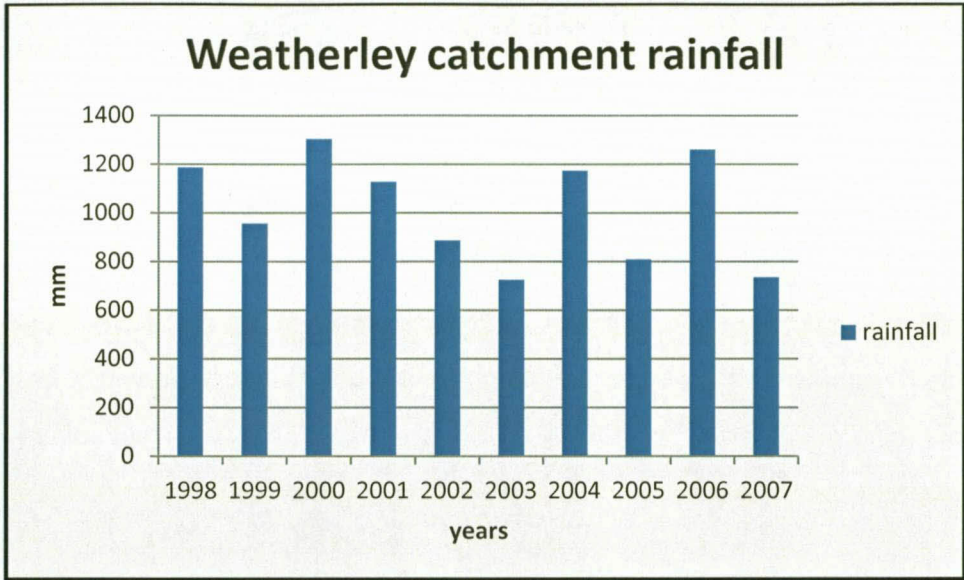


Figure 41: The annual rainfall in the Weatherley catchment from 1998 to 2007. (Source data - Lorentz, 2011)

The mean annual potential evaporation (MAE) is 1 488 mm per year (van Tol *et al.*, 2010 and Le Roux *et al.*, 2010). The winters are cold, with mean minimum temperatures of 4°C. Frost and snowfall is common, particularly in the higher-lying areas during the winter, while the summers are warm with a mean maximum temperature of 25°C.

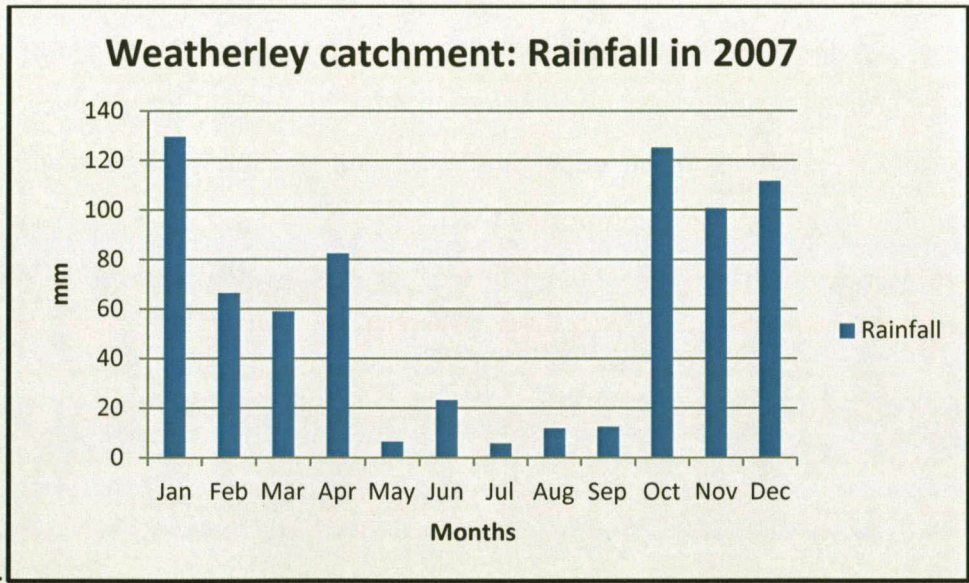


Figure 42: Weatherley catchment: monthly rainfall in 2007. (Source data - Lorentz, 2011)

3.3.4 Land cover, soils and wetlands

The natural land cover consists of Highland Sourveld grasslands with a basal cover of 50-75% on the hillslopes (van Tol *et al.*, 2010). *Eucalyptus nitens*, *Pinus elliottii* and *Pinus patula* trees were planted in selected areas during 2002. The different soil types identified display a varying degree of wetness and colour including red and yellow apedal mesotrophic soils as well as neocutanic and hydromorphoc soils.

Wetland conditions exist throughout the catchment along the stream, with a width of 100 to 400 m. The widest areas of this wetland are associated with seepage lines from contributing hillslopes.

3.3.5 Observation network and measurement of various parameters

A variety of experimental techniques undertaken at the Weatherley research catchment and used in combination, enabled identification of the dominant runoff generation processes at the hillslope and headwater scale, with particular emphasis on subsurface flow processes (Wenninger *et al.*, 2008). The classical hydrometric techniques (measurement of precipitation, runoff discharge, and soil water status and groundwater levels) were supplemented with electrical resistivity imaging and tracer studies.

In Figure 43 the observation network presented includes two transects with nests of automatic recording tensiometers and shallow and deep groundwater boreholes, as well as neutron probe access tubes in an upper and lower sub-catchment, each gauged with a crump weir. The lower transect, comprising the nests 1–11, runs east–west, from the crest of the eastern slope down into the wetland, across the stream and up the western slope.

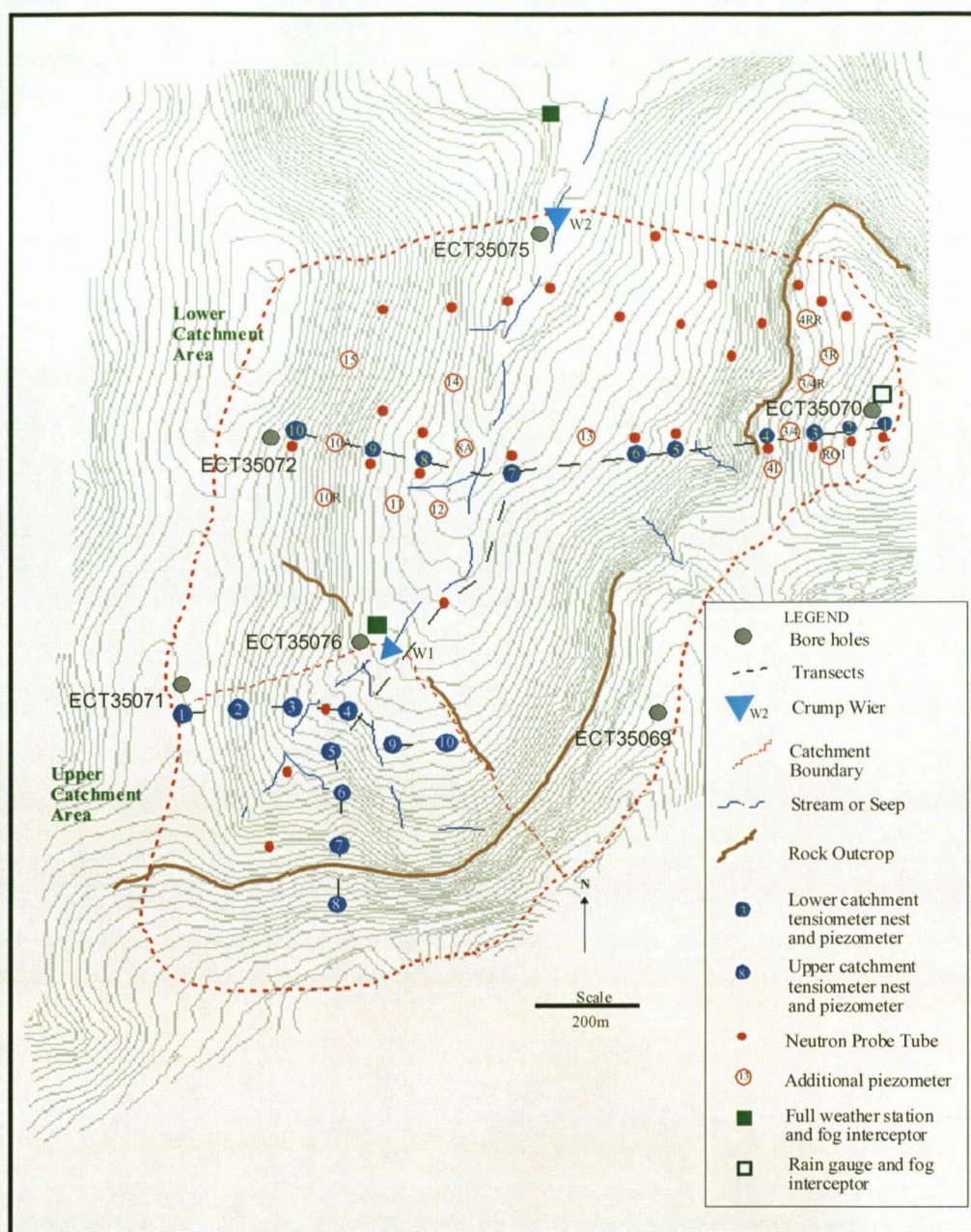


Figure 43: Observation network in the Weatherley catchment showing position of boreholes, and other observation points (Lorentz *et al.*, 2007)

3.3.6 Groundwater occurrence in the Weatherley catchment

Figure 44 and Table 6 indicate that groundwater level depth in deep boreholes, within the riparian zone (also note boreholes ECT35075 downstream and ECT35076 upstream in Figure 43) varies between 3 and 5 m, while along the hillslope the groundwater water level ranges between 12 and 19 m (ECT35069, ECT35071 and ECT35072) below surface. On the crest of the hill (ECT35070) the water level depth ranges from around 27 m to 48 m.

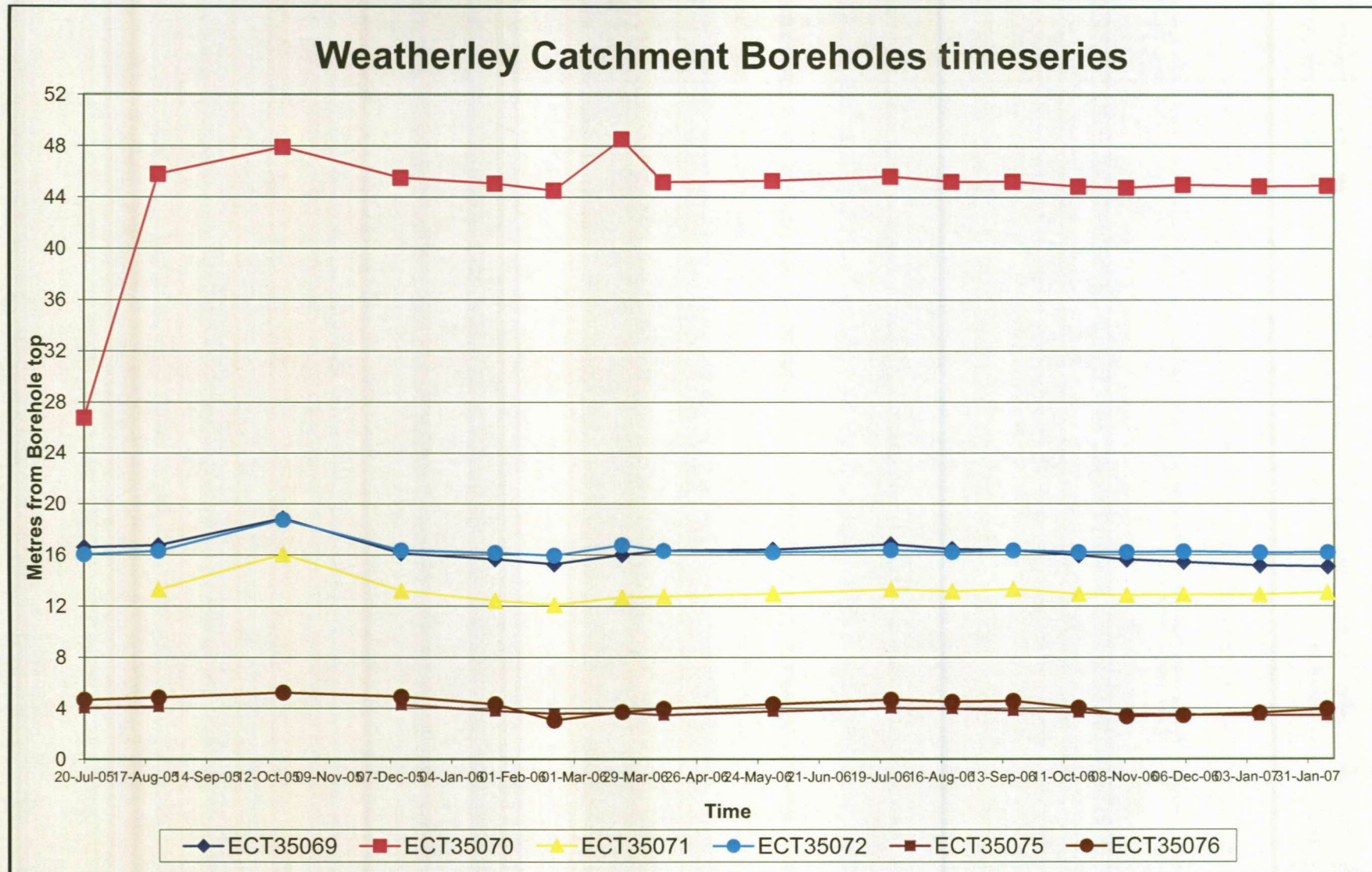


Figure 44: Groundwater level in deep boreholes in the Weatherley catchment. (Source data - Lorentz, 2011)

The relative depth to groundwater level in boreholes on the hillslope (e.g. ECT35072) and in the floodplain (ECT35075) is also quite clear from Figure 44.

Observations of boreholes near the stream revealed that the stream does not intercept the aquifer and hence the layers below the stream are also found to be dry up to 20 m depth (Lorentz *et al.*, 2007). Yet the interflow from the hillslope contributes to the stream through underlying fractured rock, thus sustaining the flow in the stream especially during low flow periods. 2D electrical resistivity surveys were undertaken in Weatherley to determine the water distribution in the system. That increase in electrical resistivity is related to a decrease in water content was used to delineate water distribution and spacing between potential electrodes enabled variation in depth and resolution to which survey could be done. Lorentz *et al.*, (2007) found that resistivity values of less than 60 Ω m in the sandstone indicate the presence of water of less than 60 ohms. An example of electrical resistivity survey is shown in Figure 45.

Table 6: Groundwater level in deep boreholes in the Weatherley catchment (Lorentz, 2011)

Date	ECT35069	ECT35070	ECT35071	ECT35072	ECT35075	ECT35076
20-Jul-05	16.62 m	26.78		16.05	4.02	4.67
23-Aug-05	16.76 m	45.8	13.28	16.32	4.14	4.86
19-Oct-05	18.87	47.9	16.02	18.77		5.23
12-Dec-05	16.13	45.5	13.18	16.36	4.27	4.91
24-Jan-06	15.66	45.05	12.39	16.17	3.78	4.32
20-Feb-06	15.26	44.51	12.06	15.95	3.58	3.04
23-Mar-06	16	48.52	12.67	16.77	3.63	3.71
11-Apr-06	16.34	45.17	12.73	16.32	3.46	3.96
31-May-06	16.42	45.28	12.94	16.2	3.75	4.32
24-Jul-06	16.82	45.62	13.28	16.37	4	4.67
21-Aug-06	16.46	45.22	13.14	16.22	3.96	4.5
18-Sep-06	16.35	45.23	13.316	16.35	3.81	4.57
18-Oct-06	15.97	44.855	12.93	16.22	3.67	4.03
09-Nov-06	15.63	44.77	12.85	16.225	3.505	3.35
05-Dec-06	15.44	45	12.89	16.27	3.5	3.44
09-Jan-07	15.16	44.88	12.89	16.19	3.43	3.63
09-Feb-07	15.08	44.92	13.06	16.2	3.43	3.95

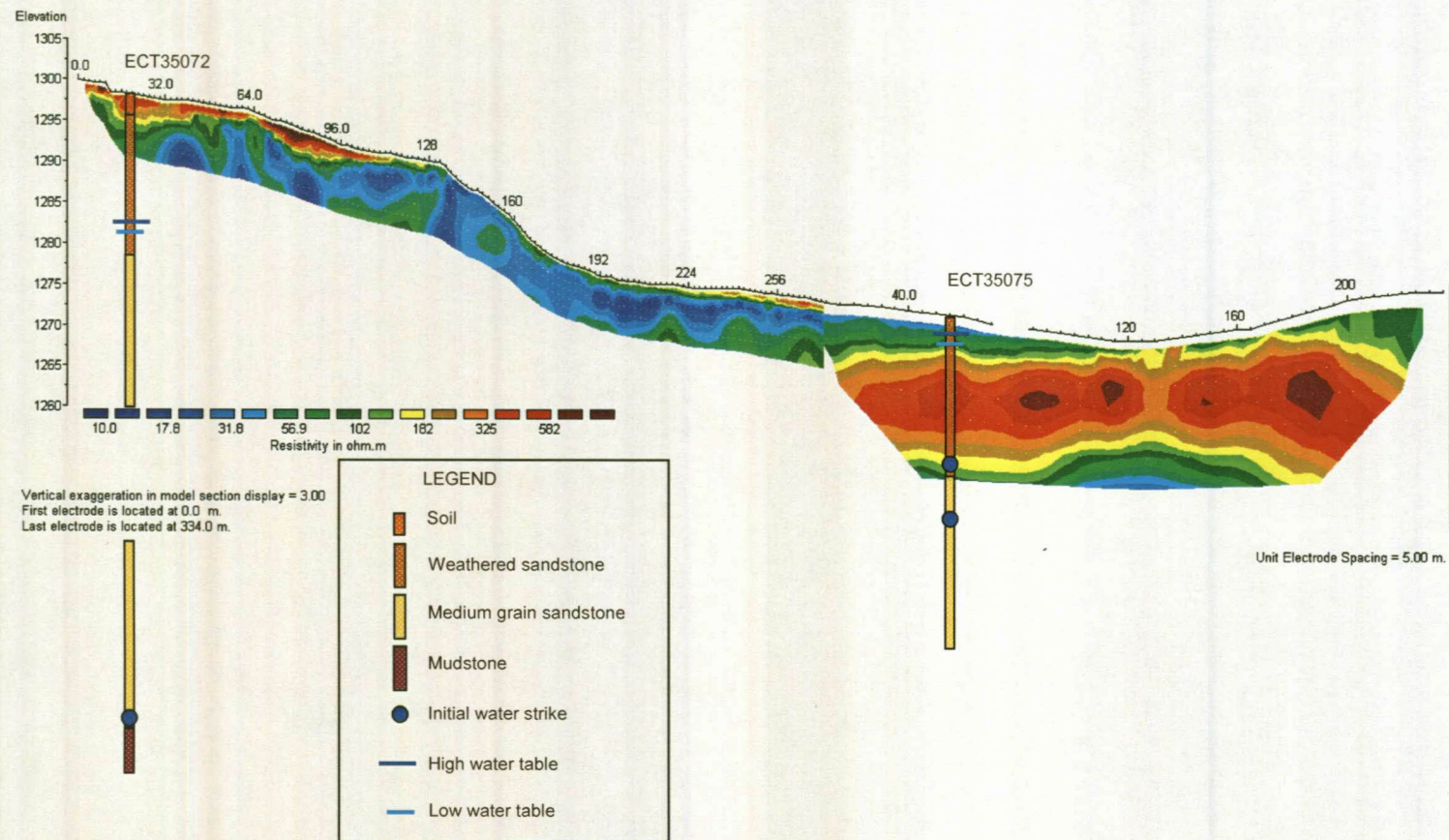


Figure 45: 2D resistivity transect (5m spacing) through stations 1-4 in the lower catchment (Feb 2004) showing the borehole log and water level (Lorentz, *et al.*, 2004)

3.3.7 Environmental isotope data analyses

Figure 46 shows that the isotopic signatures for both shallow groundwater and the river water are relatively more enriched in deuterium, probably due to evaporation of the lighter isotopes. All isotope data samples plot nearer the evaporation line. It can be deduced from isotope data analyses that water samples from boreholes in the vicinity of the weirs (ECT35075 & ETC35076) and river water samples have similar isotope signatures (both relatively enriched in terms of deuterium). This similarity in isotopic signature shows that there is some interaction between shallow groundwater and river water.

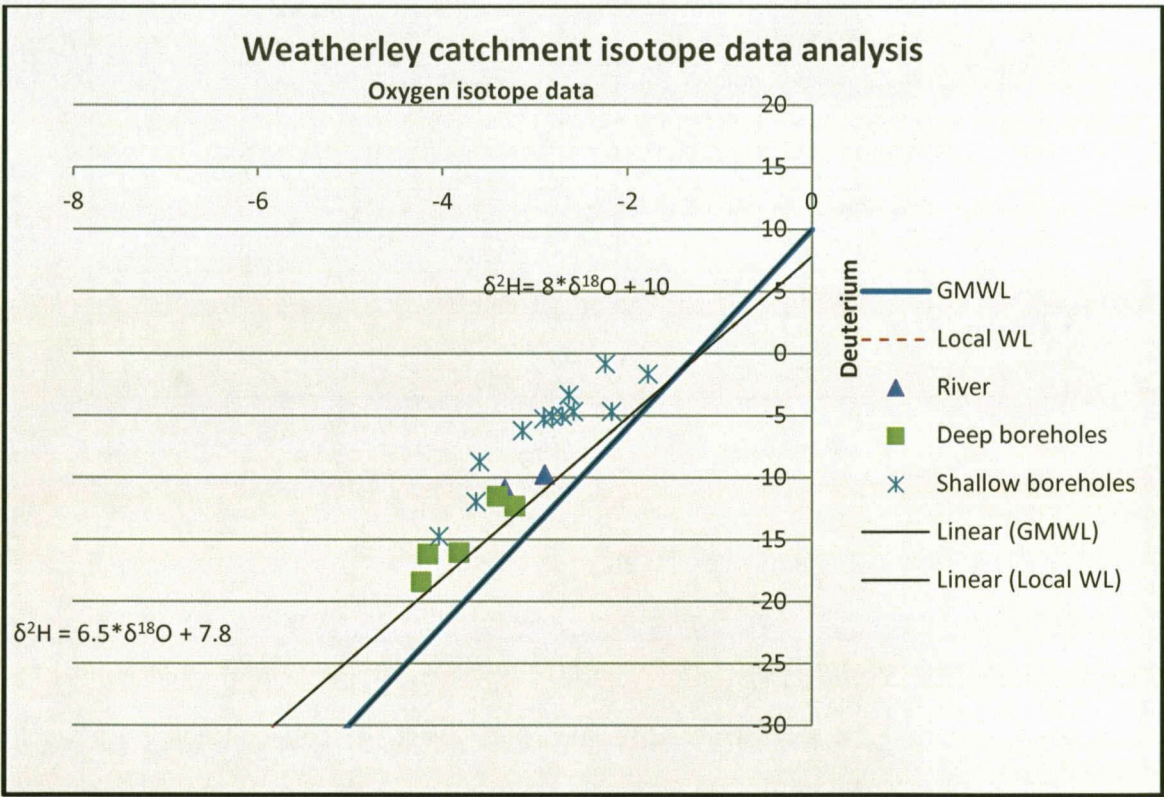


Figure 46: Isotope data analysis for the groundwater and river water in the Weatherley catchment

Table 7 depicts the isotope data captured from the Weatherley catchment, deep and shallow groundwater, and surface water as well as near surface water samples.

Table 7: The Weatherley catchment isotope data (Source data - Lorentz, 2011)

Details of observation points	Symbol used	$\delta^{18}\text{O}$	$\delta^2\text{H}$
Lower catchment weir	weirLc	-3.32	-10.91
Lower catchment weir	WeirUc	-2.89	-9.74
Deep borehole (SE hillslope)	BH69	-4.15	-16.16
Deep borehole (NE hillslope)	BH70	-4.22	-18.41
Deep borehole (NW hillslope)	BH72	-3.81	-16.04
Deep borehole (d/s near weir2)	BH75	-3.21	-12.29
Deep borehole (u/s near weir1)	BH76	-3.4	-11.49
Piezometer 4L (on NE hillslope)	P4L	-2.9	-5.25
Piezometer 8A (d/s of weir1)	P8A	-2.24	-0.82
Piezometer in the lower catchment	Lc04	-3.13	-6.21
Piezometer in the lower catchment	Lc05	-3.59	-8.72
Piezometer in the lower catchment	Lc08	-2.63	-3.33
Piezometer in the lower catchment	Lc09	-2.58	-4.66
Piezometer in the lower catchment	Lc10	-2.7	-5.01
Piezometer in the lower catchment	Lc3/4	-4.03	-14.74
Piezometer in the upper catchment	Uc01	-3.63	-11.94
Piezometer in the upper catchment	Uc05	-1.78	-1.64
Piezometer in the upper catchment	Uc06	-2.81	-5.06
Piezometer in the upper catchment	Uc3/4	-2.17	-4.65

3.3.8 Hillslope hydrological processes

The water flow regime is geologically and structurally controlled. For instance, the relatively resistant Molteno sandstone layers form the typical breaks in the hillslopes below which seepages are generally often observed, particularly at the contact with the low permeability mudstone-siltstone. When it rains, infiltration dominates in the upper areas of the hillslope (Figure 47). The gentle slopes and dense vegetation impedes overland flows and facilitate infiltration.

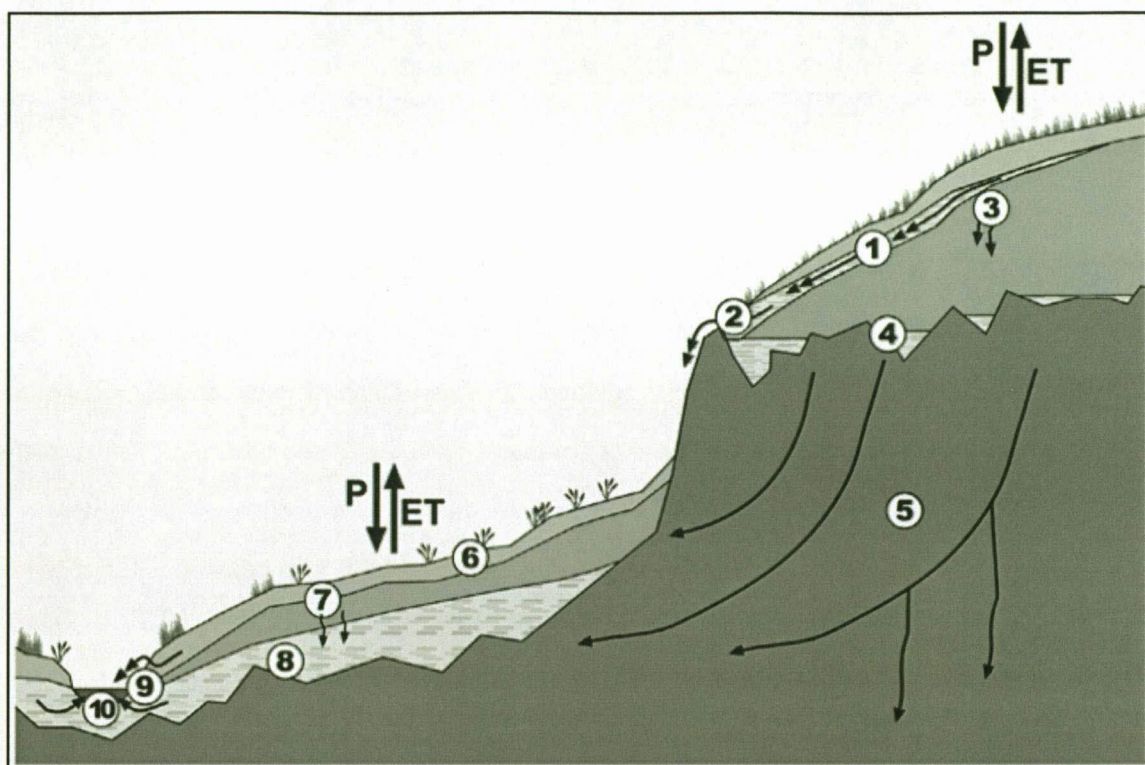


Figure 47: Flow model mechanisms in the Weatherley research catchment (Wenninger *et al.*, 2008)

There were no signs of wetness recorded in 240mm up to a depth of 1500 mm, indicating that water does not perch in the pedon (basic unit of soil) within this depth. That means some water infiltrate the fractured rock to recharge regional aquifers (Lorentz, 2001). At the soil-bedrock interface, interflow dominates with lateral flow becoming more prominent especially where flow encounters layers with low permeability. When the amount of water exceeds the storage capacity of the soil such water returns to the surface contributing to overland flow.

The valley bottom is covered by clayed soils which is an indication of long periods of saturation. These saturated conditions favour overland flow due to saturation impairing infiltration.

3.3.9 Lessons learnt from the Weatherley Research Catchment

From the Weatherley catchment case study it can be concluded that

- Understanding hill-slope hydrology requires a combination of study approaches including a variety of experimental techniques and observational monitoring over a long time.

- The water flow regime is controlled by geology, structure and vegetation as demonstrated by prominent sandstone layers that typically form breaks in the hill-slopes below which seepages (or interflow) are often observed while densely vegetated gentle slopes facilitate infiltration where overland flow is impeded.
- A combination of natural isotope and chemical species tracers sampling together with hydrometric observations of surface water and subsurface water processes and flow mechanisms is what it takes to access surface and subsurface flow exchanges.
- Lack of chemistry data (macro elements) made it difficult to use chemistry in this case study. The Weatherley catchment is one of the few field laboratories in South Africa with some of the best instrumentation that provide valuable information. Benefits may however, be enhanced if additional data (e.g. chemistry data) can be captured for research purposes.

3.4 Case Study: Seekoei River

3.4.1 Location

The Seekoei River is a non-perennial tributary of the Orange River located in the north west of the Eastern Cape Province, approximately 250 km southwest of the city of Bloemfontein (Seaman et al, 2010). The catchment, which falls in the Upper Orange Water Management Area (WMA), lies in the D3 sub-drainage region and comprises quaternary catchments D32A to H and D32J to K (Figure 48). The map also shows the location of the observation points within the riparian zone along the stem of the Seekoei River; from the south west (upstream) in the D32E quaternary catchment (e.g. EWR1 is 27 km south of Hanover) to just upstream of Vanderkloof Dam about 25 km north west of Colesberg (D3H015) in the north east part of the study area. The sites include the flow gauging weirs (i.e. D3H001, D3H007 and D3H015) and environmental water requirement sites (EWR1, EWR2, EWR3 and EWR4). Within each of the latter sites, there are also observation boreholes (as shown on this map).

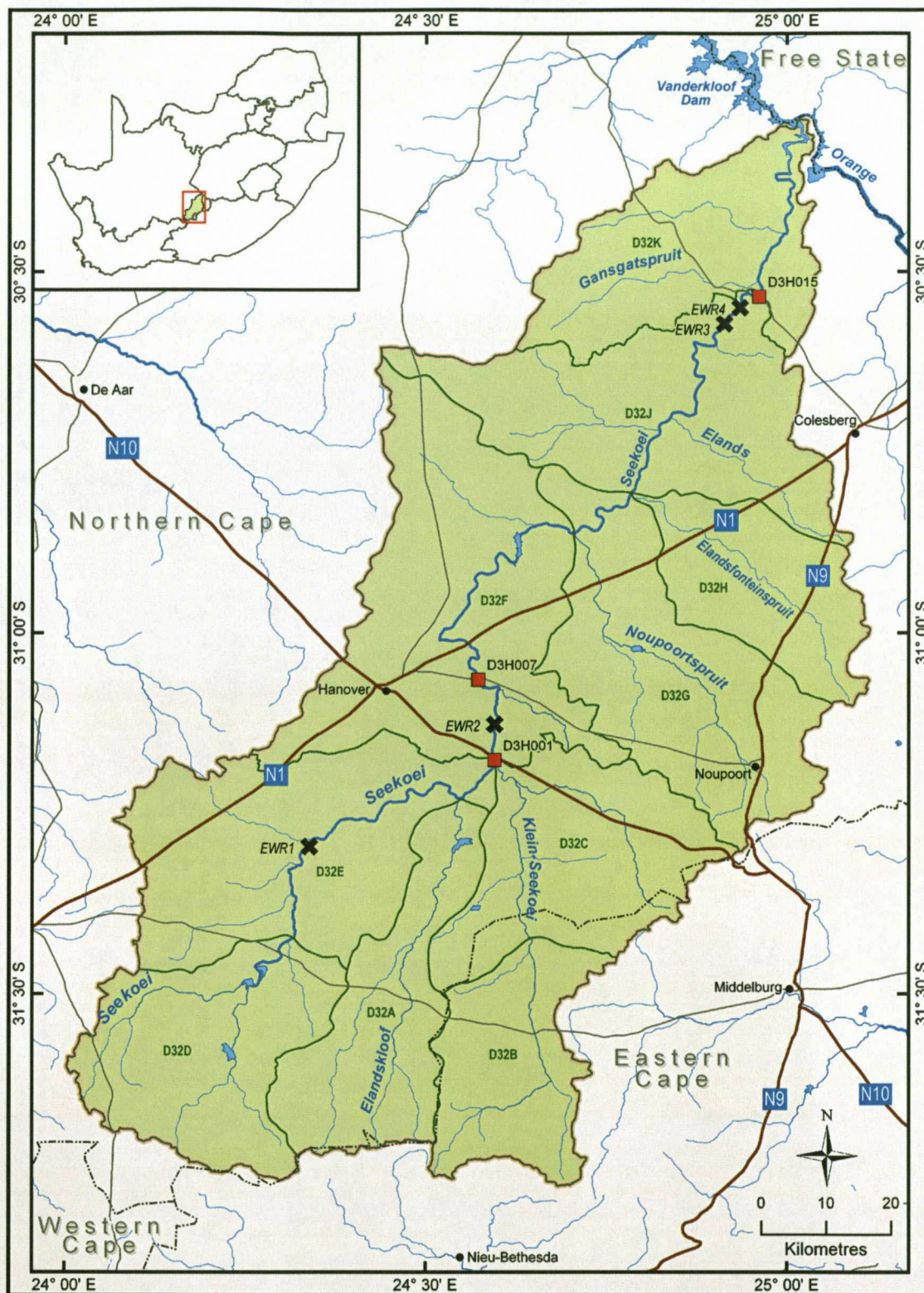


Figure 48: Location map of the study area, Seekoei River Catchment showing rivers, position of flow gauging weirs and environmental flow requirement sites. (Seaman *et al.*, 2010)

3.4.2 Topography

The near channel environments are typically very low gradient and steeper valley side slopes area mostly experienced a long way from the river close to the catchment boundary (Figures 49 and 51). Hence the topography of the Seekoei River catchment is generally flat with the mean catchment slope of around 1 to 4% (Hughes, 2008). Steeper slopes occur closer to the catchment boundaries, as well as in an isolated area in the lower part of the catchment, where the river passes through a gorge (quaternary catchment D32J).



Figure 49: The site is clearly relatively flat along the riparian zone with steeper sides (as can be viewed in the foreground). (Van Tonder, 2012)

Figure 50 shows two Google Earth images of the catchment, the left hand side being the area within D32J, while the right hand side illustrates the characteristics of the area further upstream which is more representative of the catchment as a whole (Hughes, 2008). The left hand side image clearly shows steeper topography where the river passes through a 'gorge' related to the occurrence of a dolerite ridge. This area is expected to have very different hydrological response characteristics to the rest of the catchment.

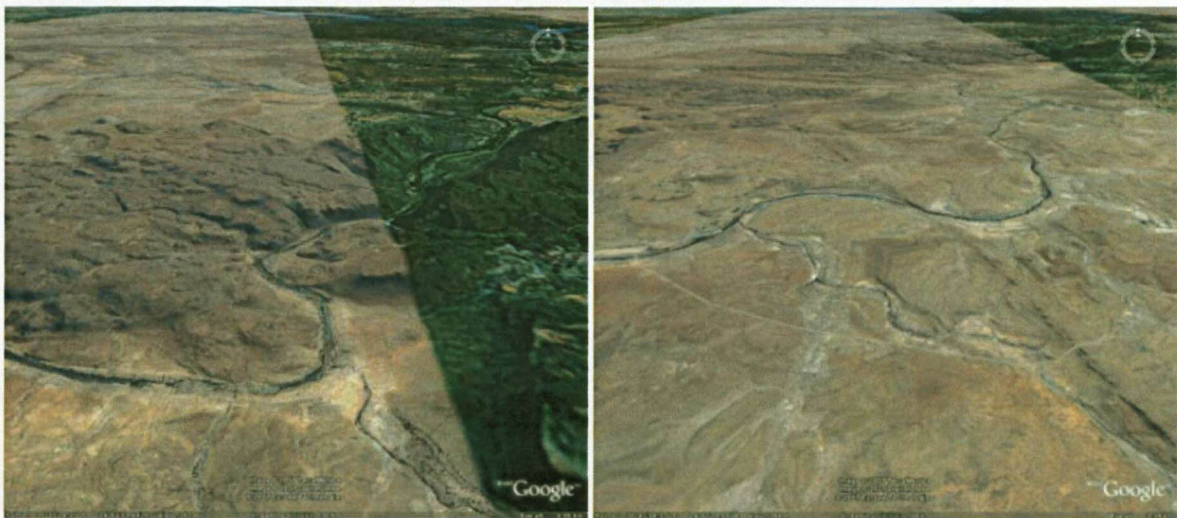


Figure 50: Google earth images of the Seekoei River catchment upstream (Hughes, 2008)

3.4.3 Geology and structure

The geology of Seekoei area consists of interbedded sandstones, shales and mudstones of the Beaufort Group with relatively frequent dolerite intrusions, some of which can be highly weathered. It appears that shallow alluvial aquifers can be found below the river channel in some areas, while in others the channel is clearly developed on solid rock (Hughes, 2008). Figure 52, depicts geology with ubiquitous dolerite intrusion that has obviously contributed to the current landscape and the older Karoo sediments. The upper and middle sections of the catchment are dominated by Adelaide Subgroup mudrocks and subordinate sandstones, with intrusions of dolerite, while the lower catchment comprise of Tierberg Formation shales, siltstones and sandstones and dolerite-capped koppies (Seaman *et al.*, 2010).

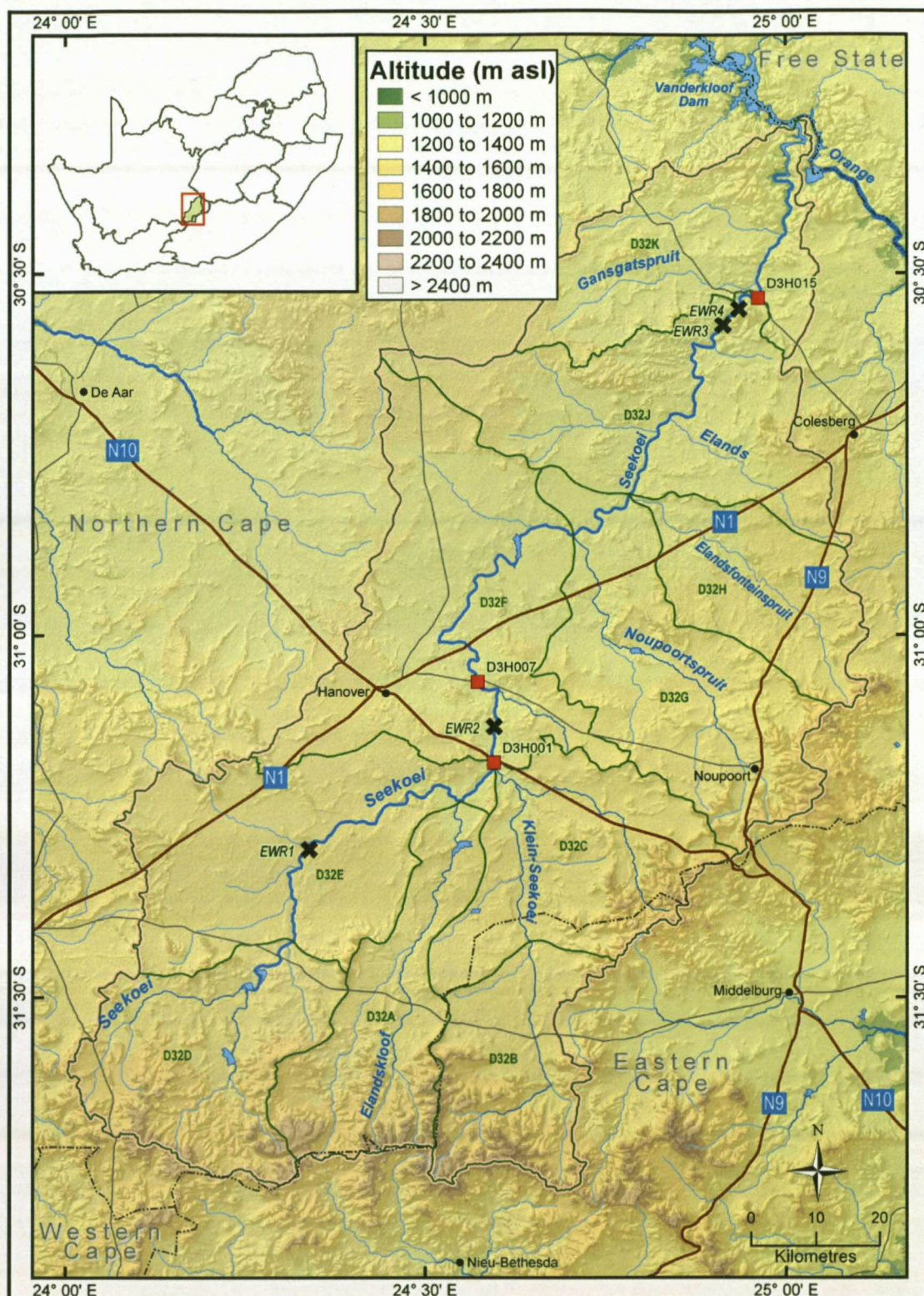


Figure 51: Topography of the Seekoei River Catchment, (Seaman et al., 2010)

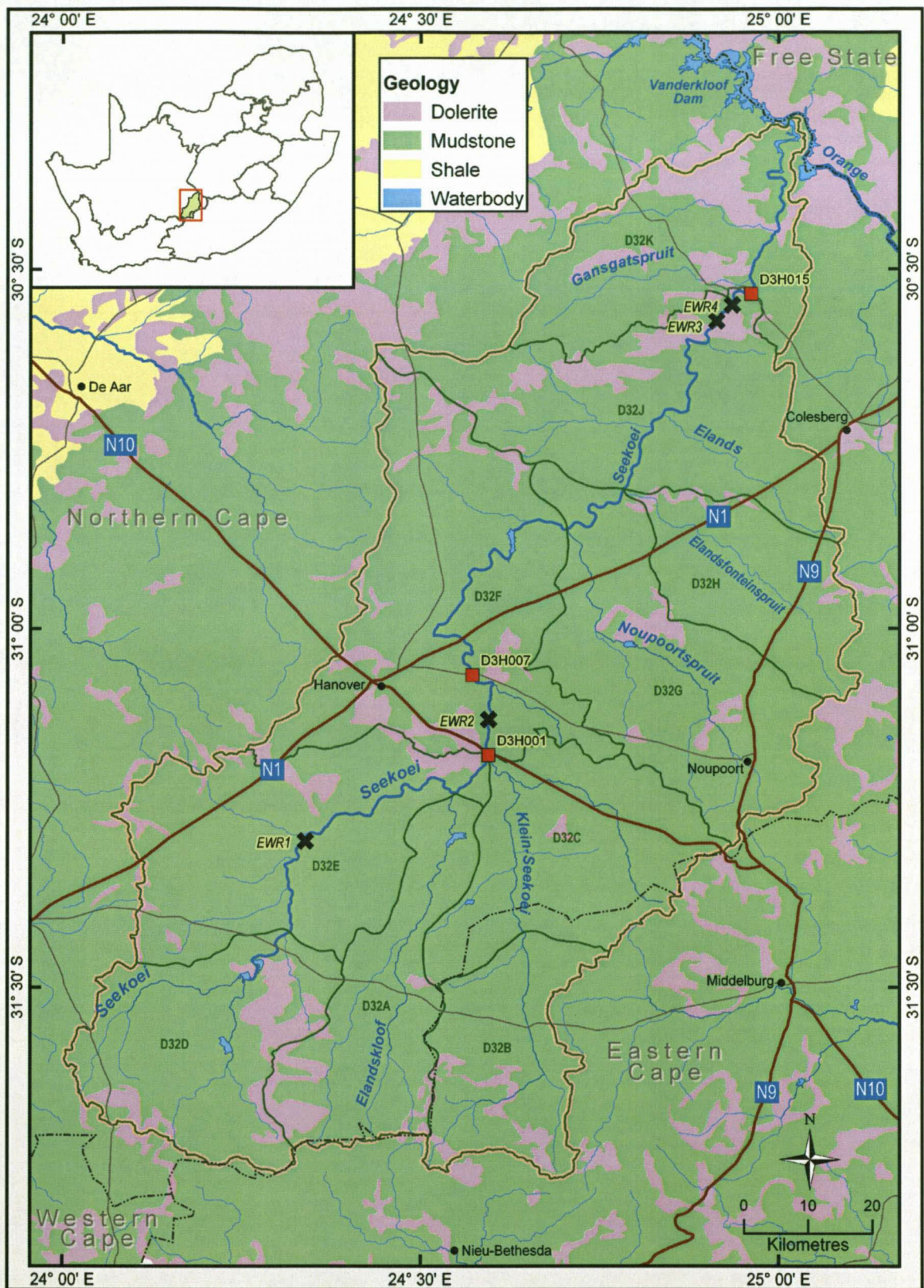


Figure 52: Geology of the Seekoei River Catchment, (Seaman *et al.*, 2010)

3.4.4 Soils, land cover and drainage network

Soils are mostly stony and thin, although they can be deep close to the river channels. Vegetation cover is very sparse except in some channel margins where presumably there is improved access to sub-surface water (Figure 53). Drainage densities are low, largely due to the low gradients and this implies that surface runoff processes will be dominated by surface sheet and shallow gully flow during heavy rainfall.

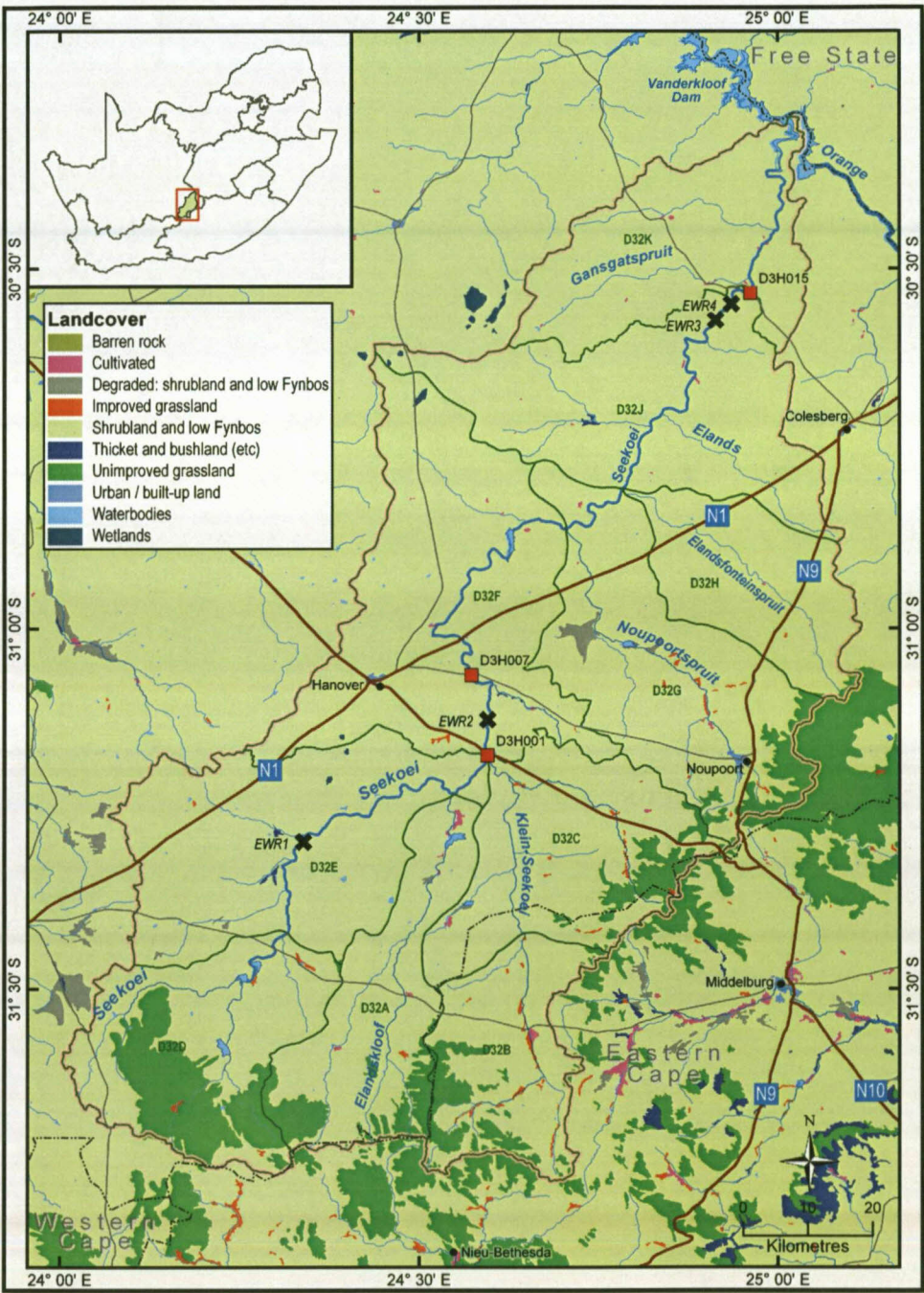


Figure 53: Landcover map for Seekoei (Seaman *et al.*, 2010)

3.4.5 Climate

Rainfall in the catchment is highly variable from month to month (with a monthly coefficient of variation of about 1.1) and between years. Figure 54 shows that the highest amount of rainfall, based on data available, occurs in summer particularly during February month, with the mean annual rainfall ranging of about 534mm. Evaporation (Figure 55) in the catchment varies between 1900 mm in the high-lying areas to 2500 mm in the western and lower part of the catchment, thus exceeding rainfall by a factor of more than six times.

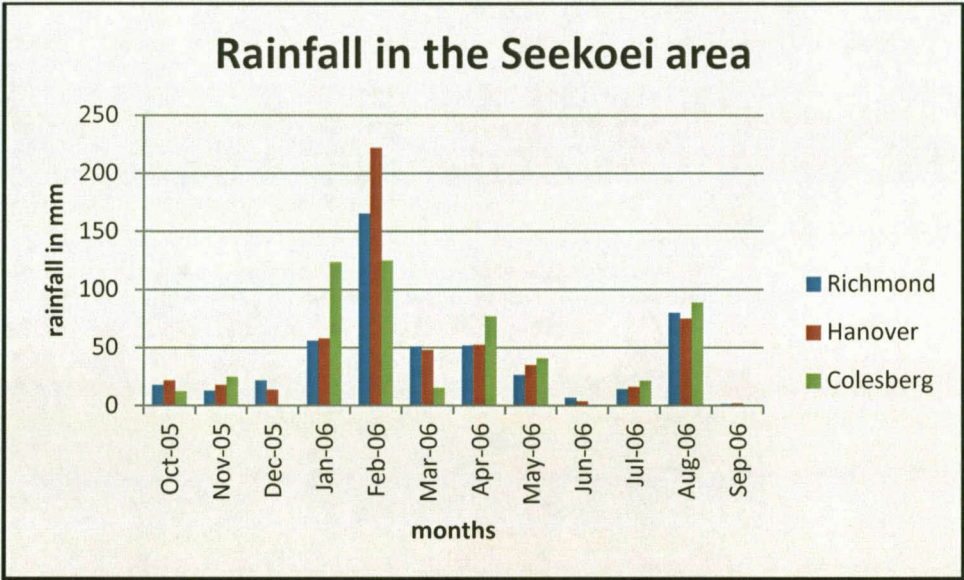


Figure 54: Seekoei – Rainfall based on measurements taken in the Richmond, Hanover and Colesberg area.
(Source data – Van Tonder, 2012)

3.4.6 Recharge and groundwater regime

The recharge of the Seekoei River catchment ranges between 10 and 25 mm per year (Figure 56). It can be deduced from this map that recharge is highest in the northeastern and south eastern parts of the catchment and lowest in the southwest where recharge occurs at a rate of approximately 3 mm per year. The ground water level in the catchment varies between 5 m below ground level to 10 m below ground level (Seaman *et al*, 2010).

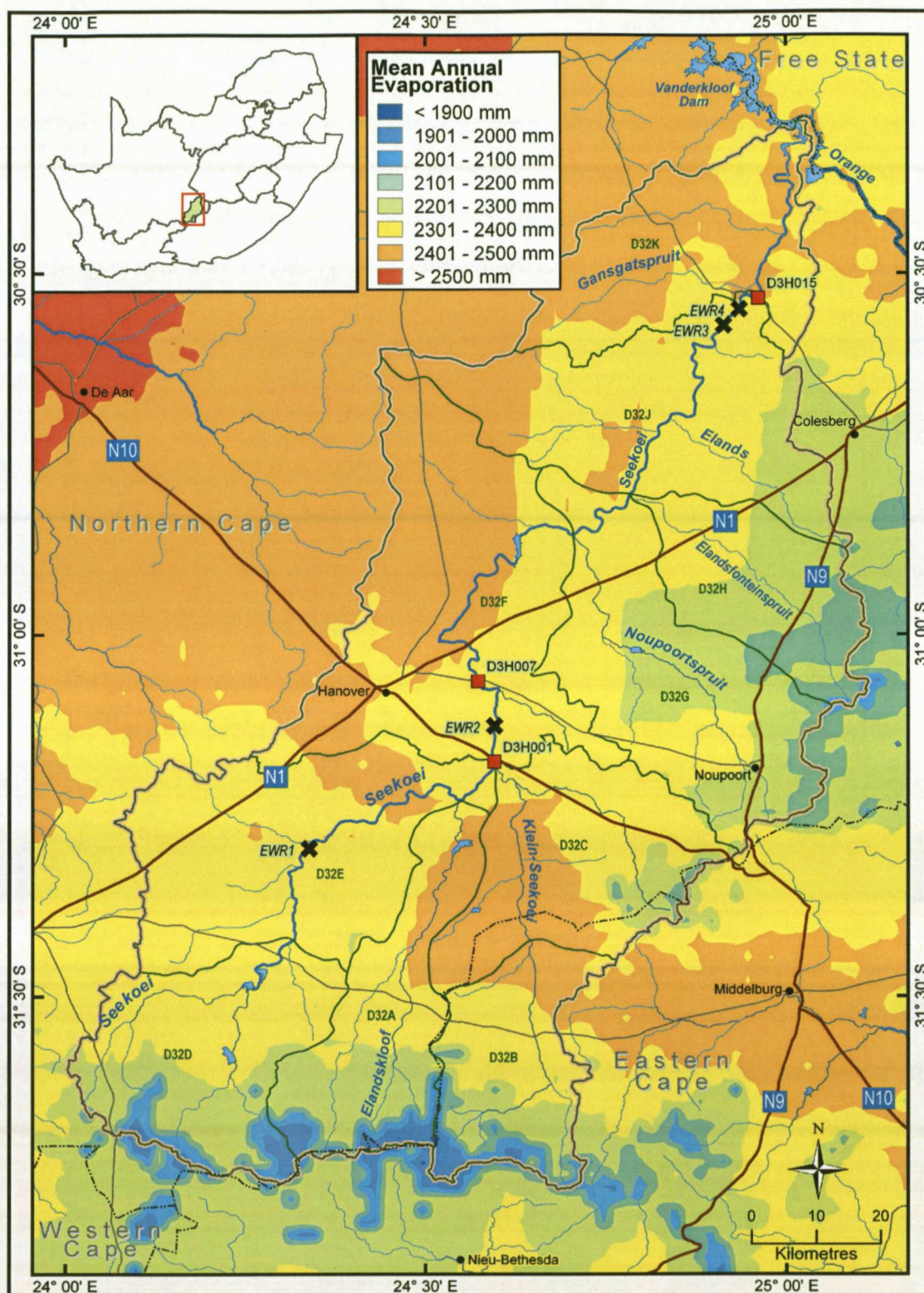


Figure 55: Evaporation rate of the Seekoei River Catchment (Seaman *et al*, 2010)

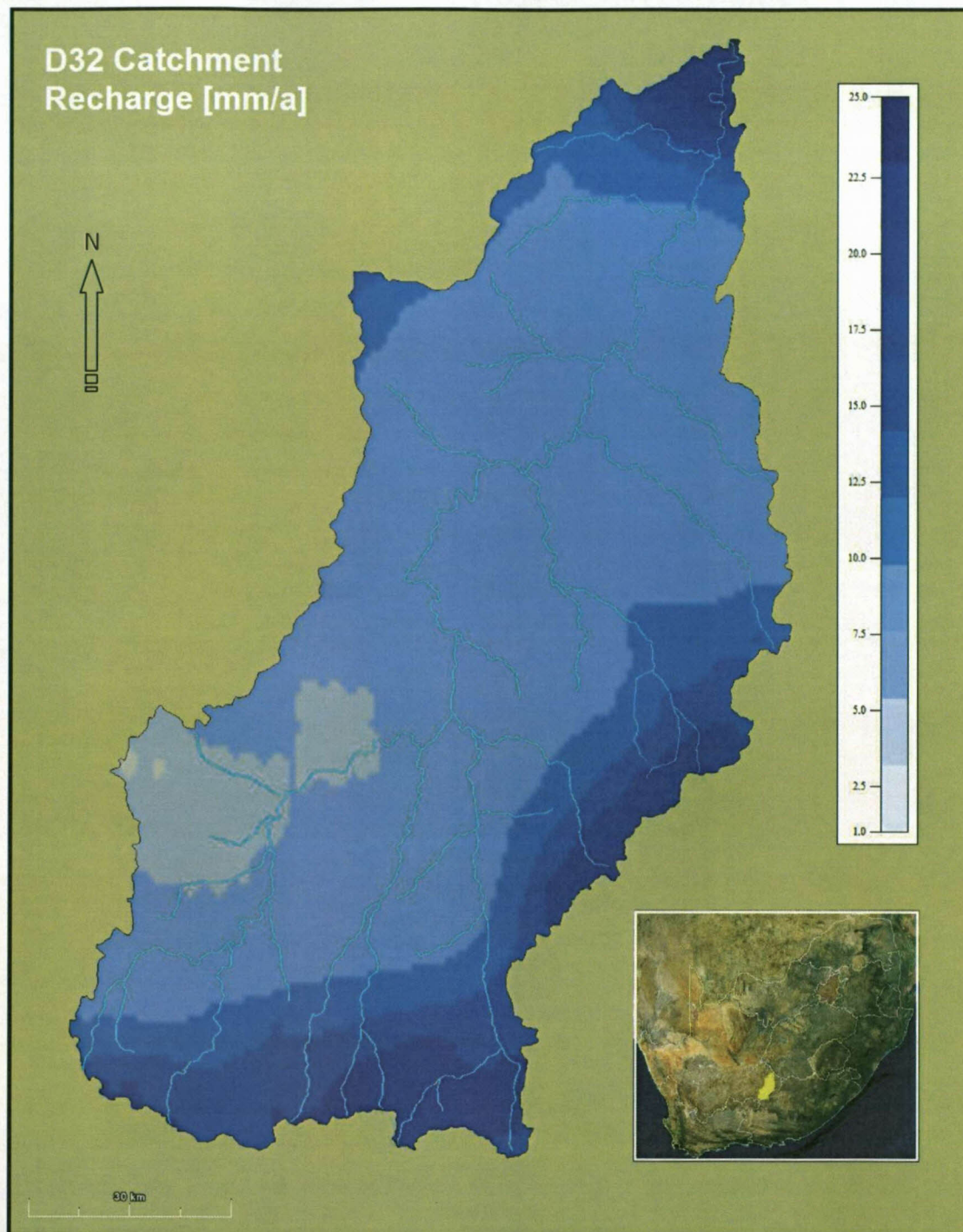


Figure 56: Recharge in the Seekoei River catchment prepared by R. Dennis (for Seaman *et al.*, 2010)

3.4.7 Conceptual model of the Seekoei area

Figure 57 represents the conceptual model of the area indicating that the channel losses to groundwater are likely to be a negligible component of the overall water balance. This is assuming that the water table is quite close to the channel bed (van Tonder *et al.*, 2007 and Hughes, 2008). There may be parts of the channel system that are losing water during

surface runoff events, while other parts of the channel system are gaining water through groundwater discharge (Hughes, 2008). However, on balance the losses are expected to be minimal.

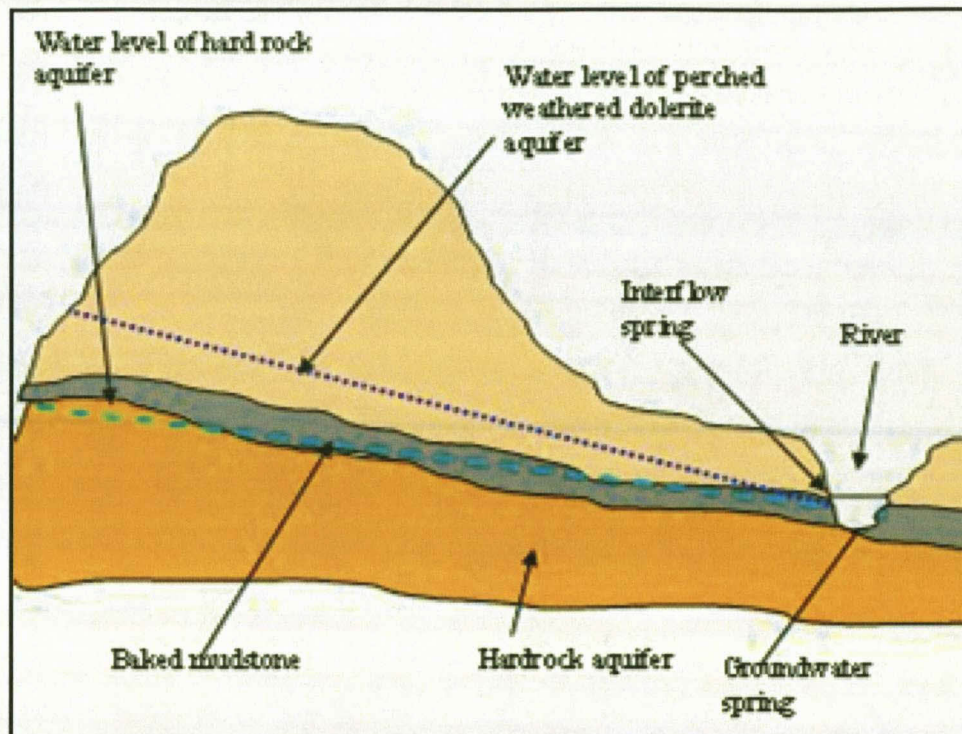


Figure 57: Conceptual model for interflow and groundwater springs in the Seekoei River (van Tonder *et al.*, 2007)

3.4.8 Groundwater flow dynamics and baseflow contributions

Studies indicate that the groundwater spring flow sustains pools during periods of no flow (van Tonder *et al.*, 2007). The spring water has a TDS of approximately 400 mg per litre; while the observed runoff at D3H015 has TDS values ranging from less than 100 to over 1 500 mg per litre and the highest TDS values occur after prolonged periods of baseflow or at the start of flow events that have very low flows. Based on this, Hughes (2008) assumed that:

- If the start of an event has very low flows, most of the runoff at D3H015 will be displaced pool water that has very high TDS values due to the concentrating effects of evaporation.
- If the start of an event has quite high flows the TDS will be more a reflection of surface runoff water quality, which appears to have low (± 100 mg per litre) TDS values.

- The rapid increases in TDS during the baseflow recession period suggest an additional mechanism apart from the spring flow and surface runoff already identified. This may be related to the storage of salts within the pools and adjacent soils which is incremented during pool drying and gradually released after pools have been re-filled.
- Relatively simple mass balance modelling of the system using assumed pool storage volumes, evaporation rates and TDS values for different water sources could provide a possible method for simulating the general trends of pool water quality under different flow conditions.

Van Tonder et al.,(2007) contends that withdrawal of water from a borehole that intercepts a perched aquifer could have significant impact on the river that is fed by interflow component of the baseflow, especially if the perched aquifer is a sustainable source. Hence consideration of interflow in the estimation of baseflow becomes important.

According to the spring owner the springs stopped flowing after about 4 months after good rains (van Tonder et al, 2007), thus indicating that such types of springs are mainly interflow from a perched aquifer. On the other hand both the Vaalkop springs (Spring 1 & 2 in Figure 58) are regarded as originating from groundwater as neither of them has ever stopped flowing during the past 40 years. It is also very interesting to note that flow of both the Vaalkop springs increases during the winter, about 30% more flow, due to delayed influence of recharge and since evapotranspiration is lower during the winter. EC of 61 and 58 mS/m were obtained for springs 1 and 2 respectively compared to 90 and 86 mS/m for EWR3 and EWR4 respectively downstream. This progressive increase in EC towards the riparian zone is primarily due to evapotranspiration of the vegetation that result in precipitation of salts.

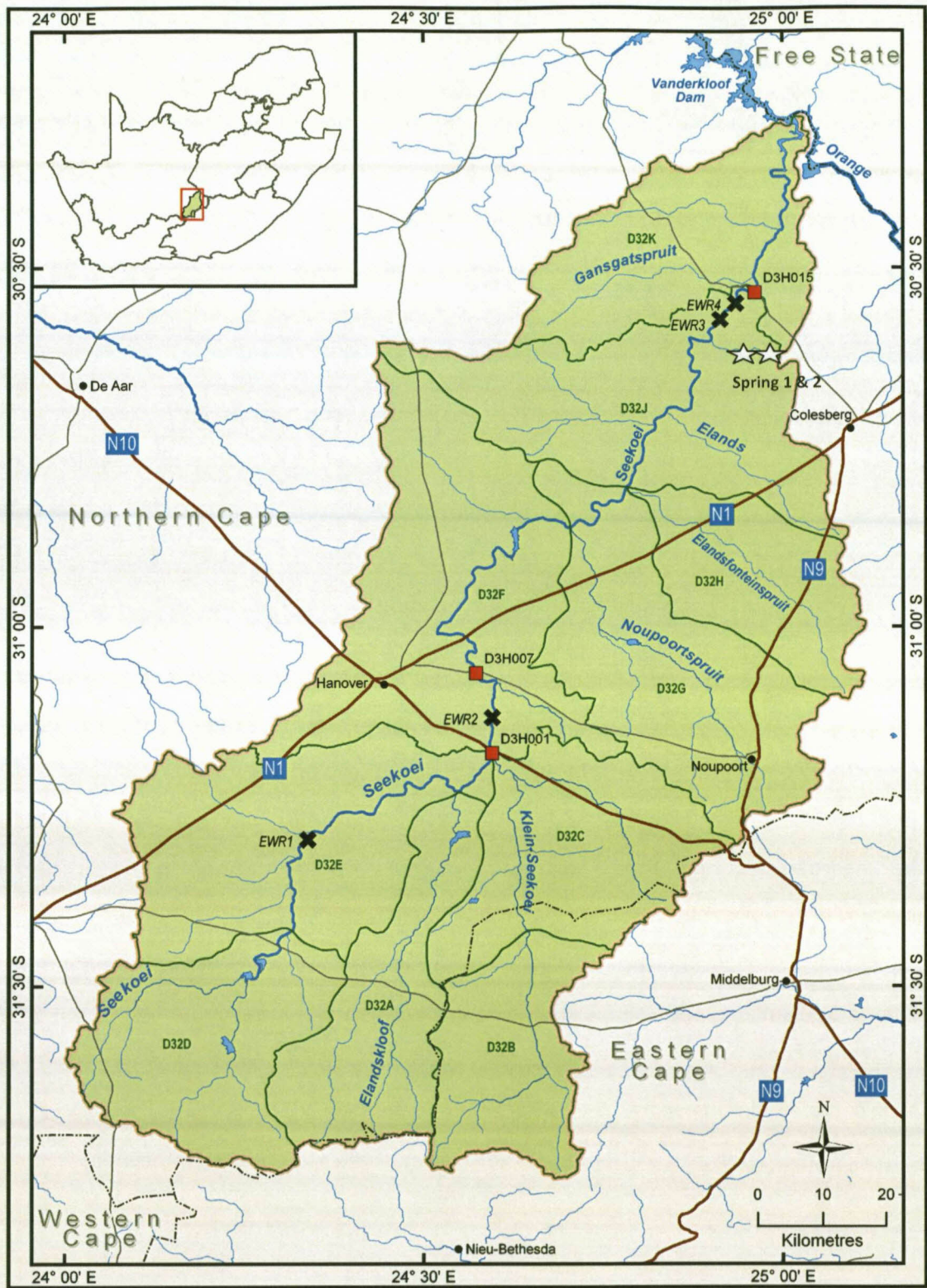


Figure 58: Location map for Vaalkop springs (1 and 2) upstream of EWR3 & 4 (adapted from Seaman *et al.*, 2010)

The movement of water to the channel is considered to occur within the perched water table associated with weathered dolerite, as well as within the hardrock aquifer. The alluvial material that includes sand, clay and weathered dolerite material beneath the channel bed is also considered to play a role in the sub-surface movement of water in the direction of the channel. Contributions to the channel are expected to be highly localized (in springs) due to structural differences and the occurrence of more transmissive fracture zones and weathered material. These contributions are expected to contribute to pool storage, support riparian vegetation while some is lost to evaporation. The exact water balance in any specific part of the channel system will largely depend on the balance between the seepage contributions and the evaporative losses (van Tonder *et al.*, 2007).

The low gradients in the side slopes even far away from the channel suggest that seepage rates would be very slow. The low gradient topography and shallow, stony soils suggest that there is a substantial opportunity for recharge during rainfall events due to surface pondage and vertical drainage through macro-pores.

3.4.9 Water chemistry analyses and surface water - groundwater interaction

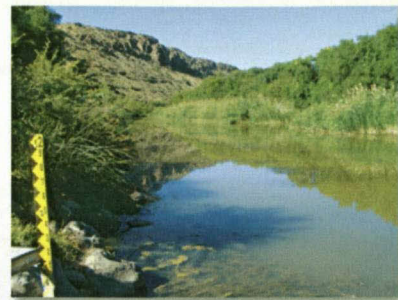
In the Seekoei area, the interaction between groundwater and surface water is evidenced by sustained pool storage in some channels during extended periods of no channel flow which implies that the storage is sustained by effluent groundwater. Hughes (2006) contends that pools that have been maintained over a long dry period in the area could partially be sustained by some source of sub-surface inflow. However, he also cautions that without additional monitoring sites speculation about the source of the water is riddled with uncertainties. The EWR3 (Figure 59) is located along Seekoei river about 35 km south west of the confluence of Seekoei and the Orange River (also see Figure 58).

Van Tonder *et al.*, (2007) found that the general groundwater flow around site EWR3 is towards the pool (Figure 61). The EC of the pool is also of the same order as EC obtained from the boreholes in the vicinity of the pool (Table 8). The EC for the river water (sampled at D3H015Q01 near and downstream of EWR4), is also of the same order as that of the pool water and groundwater in EWR3 (Figure 60).

EWR3



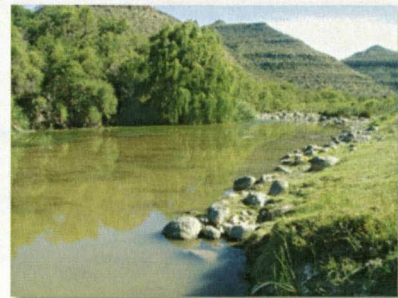
Upstream view March 2006



Upstream view March 2007



Downstream view March 2006



Downstream view March 2007

Figure 59: The Environmental Water Requirement site 3 (van Tonder *et al.*, 2007)

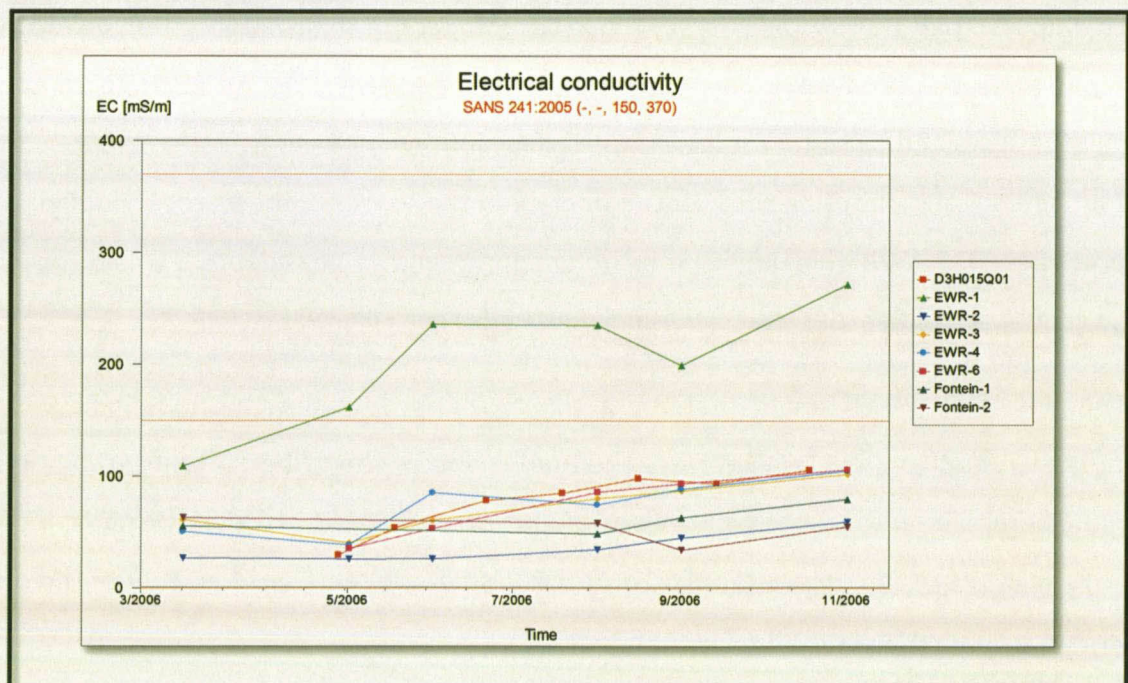


Figure 60: Seekoei - EC concentration for the river, pool and spring water (Source data – Van Tonder, 2012)

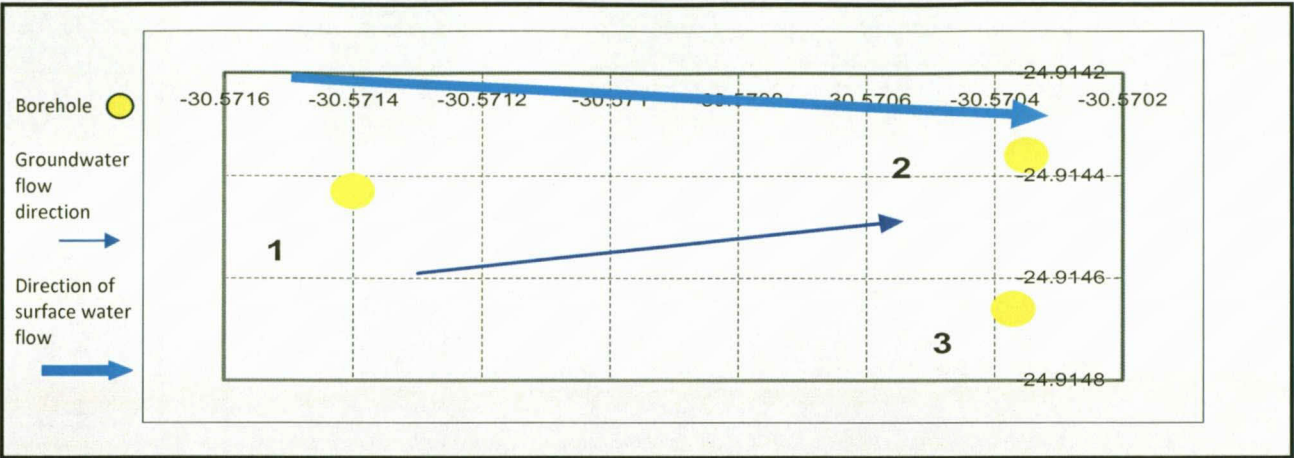


Figure 61: Direction of groundwater is towards the pool in EWR3 site (van Tonder *et al.*, 2007)

Table 8: EC of groundwater, pool and river water (Van Tonder *et al.*, 2007)

Feature (borehole/pool)	EC (average)
BG1	82
BG2	99
BG3	81
Pool	86
River (D3H015Q01)	78

Analysis of the spring water using Piper Diagram (Figure 62) indicates that both springs (i.e. Fonteins 1 & 2) plot on the CaHCO_3 field as expected since this water is mostly interflow that came from shallow perched water up the mountains. The pool water also however, plots in the same area CaHCO_3 area with relative enrichment in $\text{Na} + \text{K}$ and higher Alkalinity. The river water (D3H015Q01) also plots in the same field. This confirms that the pool water is mixed with shallow groundwater and surface water. Hence there is interaction between pool water, groundwater and pool water. Complete chemistry data is under Annexure 5.

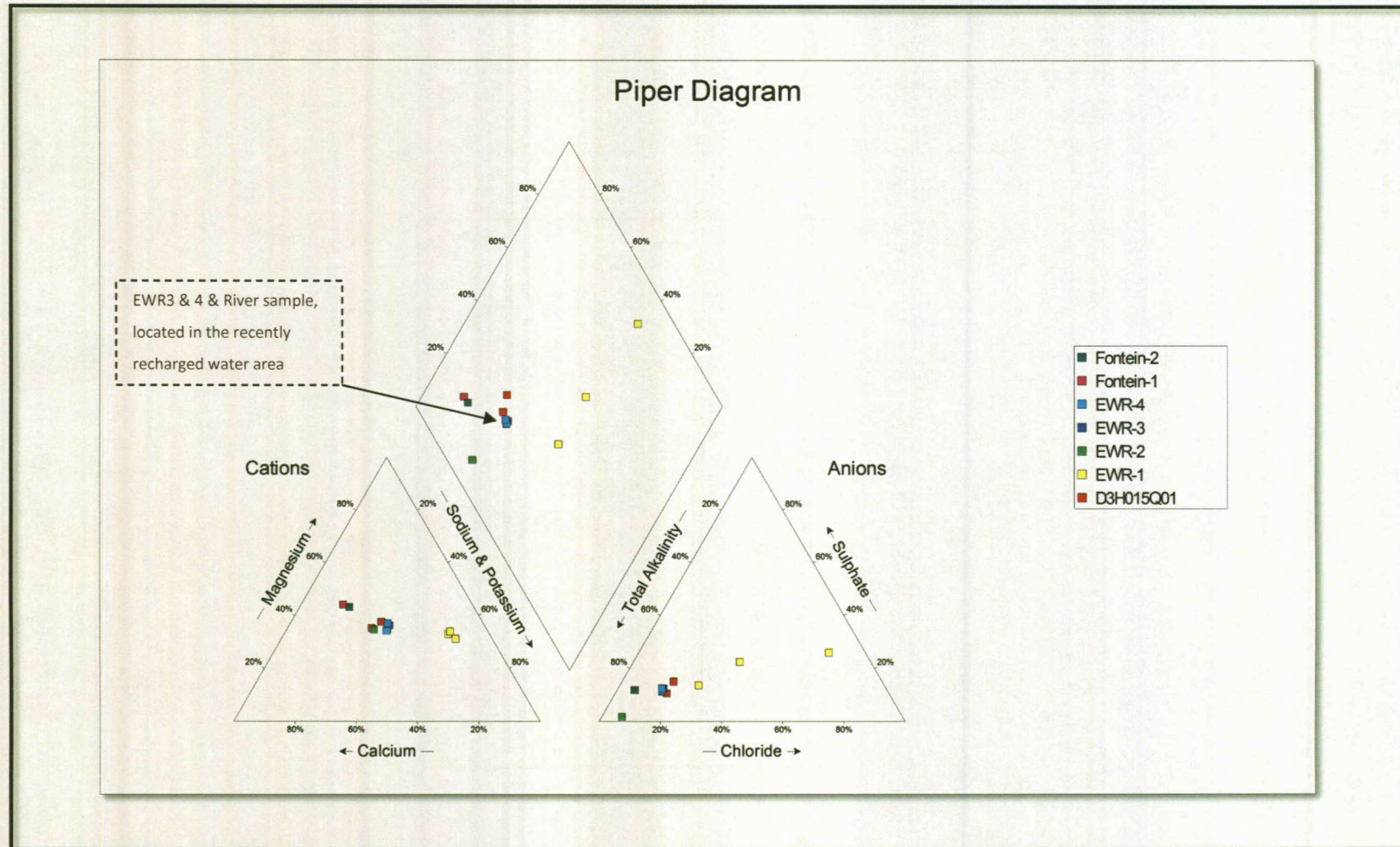


Figure 62: The EWR3 & EWR4 water, springs (Fontein 1& 2) and river water (D3HD15Q01) samples are located in the CaHCO_3 field (Source data – Van Tonder, 2012)

The stiff diagrams (Figure 63) for the river water (D3HD015Q01), EWR3 and EWR4 are also quite similar probably confirming potential interaction between groundwater and surface water. Chemistry of upstream water samples from EWR 1 & EWR2 differ.

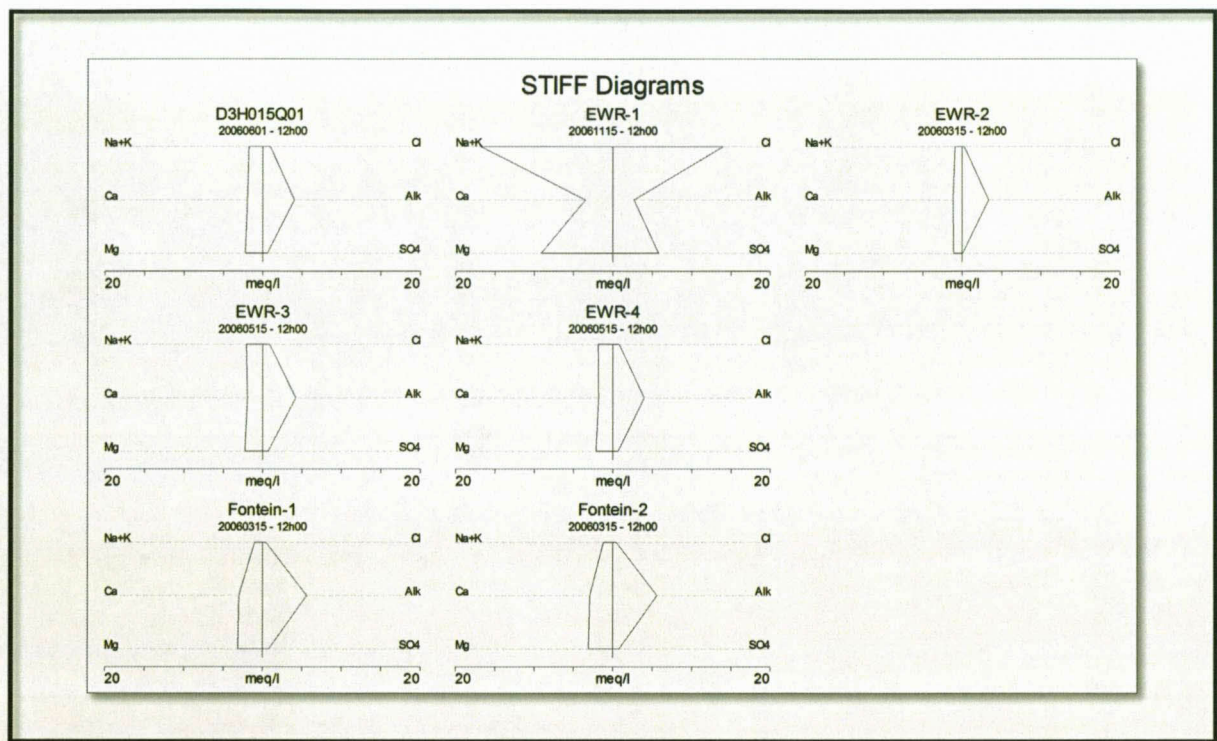


Figure 63: Stiff diagram for surface water (D3H015Q01), groundwater samples (EWR1, 2, 3 & 4) & spring water (Fontein 1 & 2) (Source data – Van Tonder, 2012)

However, when EWR1 which looks markedly different in chemistry to the rest is excluded, it becomes clearer that chemistry for EWR3 & 4 is similar to that of the surface water sample, while the concentration of spring water is slightly more alkaline and relatively Ca rich (Figure 64). This may still be indicative of the assertion that there has been exchange between groundwater and surface water.

If the upstream EWR2, is also excluded as indicated in Figure 65, the situation still remains unchanged.

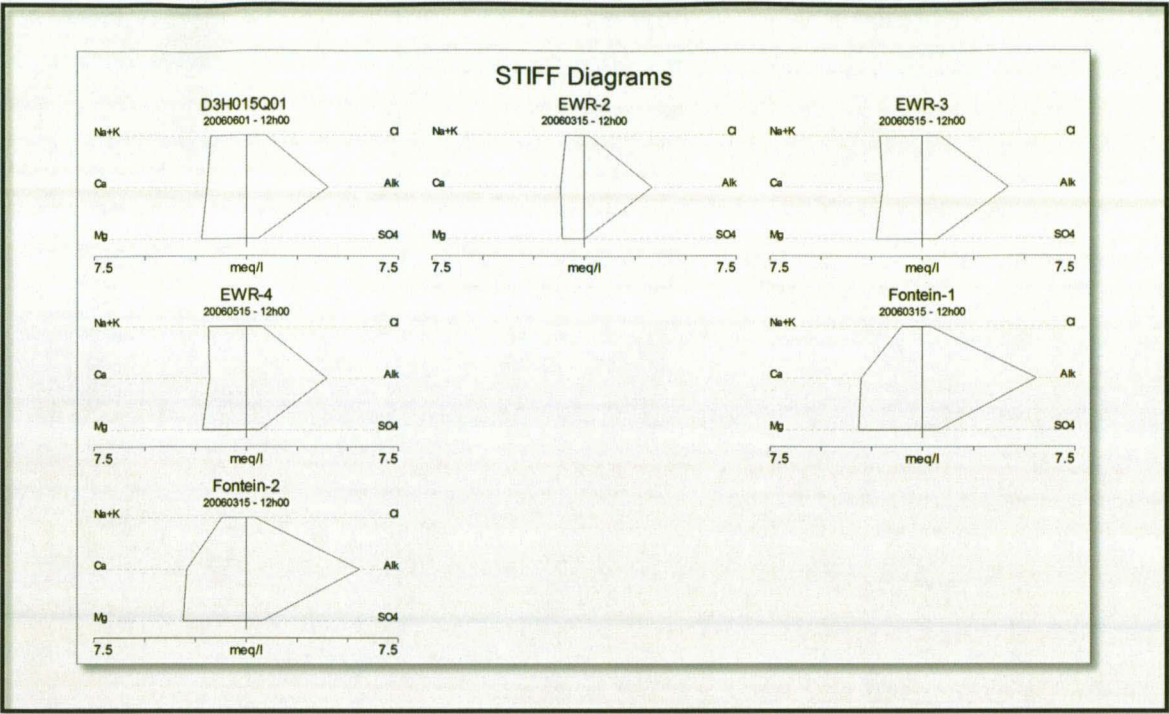


Figure 64: Stiff diagram for surface water (D3H015Q01), groundwater samples (EWR2, 3 & 4) & spring water (Fontein 1 & 2) (Source data – Van Tonder, 2012)

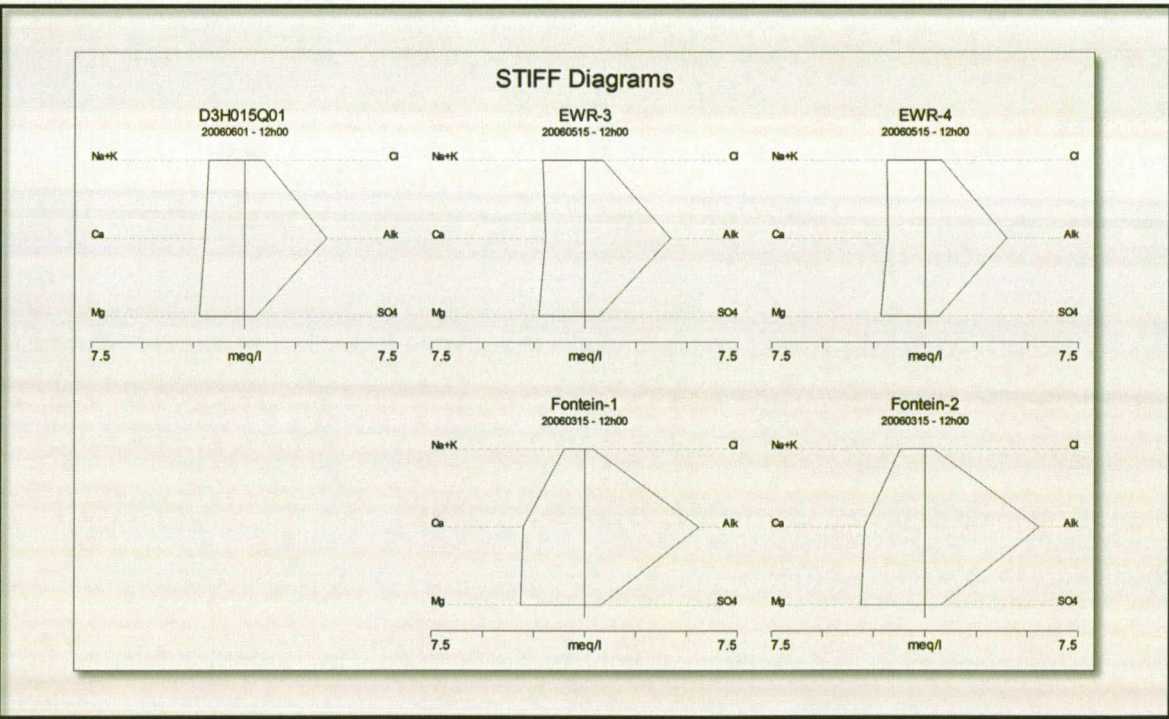


Figure 65: Stiff diagram for surface water (D3H015Q01), groundwater samples (EWR 3 & 4) & spring water (Source data – Van Tonder, 2012)

Also based on extreme changes in TDS (from 968 to 2 582 mg/l) of the pool water over a period of 3 months, van Tonder rightly concluded that the pool is groundwater fed and high TDS in groundwater is due to geology. In quantity terms most of the water is taken up by vegetation as Hughes (2008) and van Tonder *et al.*, (2007) have previously indicated. Losses from the pool storage may occur as direct evaporation from the pool surface, seepage into the banks of the pool to replenish soil moisture lost through evapotranspiration by riparian vegetation or through leakage from the pool bed as recharge to sub-surface water storage.

3.4.10 Key lessons learnt from the Seekoei case study

Some of the important issues that arise from the Seekoei case study are that

- High Total Dissolved Solids (TDS) values tend to occur after prolonged periods of low flows (or baseflow), or as a result of displaced pool water partly due to concentrating effects of evaporation. This implies is that TDS values could be used as indicators to distinguish pool water from surface water runoff, and the general trends of pool water quality under various flow conditions could also be determined
- Increased TDS during baseflow recession period is related to salts within pools and adjacent soils that are incremented during pool drying and gradually released upon refilling of pools.
- Taking interflow in the estimation of baseflow into account is important in the case of sustainable perched aquifer that is intercepted by a pumping borehole, especially if the perched aquifer is a sustainable source.
- Continuously flowing springs, particularly those in Vaalkop could be fed by groundwater; and flow increases were observed during winter due to delayed recharge and low evapotranspiration during that period.
- The water balance in any specific part of the channel system largely depends on the balance between the seepage contributions and the evaporative losses.
- Chemistry data is very useful in confirming exchange between groundwater and surface water.
- However, time series monitoring data (e.g. isotope, water levels, flows, chemical, and rainfall) need to be regularly monitored and collected.

3.5 Case Study: Mokolo Catchment

3.5.1 Location of the Mokolo case study area

This case study focuses on the interaction between the surface water and groundwater components within the Mokolo catchment, downstream of the Mokolo Dam. The area consists of quaternary catchments A42A – A42J covering an area of approximately 8100 km². The two main towns located in the study area include Lephalale (Ellisras) and Vaalwater. Figure 66 shows the general location with Mokolo River as the major river system flowing through the study area eventually joining the Limpopo River to form the international boundary between South Africa and Botswana. The study is based on work previously undertaken by various investigators (Vipond, 1987 and 1988; Department of Water Affairs and Forestry, South Africa; 2008); VSA, 2009; Dennis, 2010; Vermeulen *et al.*, 2010; and Department of Water Affairs², 2010).

3.5.2 Topography and Geology of Mokolo

Downstream to the north, the area is relatively flat at an elevation of 600-700 mamsl (Figure 67) and the geology is characterised by the mafic and ultra-mafic rocks of the Bushveld Igneous Complex underlain by the metamorphic rocks of the Basement Complex. The central part of the area consists of the sandstone, conglomerate and carbonaceous shale of the Eccra Group of the Karoo Super-group and sandstone, greywacke and shale of the Waterberg Formation as shown on Figure 68. The elevation of the Waterberg Mountain range varies from 600 to 800 mamsl (Department of Water Affairs and Forestry, South Africa; 2008).

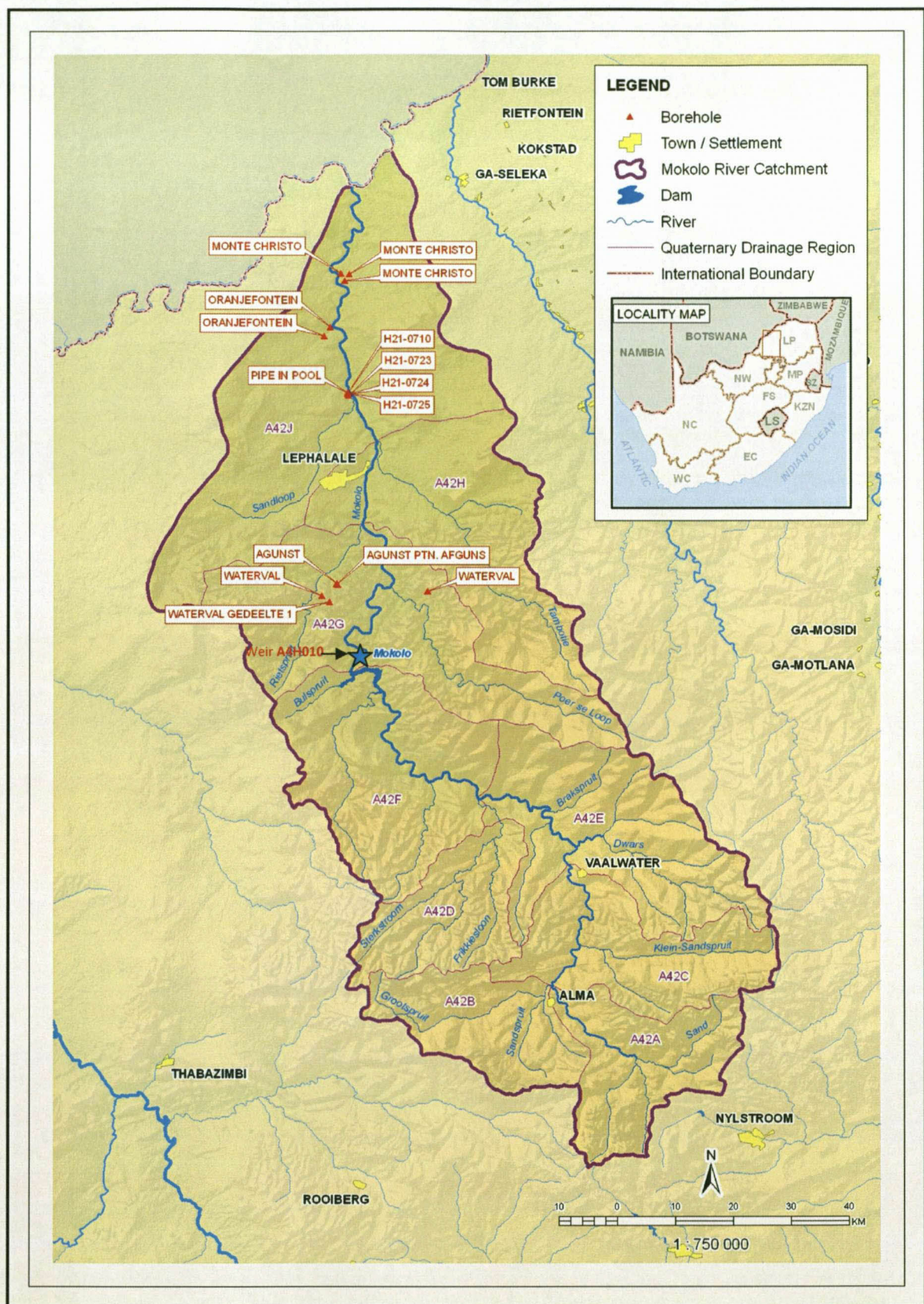


Figure 66: Location map of the Mokolo case study area

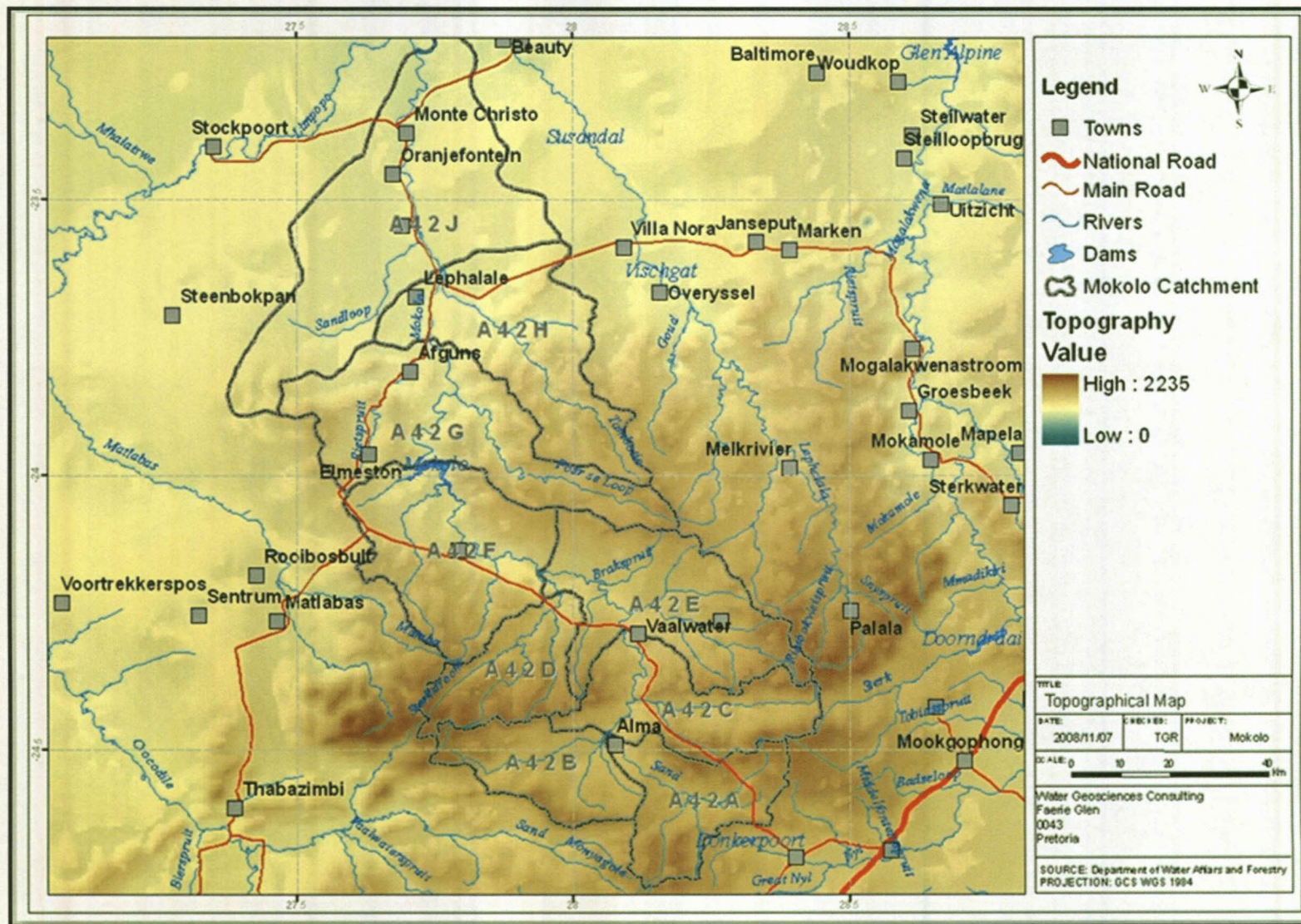


Figure 67: Topographic map of the Mokolo River System (Department of Water Affairs and Forestry, South Africa; 2008)

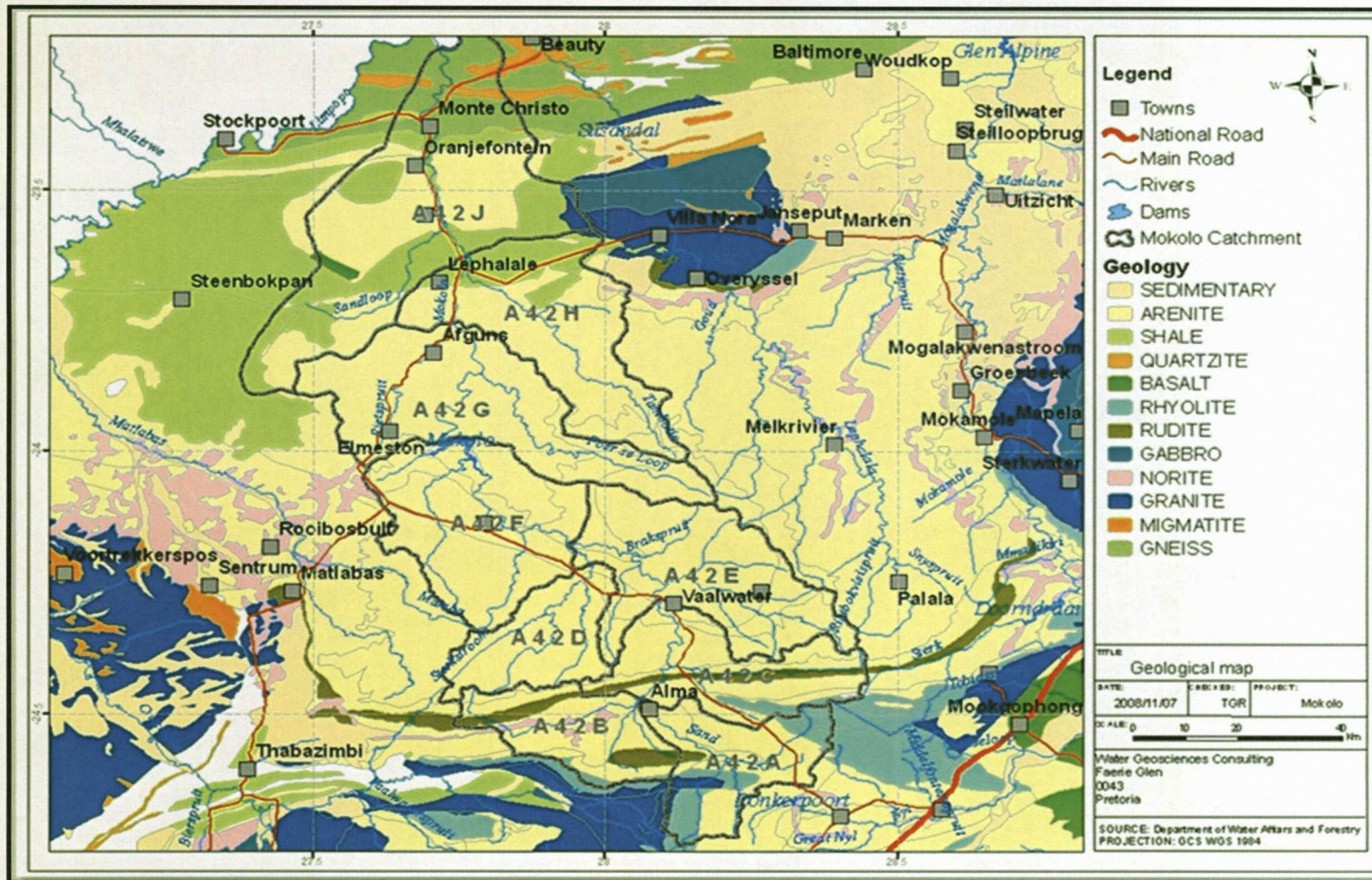


Figure 68: Geological map of the Mokolo River System (Department of Water Affairs and Forestry, South Africa; 2008)

3.5.3 Structural features

The Mokolo basin is extremely block faulted with a 250 m down-throw in the Eenzaamheid fault to the north as shown in Figure 69. Faults striking SE-NW, SW-NE and ENE orientated faults, subject to the north south tension release stresses, are targeted for the development of groundwater resources (VSA, 2009). The southern part consists of the Waterberg mountain range and rises to an elevation of up to 800 mamsl.

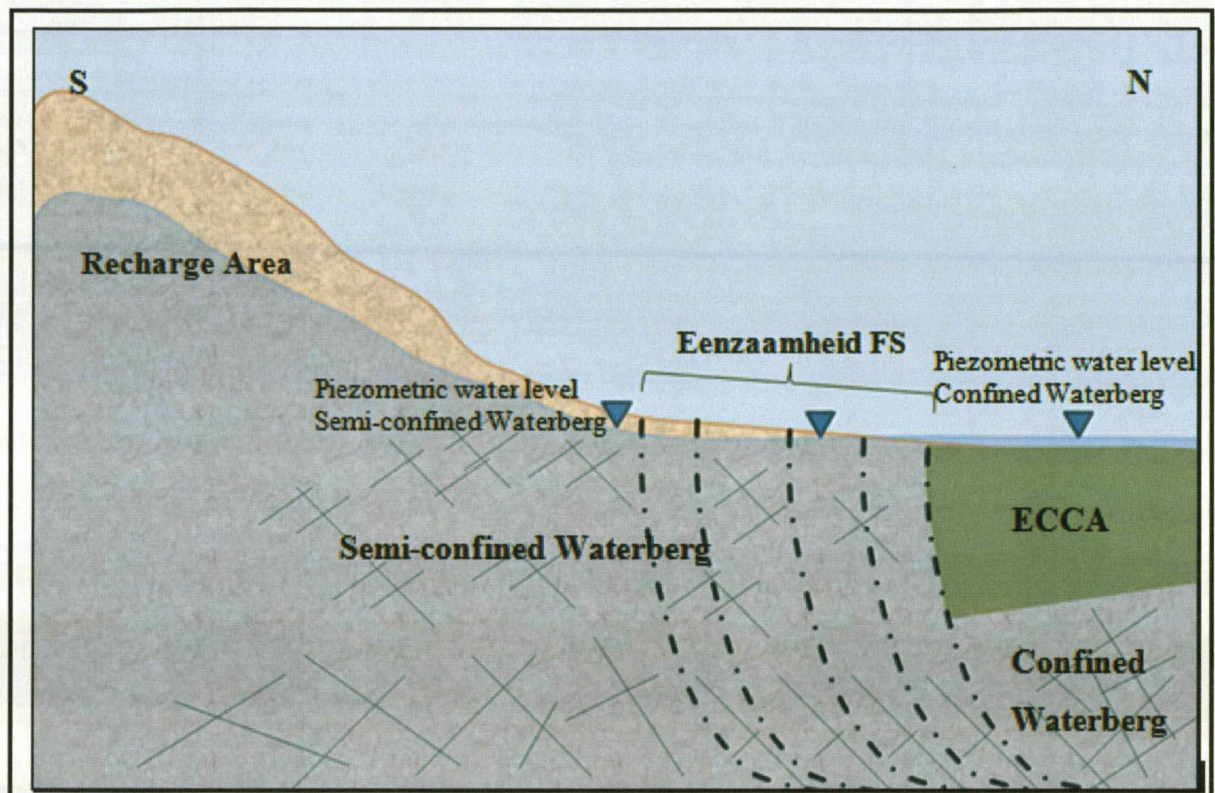


Figure 69: NS cross section of the Waterberg aquifer over Eenzaamheid fault (VSA, 2009 & Vermeulen *et al.*, 2010)

3.5.4 Climate and vegetation

Precipitation varies from 300 mm/year in the northeast to 700 mm/year in the southeast with an average of 555 mm/year (Figure 70). Evaporation rate, as shown on Figure 71 ranges between 1 600 mm/year in the south to 2 000 mm/year in the north.

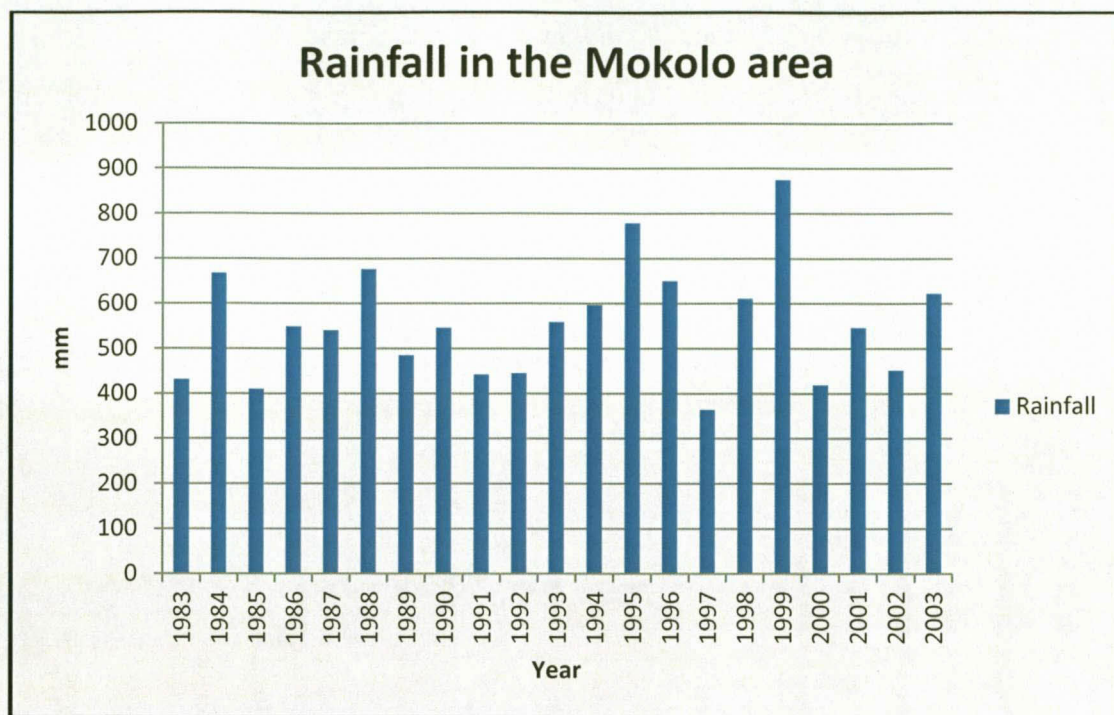


Figure 70: Rainfall in the Mokolo over a 20 year period (Source data – Water Geosciences Consulting, 2011)

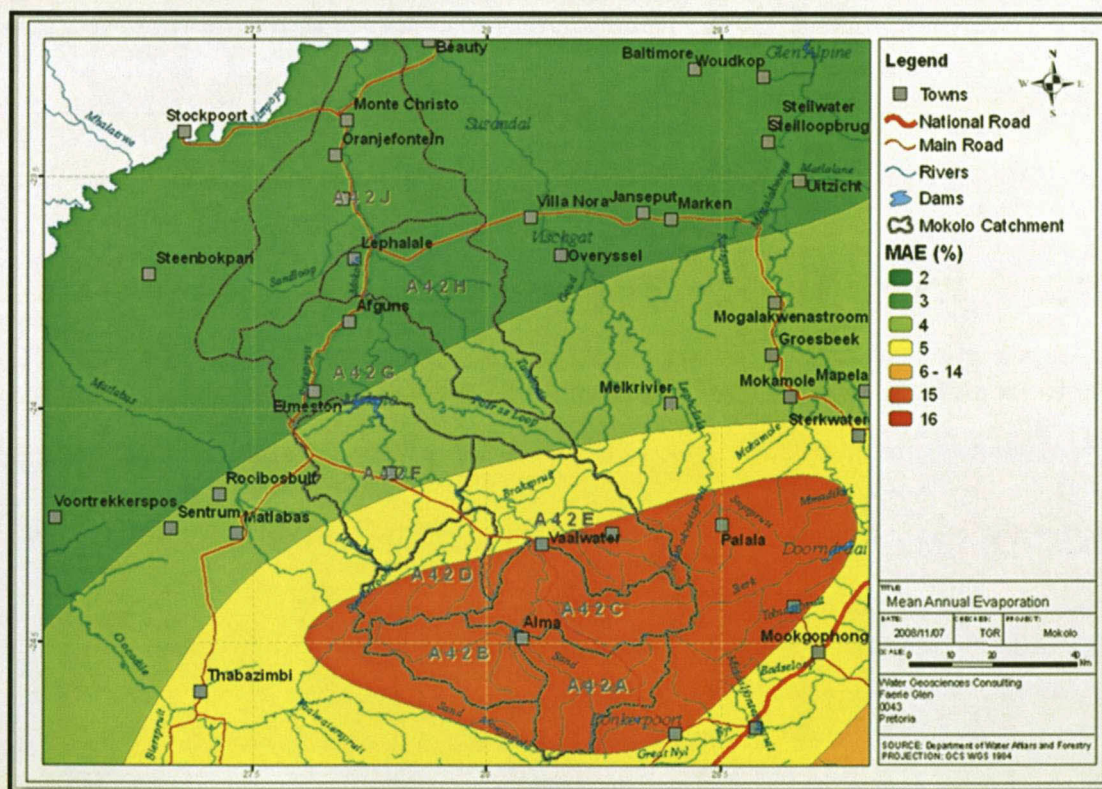


Figure 71: Mean annual evaporation for Mokolo catchment (Department of Water Affairs and Forestry, South Africa; 2008)

Figure 72 depicts the average rainfall and evaporation from 1962 to 1963. It can be deduced from the figure that evaporation rate is very high compared to rainfall in this area.

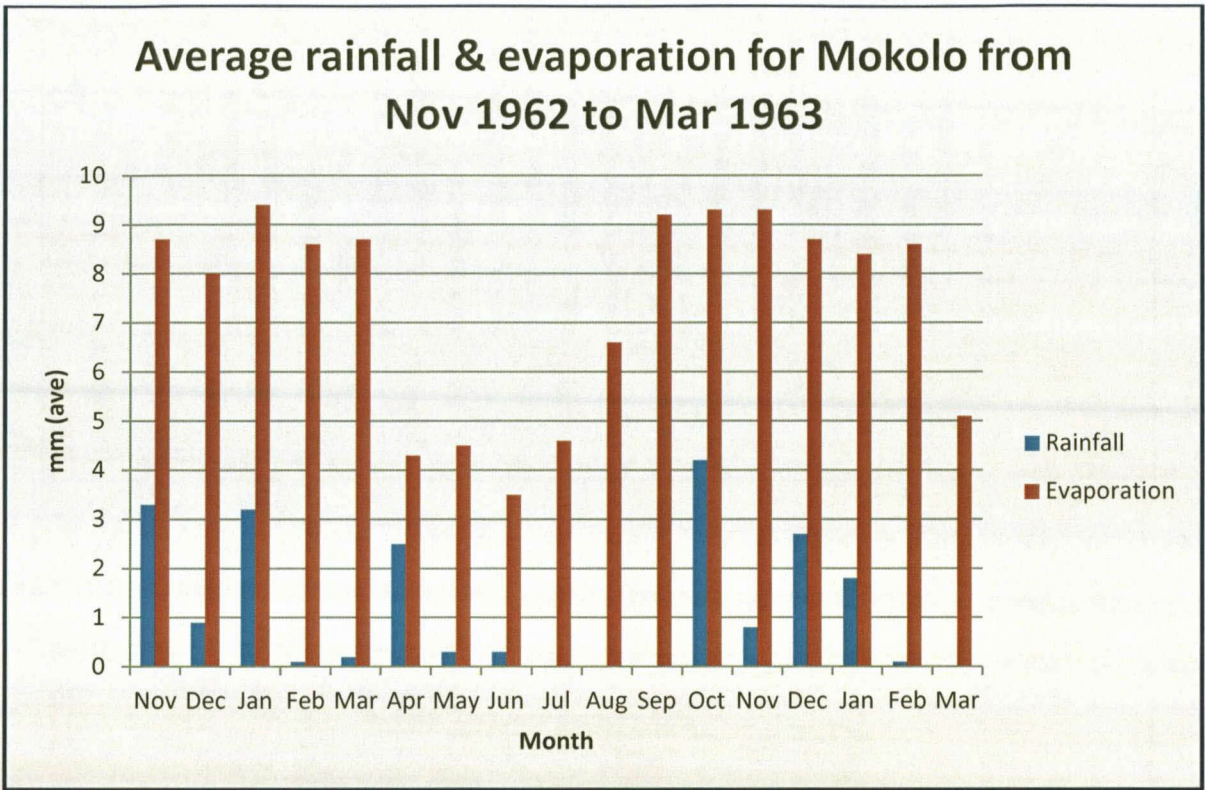


Figure 72: Average rainfall and evaporation in Mokolo catchment for a period of more than a year (Source data – Water Geosciences Consulting, 2011)

Annexure 8.6 depicts monthly rainfall for the Mokolo rainfall for a period of 12 years from October 1978 to September 2000.

The natural vegetation in the area mainly consisting of the Bushveld biomes is shown on Figure 73.

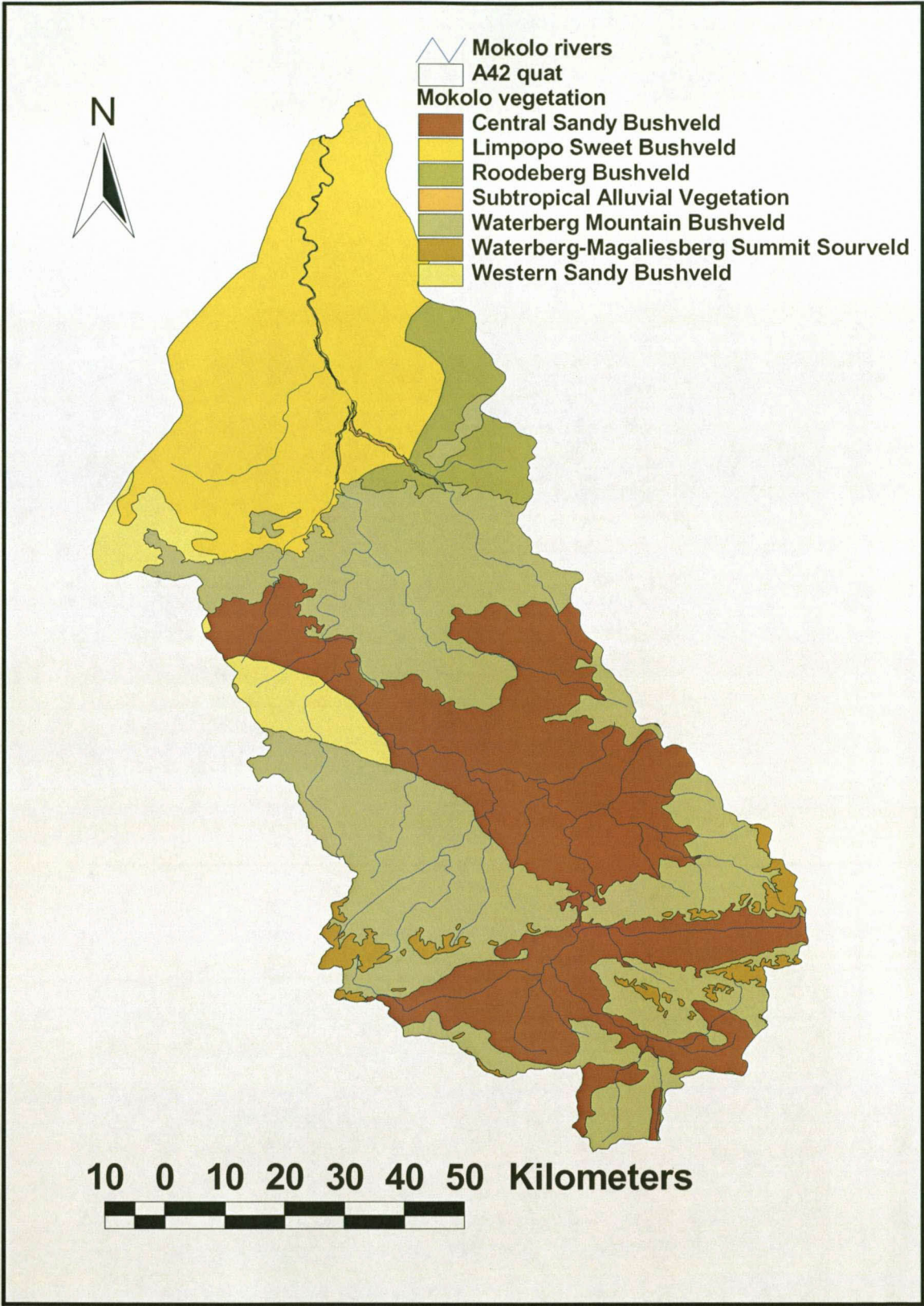


Figure 73: Vegetation types across the Mokolo Catchment (Dept of Water Affairs¹, South Africa, 2010)

The river is also characterised by riparian vegetation along side (Figure 74). Vegetation is expected to take up much of the water from the system including baseflow.

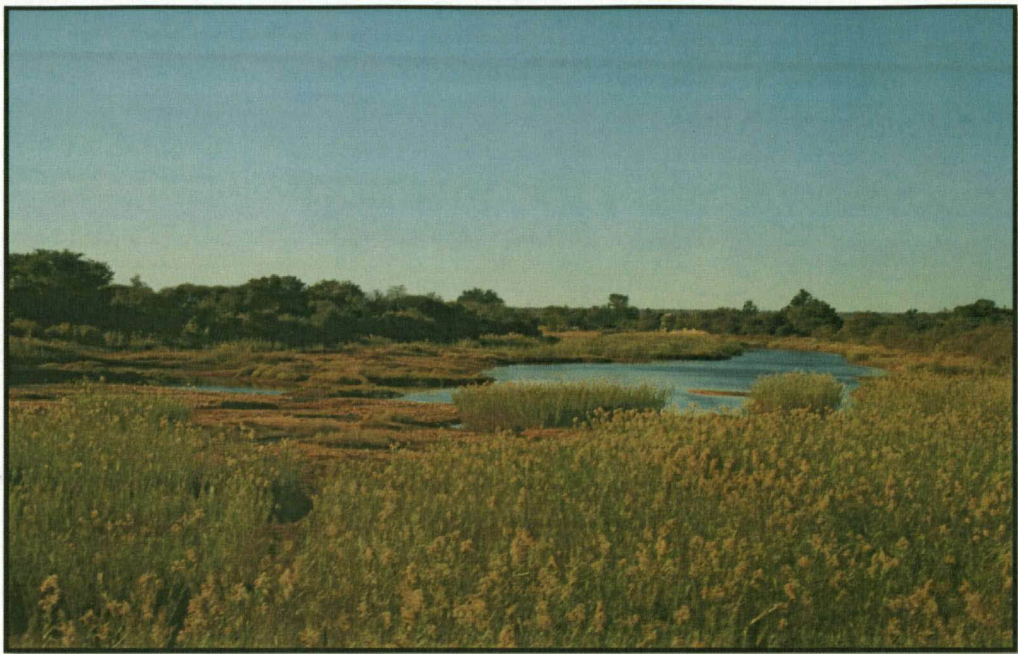


Figure 74: Mokolo River with riparian vegetation along side (picture by Lukas, 2012)

Water use in the area is mainly for agricultural purposes such as the tobacco field irrigated through borehole water in Figure 75.



Figure 75: Borehole in tobacco field along the Mokolo River (Dennis, 2010)

3.5.5 Aquifer types in the Mokolo Catchment

The Basement comprises of deeper fractured aquifers overlain by a weathered horizon of variable thickness. The significant aquifer within the basement rocks is within the ENE trending zones of shearing, faulting and brecciation that are typically covered by Quaternary deposits contributing to the aquifer's storage potential.

The Waterberg aquifer is predominantly of a fractured and weathered type potentially connected to alluvial deposits occurring along the Mokolo River, while the main groundwater targets are associated with fractured dyke contacts and fault zones. This formation is also associated with steep topography and constitutes recharge to the aquifer, often discharged on the steep slopes, and contributes baseflow/interflow to the rivers. A weathered zone aquifer is found only where deep weathering occurs and provides groundwater storage that feeds the underlying fractured aquifer.

The Karoo aquifer shows similar hydraulic properties as the Waterberg aquifer comprising fractured rocks with a porous matrix. However, groundwater resources and especially the development thereof, are limited due to the low recharge to these aquifers.

The Mokolo River is underlain by about 5 m thick alluvial deposits, with a coarse sand base that is recharged during periods of high stream-flows and discharge events (from the Mokolo dam) as well as during the rainfall season. It is an important local, major aquifer and exists in equilibrium with surface water, adjacent groundwater systems and ecosystems along the rivers.

3.5.6 Hydraulic properties of aquifers

Estimation of hydraulic properties was obtained by various researchers in this area and different values obtained for various reasons that is discussed here.

3.5.6.1 Hydraulic properties based on work by VSA (2009)

Recharge for the drainage area of quaternary catchments A42G, A42H and A42J where the study area is mainly located, occurs in the higher Waterberg Mountains.

Figure 76 shows a cross section of the alluvial and confined Waterberg underlying the Ecca Formation, north of the Eenzaamheid Fault, at the river. This fault was shown earlier in

Figure 69, under section 3.5.3. Based on differences in water levels of monitoring boreholes in the alluvial aquifer and the Waterberg aquifer it was established that these aquifers are unconfined and semi-confined respectively. Hence the alluvial aquifer in this area does not recharge the Waterberg aquifer, since hydrostatic pressure and the confining Eccca layers act as barrier to this, and that the semi-confined Waterberg aquifer discharges the weathered zone and the alluvial aquifer at the contact of the Eenzaamheid Fault with the alluvial aquifer. The alluvial aquifer is separated from the Waterberg aquifer by the impermeable Eccca Formation.

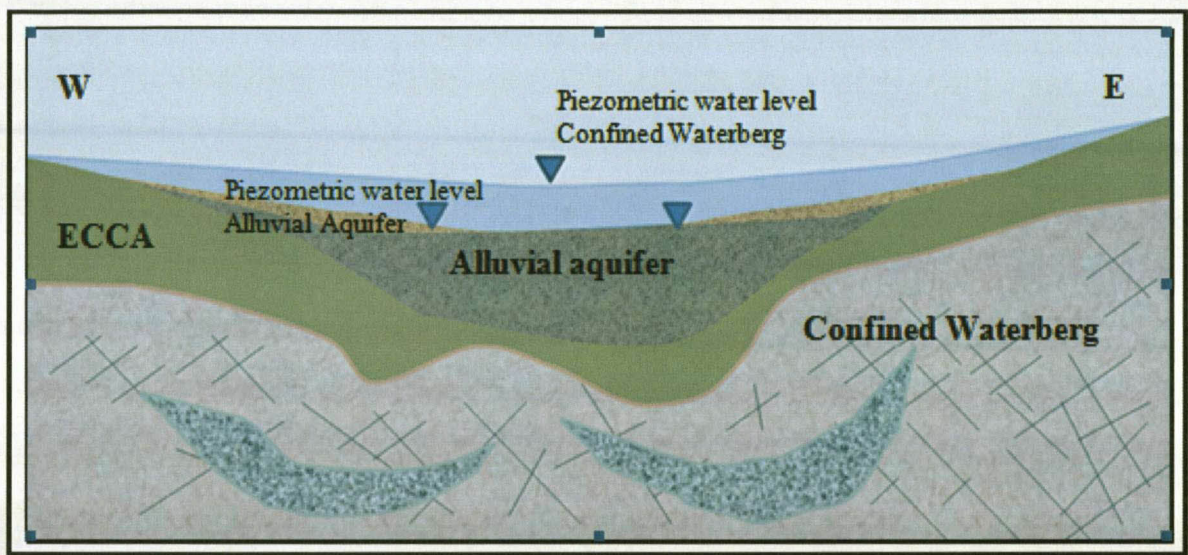


Figure 76: East-west cross section of the Alluvial and Waterberg aquifers north of the Eenzaamheid Fault (VSA, 2009)

Upon conducting geophysical surveys to locate structures, identify appropriate sites and subsequent drilling of 30 exploration boreholes, VSA (2009) analysed results of water strikes and blow yields to inform hydraulic testing of the aquifers. The blow yields in Waterberg aquifer ranged from 0.2 to 20 l/s while those drilled in the shallow aquifer ranged from 0.7 to 10 l/s. Of those drilled, 19 boreholes were selected for the 24 hour constant discharge pump tests and then Flow Characterisation (FC) method developed by the Institute for Groundwater Studies, in Bloemfontein used for analyses and estimation of aquifer properties. Table 9 depicts aquifer parameters (transmissivity as T and storativity as S).

Table 9: Constant discharge test results from the exploration boreholes in the Moloko Catchment study area (VSA, 2009)

Drill Target	Drill Site ID	Blow Yield (L/s)	24h Yield (L/s)	T _{early} (m ² /d)	T _{late} (m ² /d)	S
Semi-confined Waterberg Eenzaamheid Fault (S)	H21-0637	8	4	13.81	6.63	2.2 x 10 ⁻³
	H21-0638	15	7.5	34.7	26.8	2.2 x 10 ⁻³
	H21-0663	8	3	8.9	5.2	1.0 x 10 ⁻⁶
	H21-0665	>20	14	250	45	2.2 x 10 ⁻³
	H21-0669	0.8	0.12	1	0.5	2.2 x 10 ⁻³
	H21-0670	1.5	0.3	4.5	1.8	2.2 x 10 ⁻³
In between	H21-0700	12	9	13.53	10.54	1.7 x 10 ⁻³
Semi-confined Waterberg Eenzaamheid Fault (N)	H21-0666	8	3	61.9	11.2	8.2 x 10 ⁻⁶
	H21-0667	3.5	0.8	4.4	2.8	3.5 x 10 ⁻⁶
Semi-confined Waterberg	H21-0708	2.5	0.7	11.6	7.1	2.2 x 10 ⁻³
	H21-0712	8	2	3.4	3.33	1.17 x 10 ⁻³
Confined Waterberg	H21-0671	>20	9	85	18.6	9.9 x 10 ⁻⁶
Confined Waterberg underneath Alluvium	H21-0701	3.5	0.7	6.6	2.6	2.0 x 10 ⁻³
	H21-0702	4.2	2	3.3	2.8	2.2 x 10 ⁻⁵
	H21-0704	2.5	1.4	61.3	2.7	1.4 x 10 ⁻³
Alluvium	H21-0703	0.7	0.15	7	4.7	*8.8 x 10 ⁻⁴
	H21-0715	10	2	370.75	56.59	*2.2 x 10 ⁻³
	H21-0706	0.7	tested but found very weak – no evaluation done			
Confined Waterberg Daarby Fault	H21-0681	2.2	0.6	1.1	1.1	2.1 x 10 ⁻³

*For the unconfined alluvial aquifer S is measured as S_y or specific yield – the FC calculation is not considered sufficient for this type of aquifer.

It is clear from pump tests that, except for the alluvium, Transmissivity values are generally low.

However, earlier estimates of transmissivity values determined by previous investigators (Vipond, 1987 and 1988; Water Geoscience Consulting, 2008; and Department of Water Affairs², 2010) were comparatively higher. It turned out that although consideration of hydraulic properties in the latter case were with reference to various aquifers (i.e. Basement, Waterberg, Karoo and Alluvial) the focus was on the alluvial aquifer only, covering the area underlain by a particular hydrological unit. In other words, for instance, what was referred to as Basement aquifer was actually the alluvial aquifer underlain by the Basement Complex.

3.5.6.2 Hydraulic estimates by Vermeulen et al.,(2010)

Vermeulen et al.,(2010) obtained transmissivities for boreholes drilled in the area varying from 0.31 to 600 m²/d. The calculated harmonic mean for the transmissivity values for all the tested boreholes to be 0.4 m²/d. That the transmissivity values are low is markedly attributed to the geology of the area that is characterised by alternating layers of shale and mudstone interspersed with sandstone and coal. However, during a modelling exercise for dewatering purposes, more realistic values were found to range from 0.12 m²/d (for layered Formation) to 500 m²/d (for faults).

3.5.6.3 Summary of the water exchange among the river, the pool and the alluvial aquifer (Vipond, 1987 and 1988; Department of Water Affairs and Forestry, South Africa; 2008; and Department of Water Affairs², 2010)

During the period between October and November 1987 a total volume of 7.4 Mm³ was released from the Mokolo Dam (Figure 77) and the water level response was monitored in piezometers installed in the alluvium along the entire channel length (Vipond, 1987 & 88; and Department of Water Affairs², 2010).

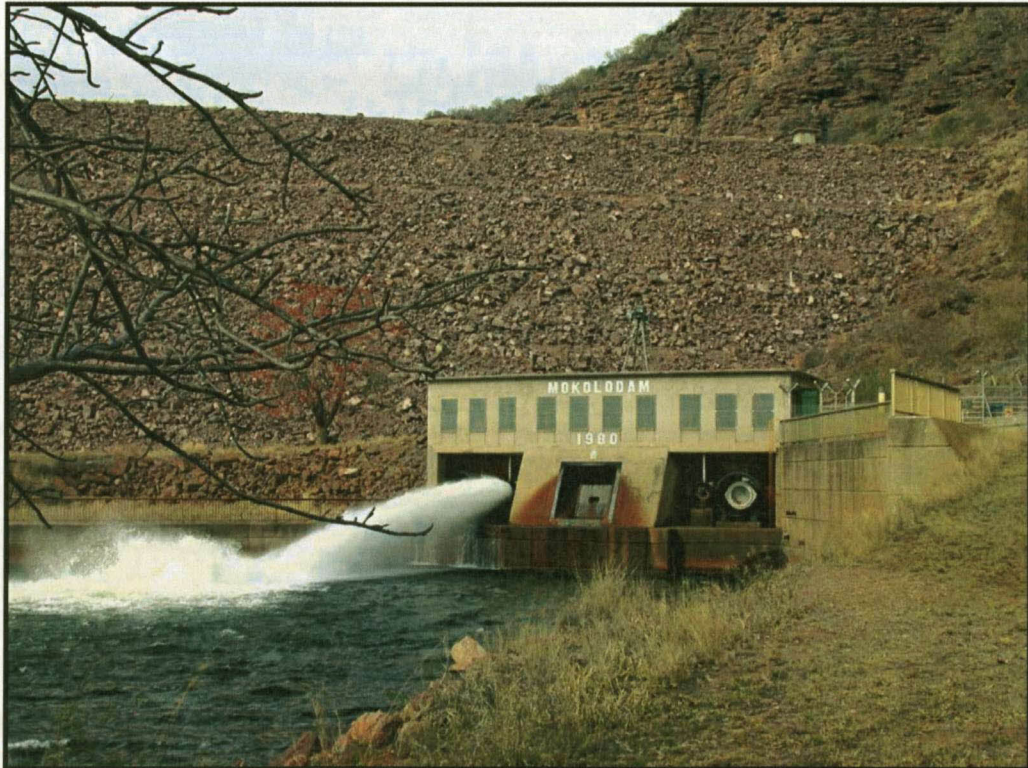


Figure 77: Water release at Mokolo dam (Vermeulen, 2012)

Observations from these earlier investigations were that:

- Transmissivity values of the alluvial sands overlying the Basement Complex, Karoo and Waterberg Formation ranged from 591 to 1835 m²/d as indicated in Table 10. However, it is very strange that T values upstream are higher than downstream.

Table 10: Transmissivity values as determined by Vpond (1987) and Dept of Water Affairs² (2010)

Aquifer type alluvial	Vpond (1987) T values	Dept of Water Affairs ² (2010) T-values
Waterberg flat mountains	1835	1489
Waterberg flat area	1835	1489
Karoo (Ecca)	865	999
Basement Complex	591	810

- Through-flow indicated by recession plots (Figure 78), occurs at uniform rate throughout most of the river profile.

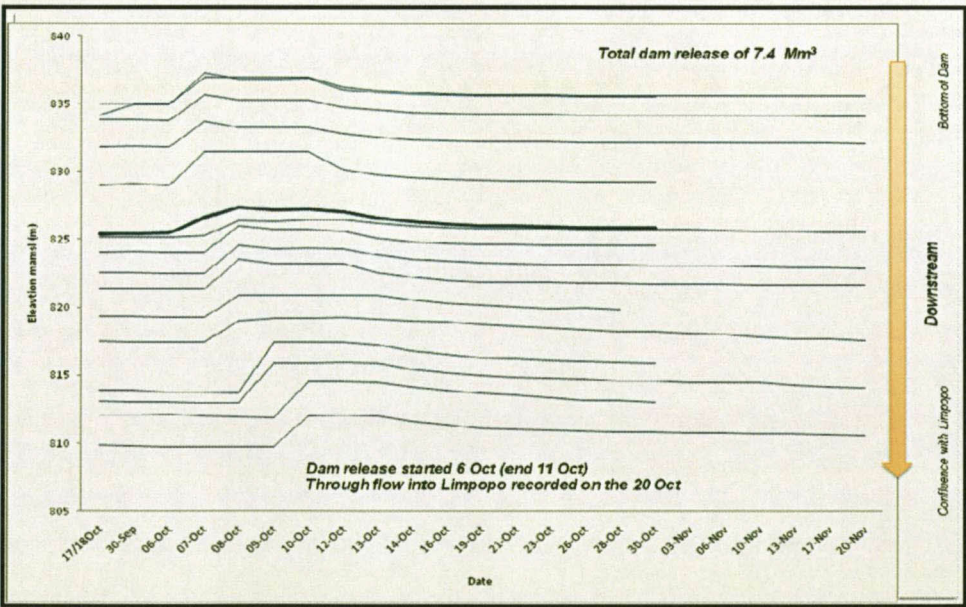


Figure 78: Comparison of water level recessions along the Mokolo River following the dam release (7.4 Mm³/a) during October 1987 (Vipond, 1988 and Department of Water Affairs², 2010)

Figure 79 represents groundwater flow contours based on groundwater levels measured in boreholes close to the Shot Belt Pool. Clearly the flow is generally away from the Shot Belt Pool westwards and eastwards towards the alluvial aquifer and Mokolo River. The groundwater flow vectors indicate influent conditions with respect to the Shot Belt Pool and Mokolo River. That is both the river and the pool lose water to the alluvial aquifer.

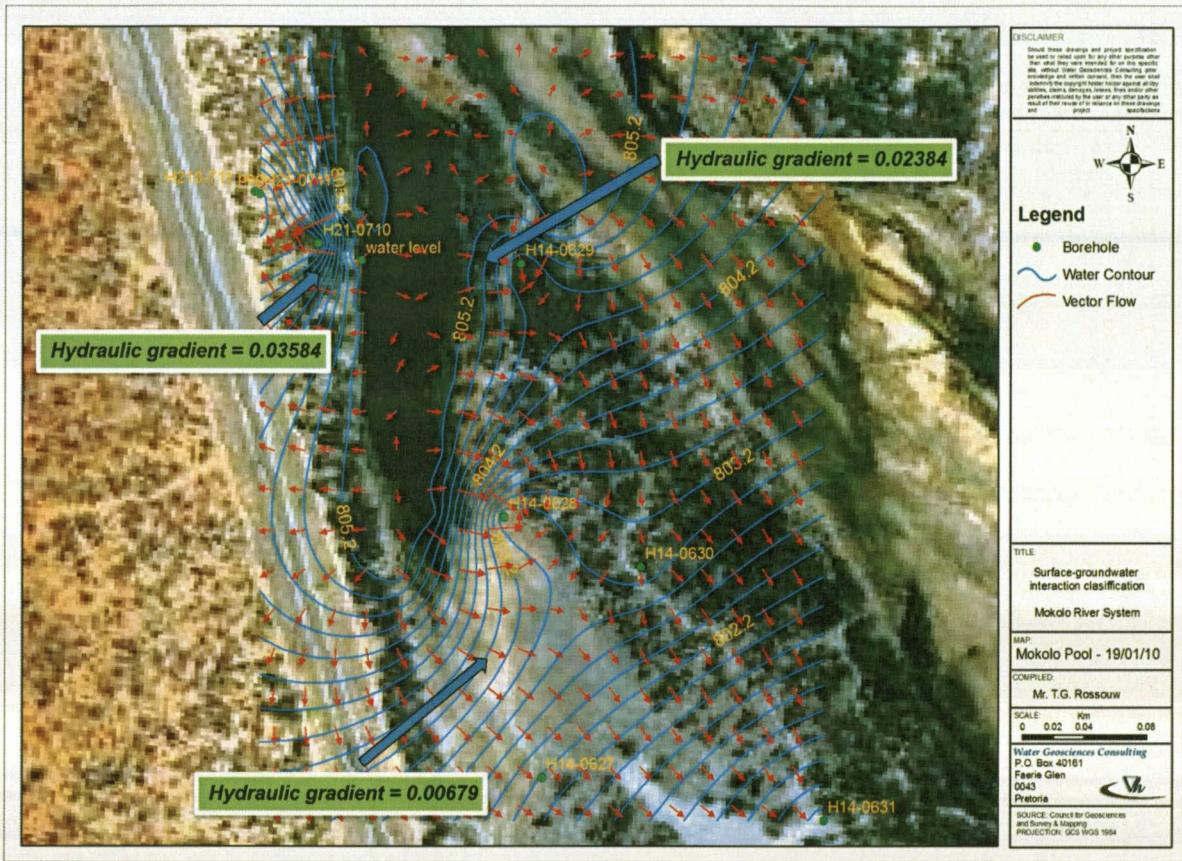


Figure 79: Groundwater flow vectors indicating flow in the vicinity of the Shot Belt Pool (Department of WaterAffairs², South Africa, 2010)

The water level in the Shot Belt Pool is also higher than in the boreholes around and in the vicinity of the pool (i.e. H21-0710, H14-0627, H14-0728, etc) as indicated in Figure 80. This confirms that the water may be lost to the aquifer given the relative water stage in the aquifer and the pool.

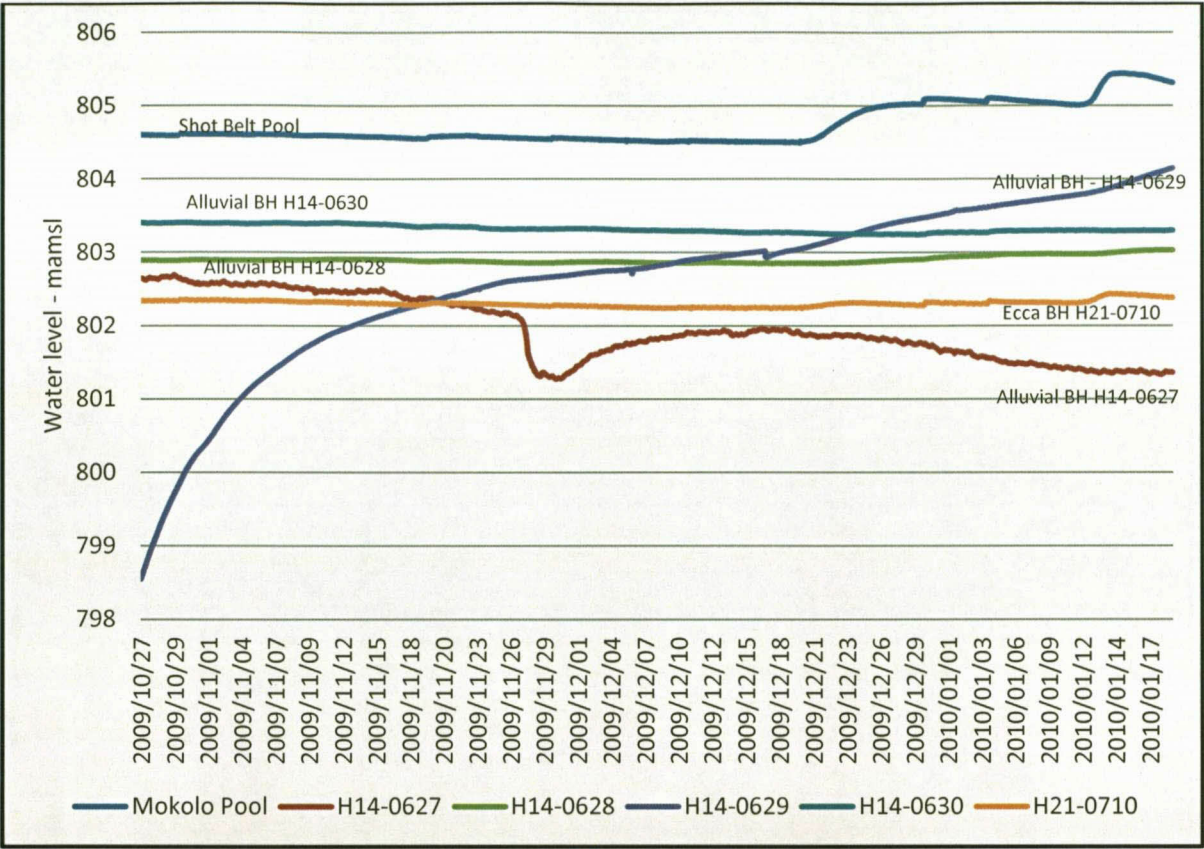


Figure 80: Variation in pool and groundwater level with time (after Dept of Water Affairs², 2010)

The fluctuation in water level (Figure 81) for the alluvial aquifer (borehole H21-0703) is similar to that of the river (weir A4H010). Figure 66 indicates the location of the weir. This may indicate that there is leakage between the two systems.

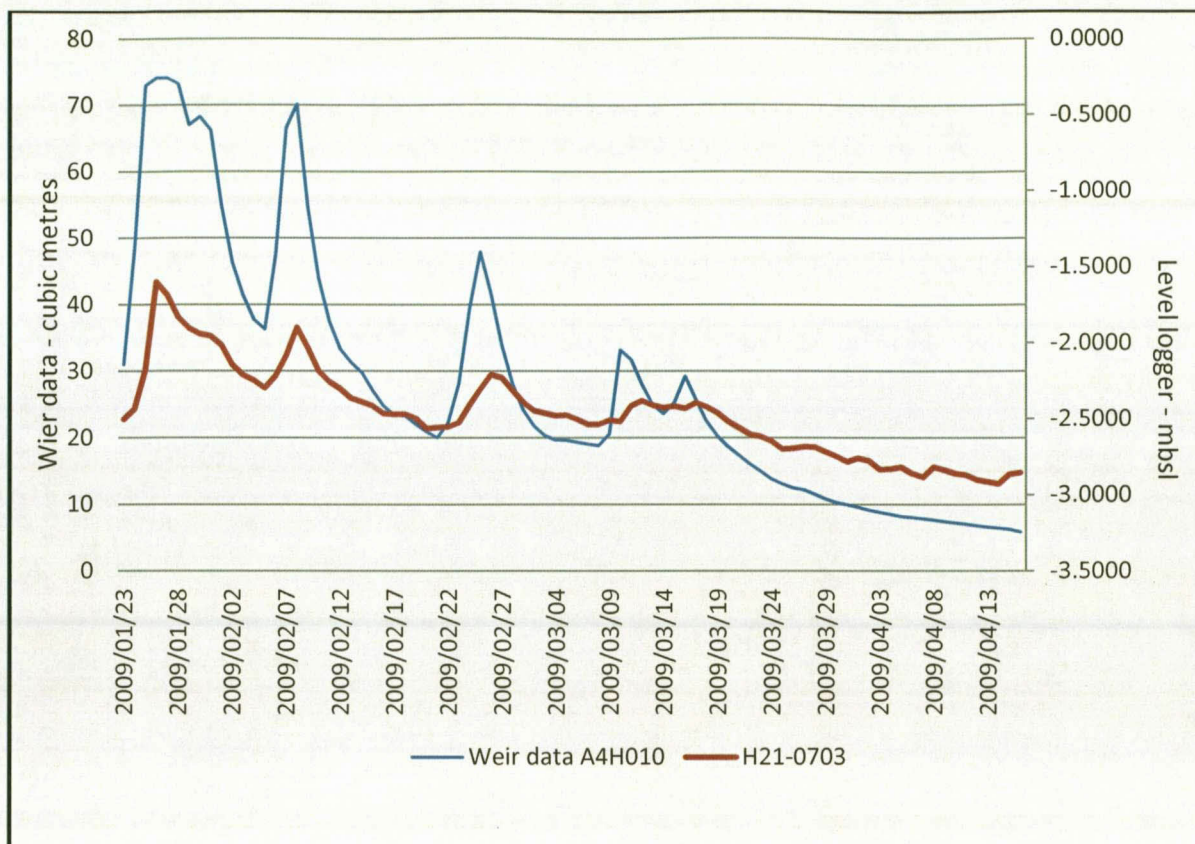


Figure 81: Water level fluctuation in the alluvial borehole (H21-0703) and weir data at observation point (A4H010) by Dept of Water Affairs², (2010)

The water level measurement at the weir correlates to rainfall events as shown on Figure 82. This is due to a spill from the dam due to precipitation and dam releases upstream.

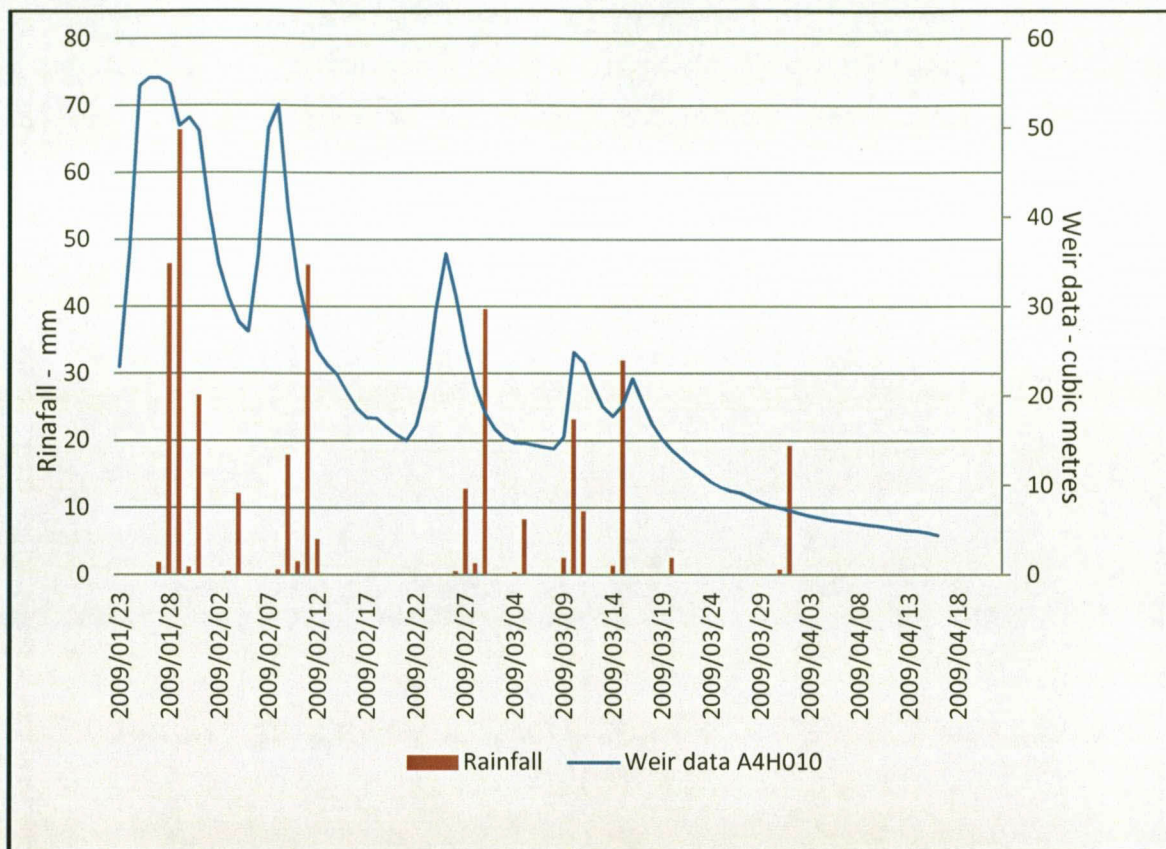


Figure 82: Rainfall and flow at the weir (Dept of Water Affairs², 2010)

3.5.6.4 Chemistry analysis for the pool, the river and boreholes in the vicinity

The chemistry analyses for the Shotbelt Pool, Mokolo River and H21-0710, H14-0627 & A14-0629, as indicated in Figure 83 shows that H21-0710 is Na rich C/ type water in the typical deep groundwater area, which is not surprising since this borehole was drilled in Eccca group of fractured rocks. On the other hand, other groundwater samples (H14-0627 and H14-0629) plot in the same area as river water and pool water which are all HCO₃ type water. The similarity in chemistry shows that water from these boreholes that are drilled in the alluvial have probably mixed with pool and surface water.

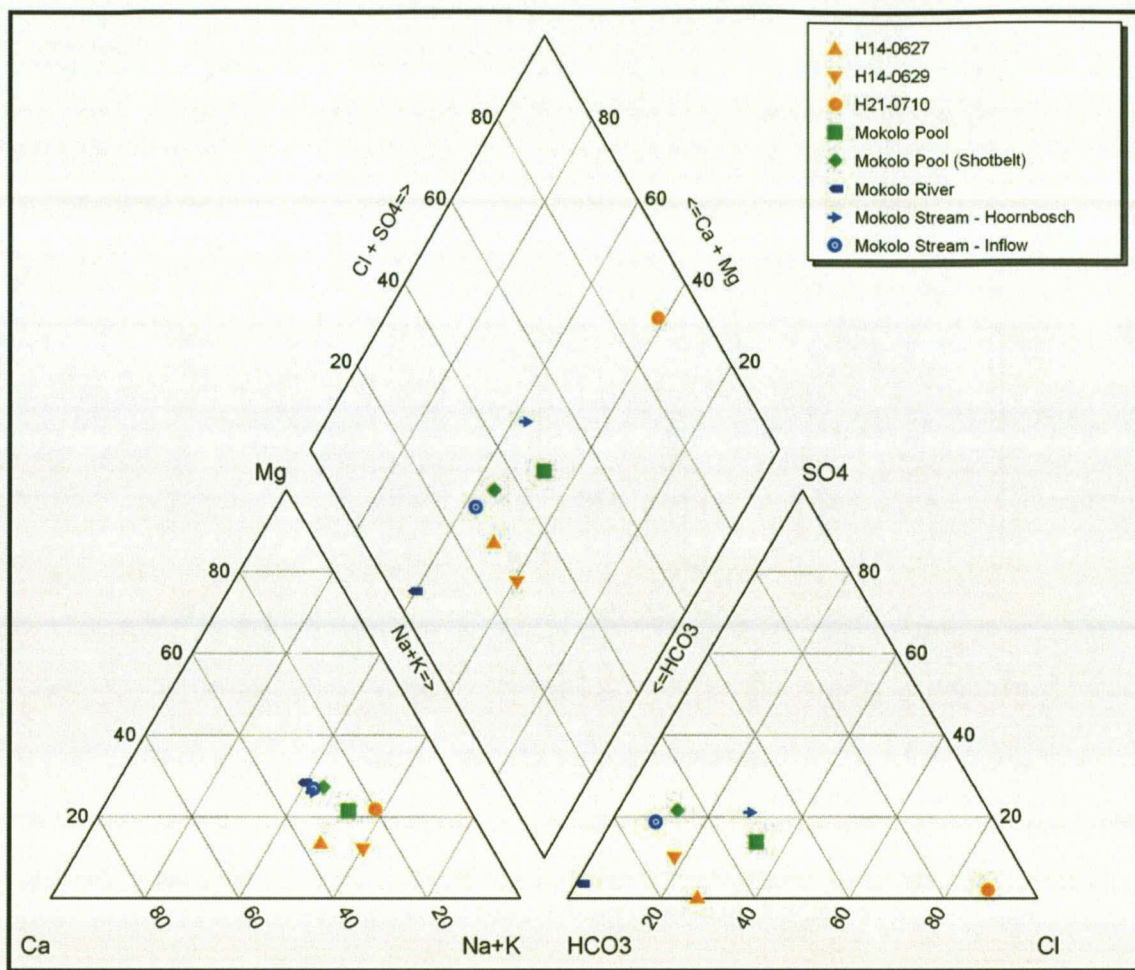


Figure 83: Piper diagram for river, pool and groundwater in Mokolo (Source data – Water Geosciences Consulting, 2011)

3.5.6.4.1 Stiff Diagram

It is clear from the Stiff Diagram (Figure 84; Mogol is the earlier name for Mokolo) that borehole H21-0703 and H21-0715 (that is also located in the alluvial) have chemical signature that is quite similar to that of the Mokolo River (named Mogol River in the Figure). This confirms that there must have been water flux exchange between the river and the alluvial aquifer in which the boreholes are located.

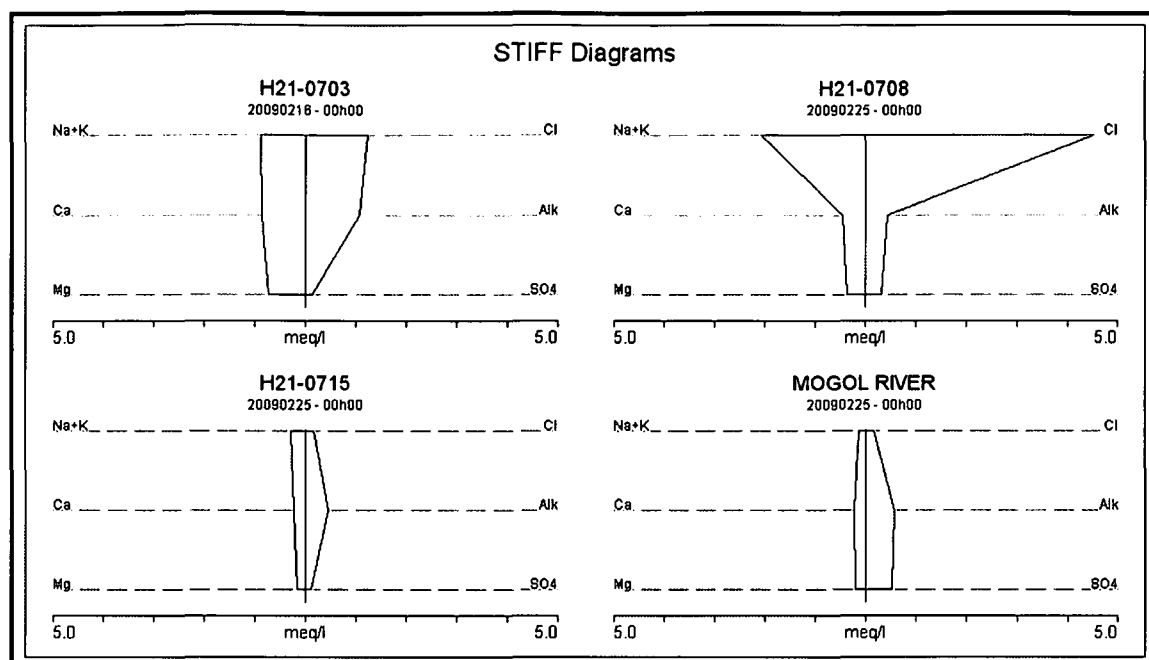


Figure 84: Stiff Diagram for Mokolo (Source data - Water Geosciences Consulting, 2011)

3.5.6.4.2 Isotope hydrology

Boreholes H21-0703 is highly enriched in deuterium isotope signature whereas typically surface water that is often subject to high evaporation rate is expected to be comparatively D enriched rather than groundwater (Figure 85). Hence this yet again confirms interaction between surface river water and the groundwater (alluvial aquifer). Both the Mokolo River (Mogol River) sample and the groundwater samples (H21-0703 as well as H21-0706) are on the evaporation line (Figure 85) in relation to the GMW line.

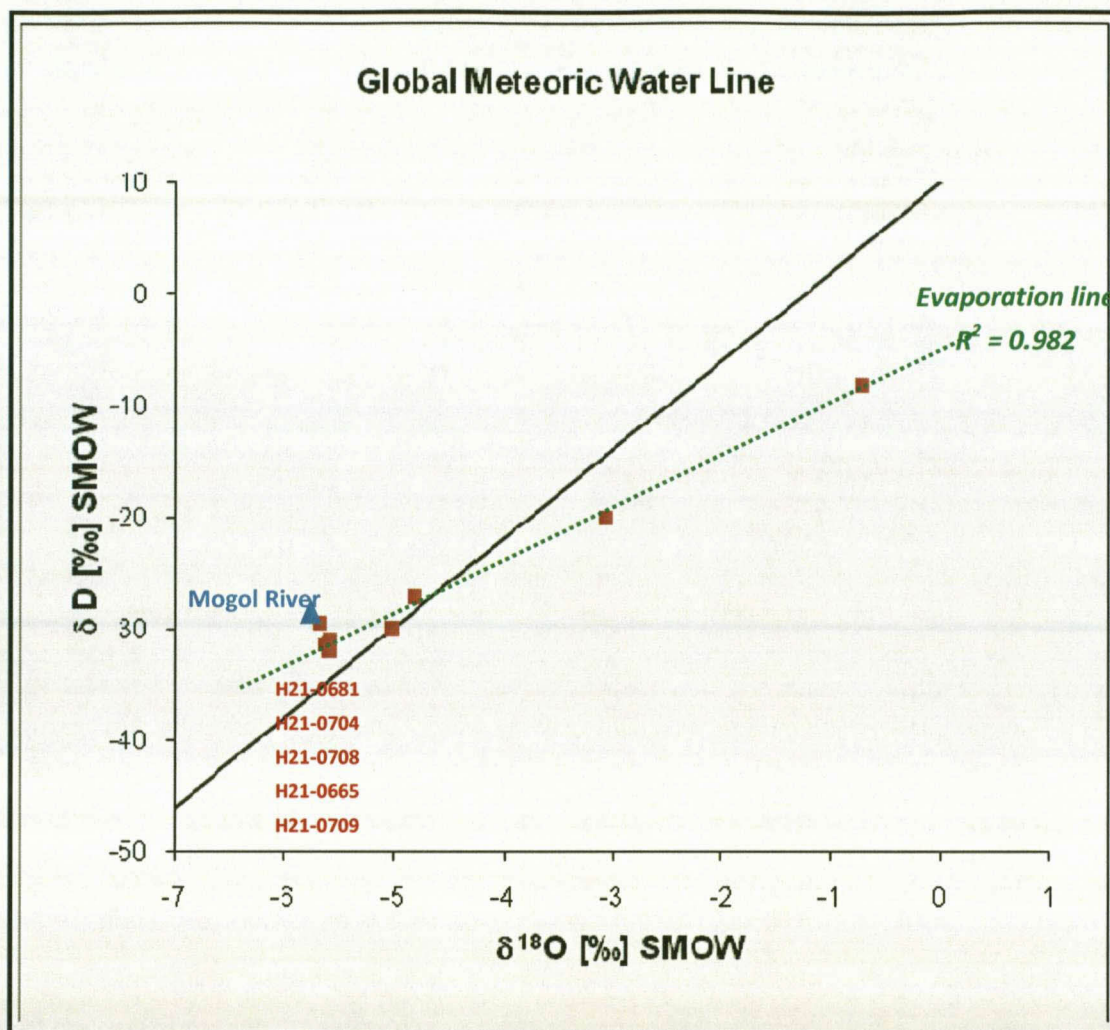


Figure 85: Stable isotope data in relation to the Global Meteoric Water Line (VSA, 2009)

3.5.6.5 Groundwater surface water interaction and baseflow estimates

Groundwater contribution to baseflow is the component that derives from the aquifer adjacent to a surface water body. From the published values (Table 11), the groundwater contribution to baseflow is mostly equal throughout the whole area except for quaternary catchment A42J, which is dominated by flat lying topography underlain by Karoo and Basement rocks with much smaller runoff volumes.

Table 11: Baseflow estimates by Schulz, Pitman and Hughes (Department of Water Affairs², 2010)

Resource unit	MAR (Mm ³ /a)	Baseflow [Schultz] (mm/a)	Baseflow [Pitman] (mm/a)	Baseflow [Hughes] (mm/a)	GW contribution to baseflow [Sami] (mm/a)
A42G	39.97	0.55	4.4	9.39	0.55
A42H	27.74	0.51	4.1	7.5	0.51
A42J	7.42	0	0	0.05	0

As part of the groundwater assessment of the Mokolo catchment Dennis (2010) delineated the area into groundwater response units (Figure 86) according to the following criteria:

- The geology and associated aquifer types are the same as those mentioned in sections 3.5.2 and 3.5.5
- Topography and associated aquifer recharge (as percentage of mean annual rainfall)
- The potential of surface water - groundwater interaction.

Details on response units investigated are summarised on Table 12. The conductivity values based on pump tests are recommended (section 3.5.6).

Table 12: Response units of the study area (adapted from study by Dennis, 2010)

Unit	Formation	Recharge	Land use	Depth to groundwater	Alluvial
A	Waterberg – mountainous	High (10 -19 mm/a)	Not much	-	Not much
B	Waterberg - mountainous	Moderate (5-9 mm/a)	Intense agriculture	20-25 mbgl	15 m thick alluvial
C	Waterberg - foothill	Average (4-5 mm/a)	Intense agriculture	20-25 mbgl	12 m thick alluvial
D	Waterberg - foothill	Average (4 mm/a)	Intense agriculture	20-25 mbgl	12 m thick alluvial
E	Karoo & basement – low land	About (3 -6 mm/a)		20-0 mbgl	8 m thick alluvial

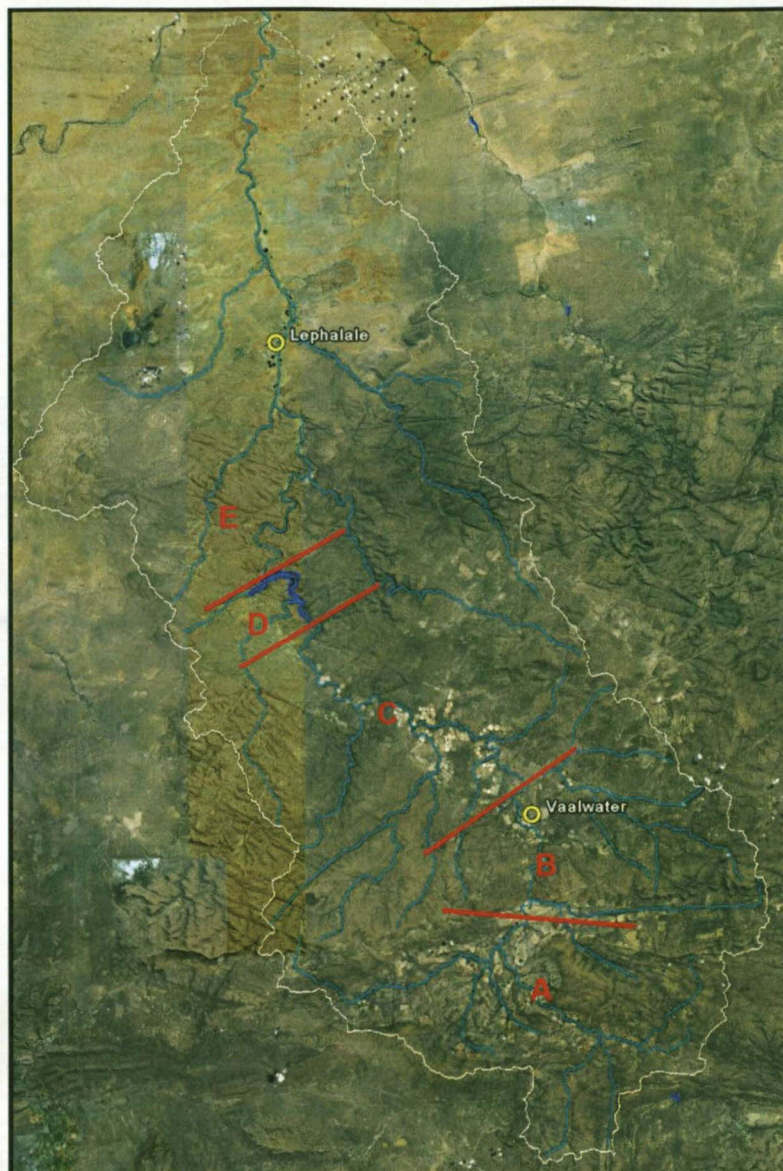


Figure 86: Groundwater response units (Dennis, 2010)

Detailed biophysical studies were undertaken at 5 sites selected on the basis of the response units and further observations. Figure 87 shows the location of sites while Table 13 summarises baseflow values or groundwater contribution to surface water estimated by using Herold's separation hydrograph method (Dennis, 2010). Depth to groundwater, average blow yield and recharge values are also included in the table. The difference in recharge up the mountains (at 80 Mm/a) compared to (3.3 Mm/a) the riparian zone is huge.

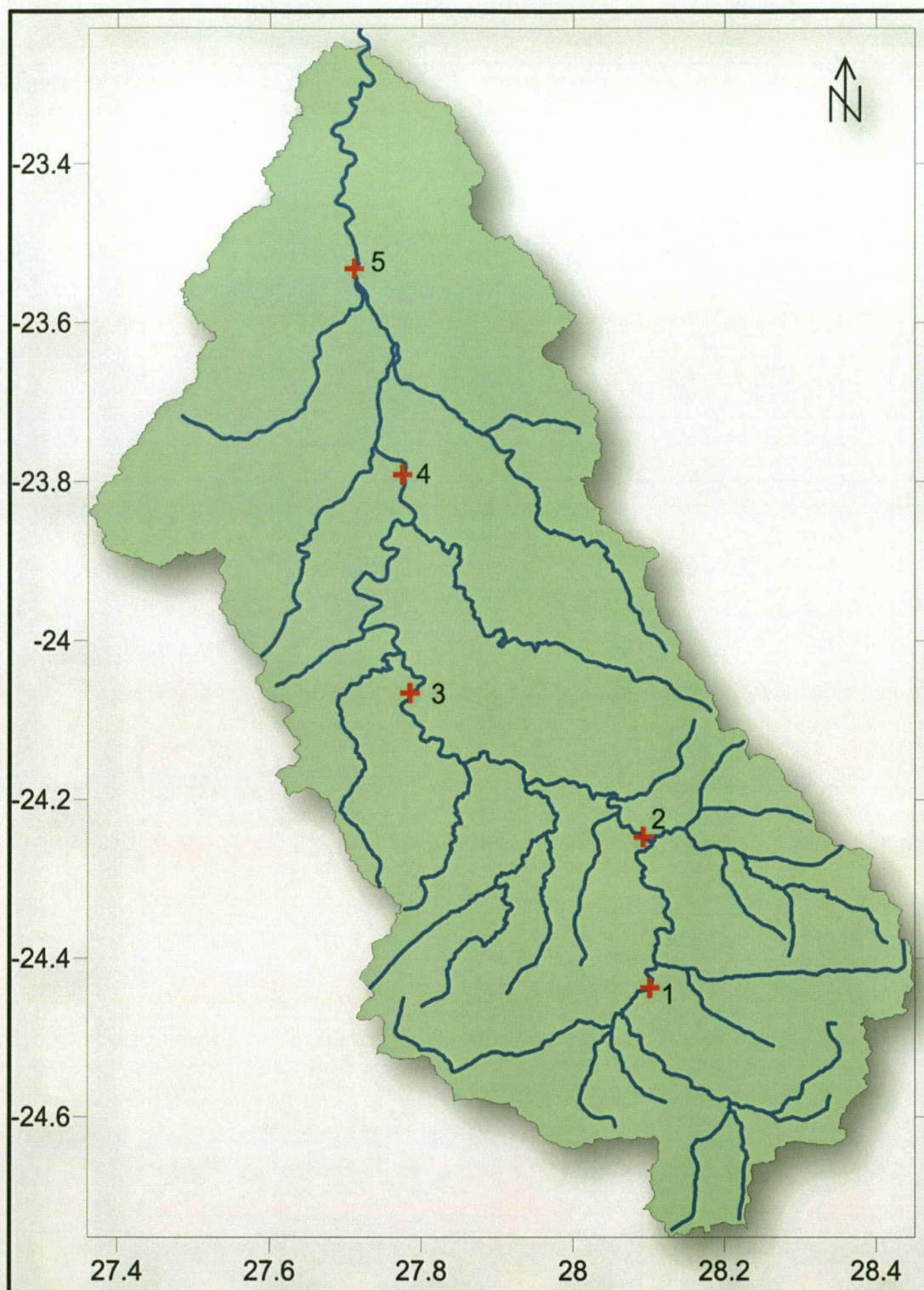


Figure 87: Location of representative sites for groundwater response units (Dennis, 2010)

Table 13: Groundwater contribution to surface water in Mokolo Catchment (Dennis, 2010)

Site	Depth to Groundwater (m)	Average blow yield (l/s)	Recharge (Mm/a)	Groundwater contribution to surface water (Mm ³ /a)
1	12	>3	80	4.8
2	4	>5	4.5	2.5
3	4	>5	6	1
4	4	>3	3.2	1
5	>10	>3	3.2	0.5

3.5.6.6 Surface water - groundwater interaction assessment using classification system

A two tiered classification scheme comprising of the geological classification of the river setting and a hydraulic classification of the surface-groundwater interaction developed by Dennis and Witthueser (2007) was proposed (Figure 88 and 89). That tiered classification system enabled investigators to recommend appropriate methods that are applicable to evaluation of surface water - groundwater interaction. In this particular case the system indicates that, given the conceptual model of the aquifer and the river as well as the intermittent exchange of water between the alluvial aquifer and the river, either the analytical or numerical model is applicable.

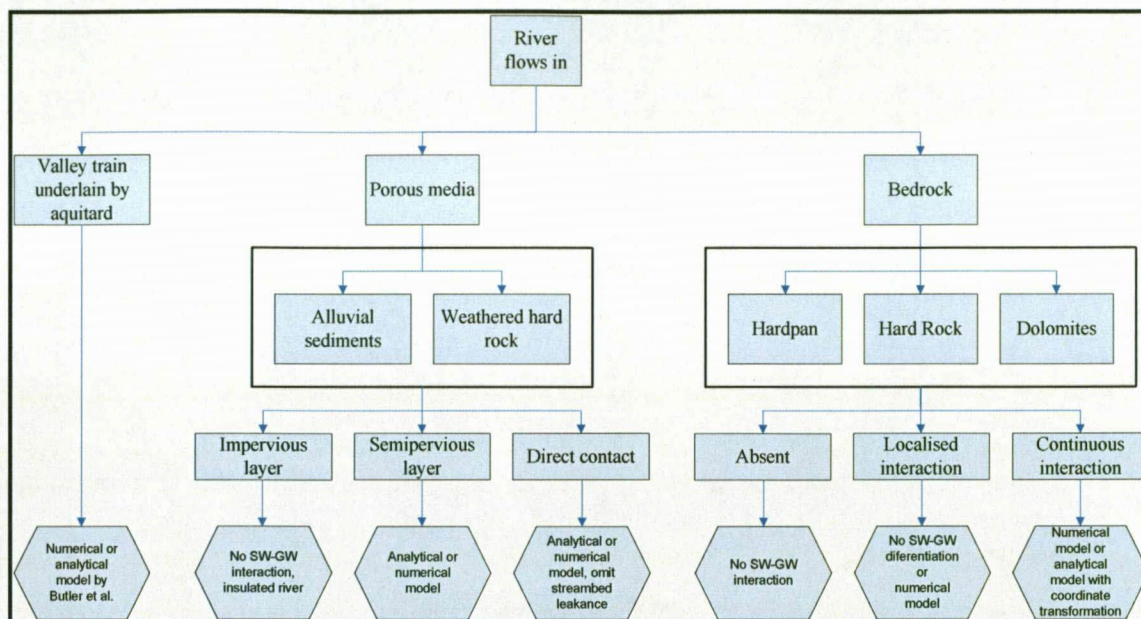


Figure 88: Primary geological classification scheme for surface water - groundwater interaction assessment
(Dennis and Witthueser, 2007)

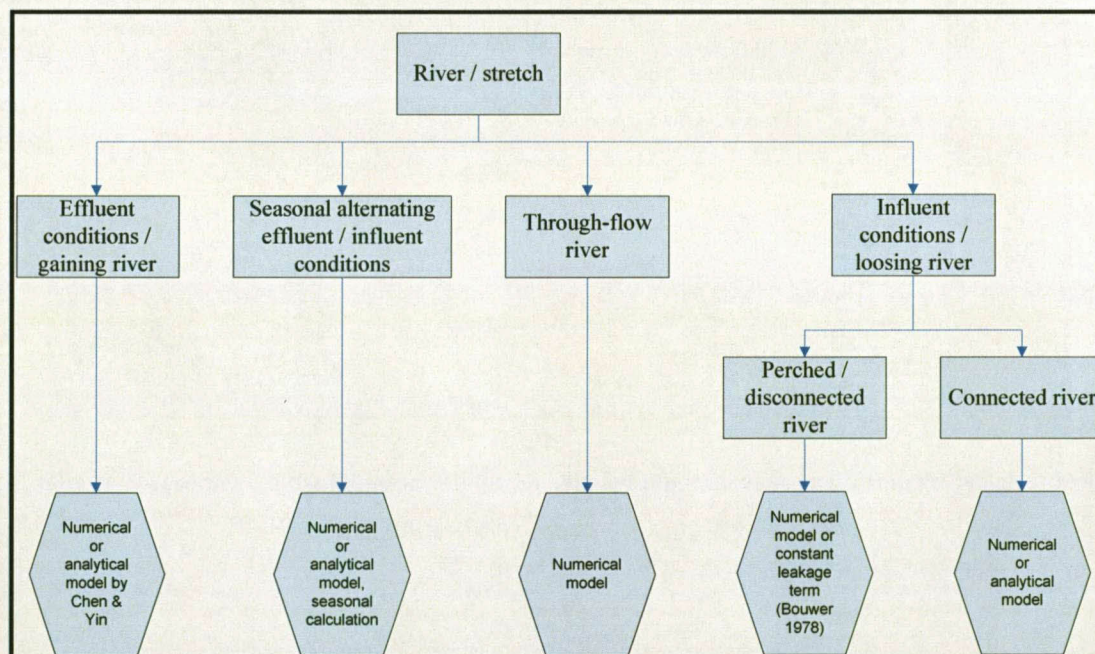


Figure 89: Secondary hydraulic classification scheme for surface water - groundwater interaction assessment
(Dennis and Witthueser, 2007)

Out of the 7 boreholes drilled in the vicinity of the Shot Belt Pool site, 2 were drilled into rocks of the Ecca Group to depths of 50 metres and 100 metres respectively while the rest were drilled in the alluvial aquifer system to depths of 25 metres. A typical borehole log is as indicated in Figure 90. The yields in drilled boreholes were very low and pump testing was consequently unsuccessful. Hence aquifer characteristics previously calculated were obtained from the literature.

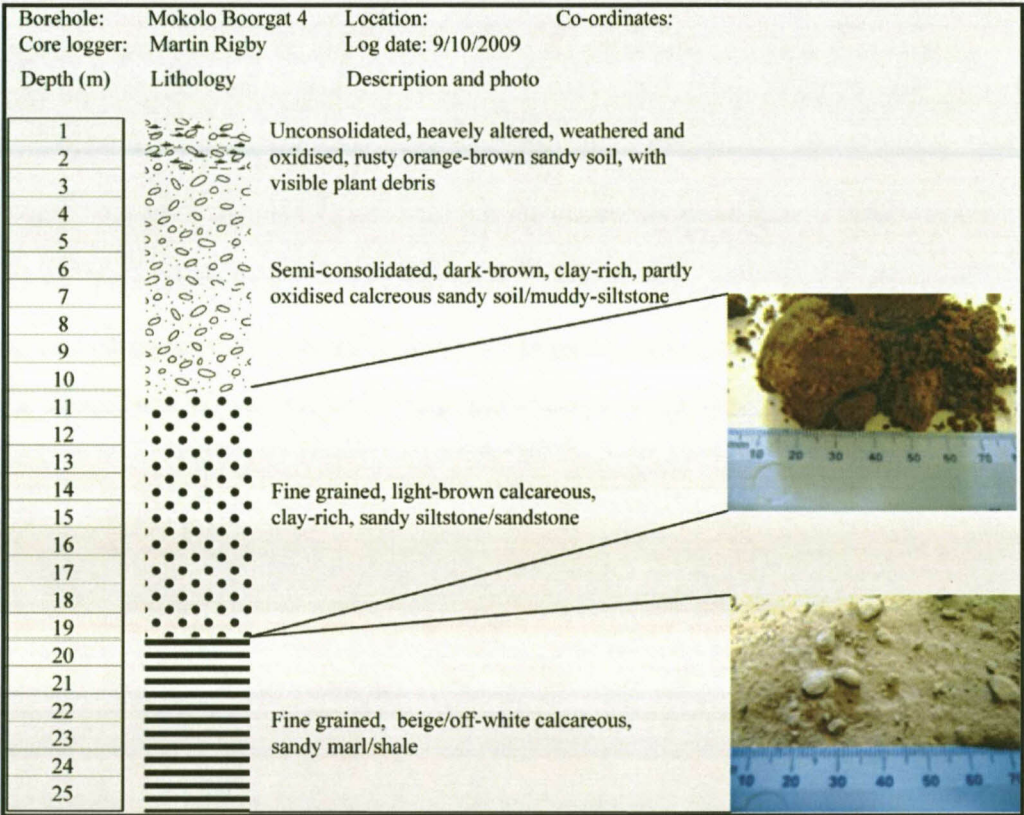


Figure 90: Typical log of the borehole drilled n the alluvial aquifer of the Mokolo River System (Department of WaterAffairs², South Africa, 2010)

3.5.7 Conceptual model of the Mokolo

The alluvial aquifer was then subdivided into 4 categories based on geology and aquifer properties determined by Vipond (1988). It should be noted that the following sections are conceptual, and not taken from a specific place in the area, and based on the knowledge of the geology in (A) Waterberg Formation (mountainous area); (B) Waterberg Formation

(flatter area); (C) Alluvial aquifer (underlain by Basement rocks) and (D) Alluvial aquifer (underlain by the Eccca Group). In summary the subdivisions were as follows:

- Section A (Figure 91) is characterized by a mountainous region underlain by sediments of the Waterberg Formation with deep eroded valleys and cliffs. The average depth of the alluvial sands in the Mokolo River is 12.4 m with a very high Transmissivity (T). The hydraulic gradient towards the river bed from the surrounding Waterberg Formation is 0.00073 indicating a steeper gradient compared to the other sections.

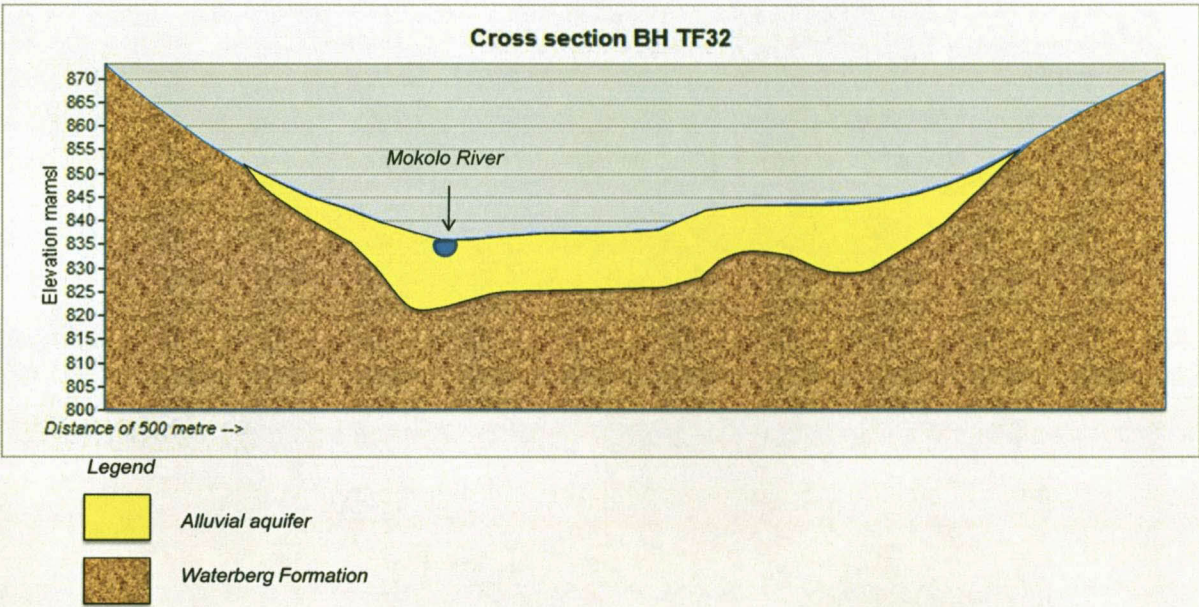


Figure 91: Section A: Cross section of the alluvial aquifer underlain by the Waterberg Formation in the mountainous area (Department of WaterAffairs², South Africa, 2010)

- Section B (Figure 92) is characterized by a flat lying area underlain by sediments of the Waterberg formation with an average alluvium depth of 8.5 m extending over a larger area compared to the mountainous regions of the Waterberg Formation. The hydraulic gradient towards the river is 0.00043 indicative of a more gentle topography.

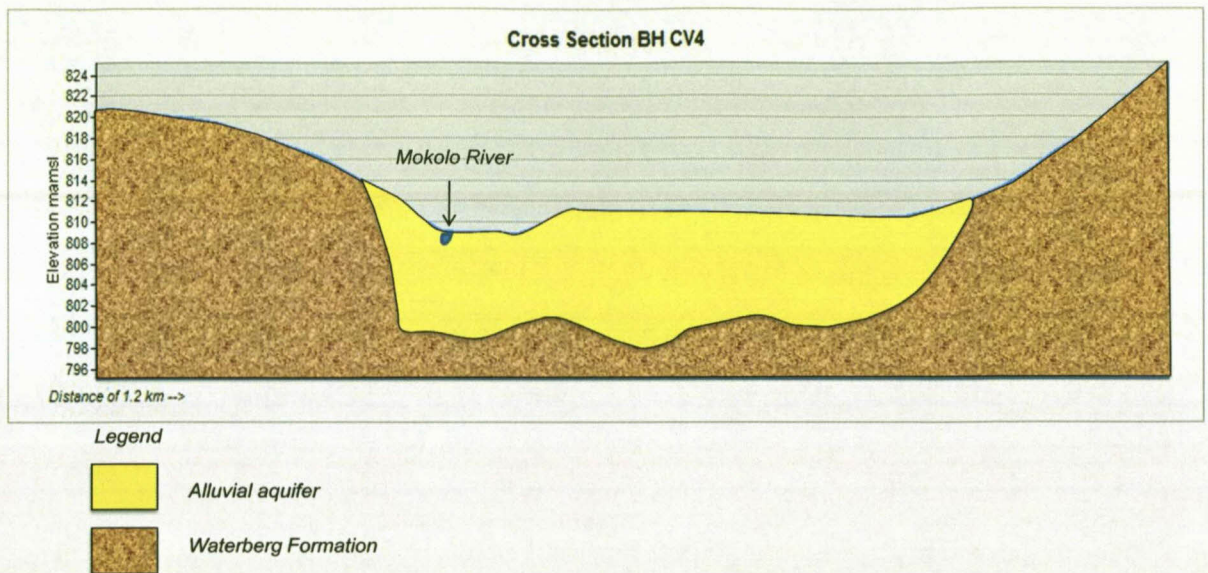


Figure 92: Section B - Cross section of the alluvial aquifer underlain by the Waterberg Formation in the flatter areas (Department of WaterAffairs², South Africa, 2010)

- Section C (Figure 93) is characterized by alluvium underlain by rocks of the Eccca Group. The alluvium has an average depth of 7.5 m. The hydraulic gradient away from the river is estimated at 0.00045 indicative of a more gentle topography.

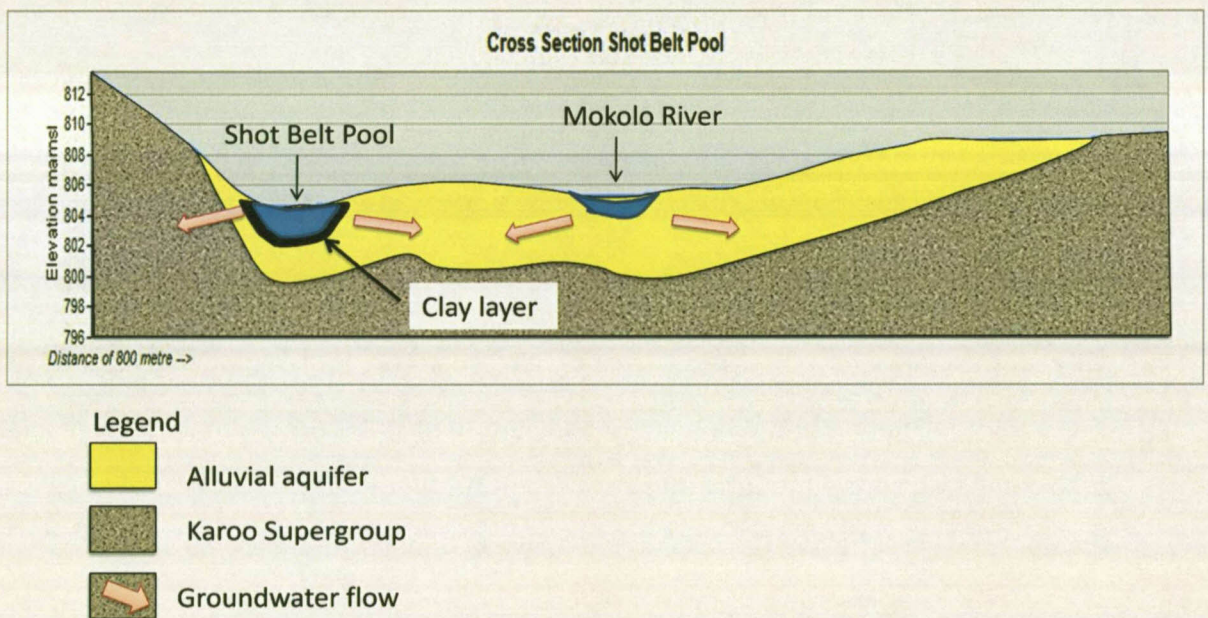


Figure 93: Section C - Cross section of the alluvial aquifer underlain by the Eccca Group at the Slot Belt Pool (Department of WaterAffairs², South Africa, 2010)

- Section D (Figure 94) is characterized by alluvium with an average depth of 6.5 m underlain by igneous and metamorphic rocks of the Basement Complex. The hydraulic gradient towards the river is estimated at 0.00043 again indicative of a more gentle topography.

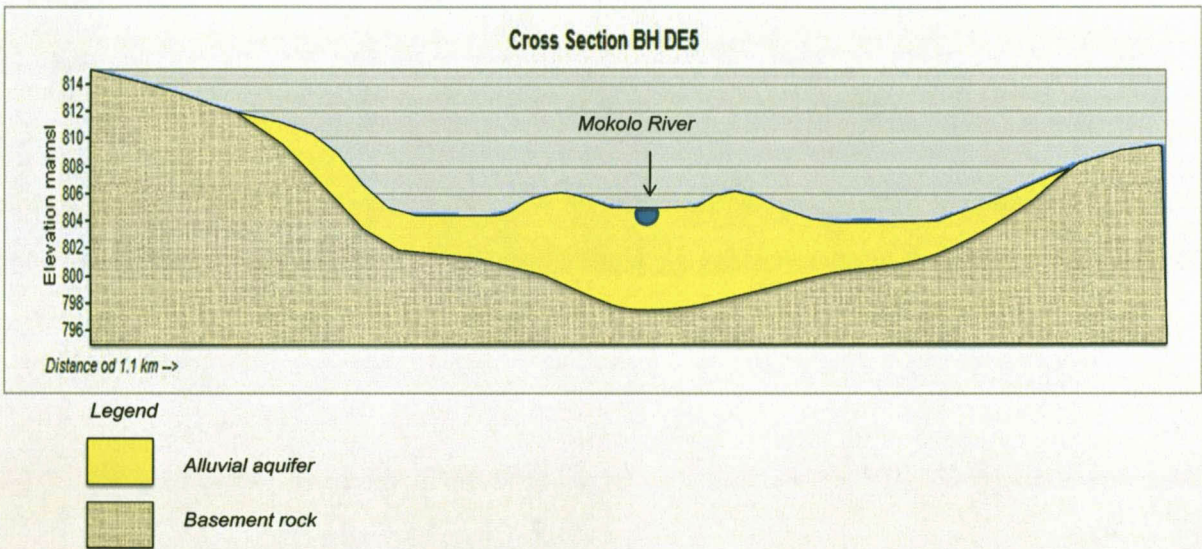


Figure 94: Section D - Cross section of the alluvial aquifer underlain by the Basement Rocks (Department of WaterAffairs², South Africa, 2010)

3.5.8 Chemistry analysis

Groundwater samples located in the quaternary catchment (A42G) that mainly constitutes the Waterberg Formation in the mountainous region are enriched in HCO_3 and lower in TDS values. This could as pointed out by the investigators be due to this area being a recharge area, or least impacted by land based activities. On the other hand further downstream groundwater samples in the quaternary catchment (A42J) that is underlain by Karroo rocks and basement is characterised by Na-Cl rich water type and higher TDS values.

Figure 95 is also a cross section (conceptual) indicating the relationship between the river, pool and the alluvial aquifer with respect to water flow exchanges based on groundwater level trends in the alluvial aquifer and flow monitoring in the vicinity of pools, the river as well as other geohydrological information.

Groundwater contribution to surface water or baseflow could not be determined in this study due to lack of adequate and appropriate data, but rather water exchange that may entail interflow, leakage from surface water bodies and possible groundwater contribution as lumped rather than discerned contributions.

3.5.9 Conceptual model of the study area

Based on detailed field work to classify the groundwater-surface water interaction of the Mokolo River as part of the groundwater Reserve determination study (Department of Water Affairs², 2010) the following conceptual aquifer model for the area is proposed:

- The alluvial aquifer associated with the Mokolo River is in direct contact with the river, i.e. no significant clogging layer (impermeable layer) is present in the river bed itself (refer to Figure 95). However, clay later exists between the pool and the alluvium.
- The alluvial aquifer is generally unconfined
- The regional fractured aquifers (Waterberg, Karoo and Basement aquifers) of moderate hydraulic diffusivity are in limited interaction with the alluvial aquifer.
- The regional aquifers show marginal gradients towards the Mokolo River course and exchange water with the river only indirectly via the alluvial deposits.

The surface water - groundwater exchange between the alluvium and the Mokolo River occurs on a far shorter time scale in comparison to the interaction between the regional and alluvial aquifers.

Water exchange broadly entails dam overflows into ecologically important pools (such as the Shot Belt Pool) and the groundwater contributions from the alluvial beds into the river. The latter is deduced from water level responses in piezometers and river stage measurements as well as hydraulic gradients. The interaction is also confirmed by similarity in chemistry between pool water and groundwater even though the rate is found to be low partly due to a low permeability layer between the alluvial bed and the river. On the other hand the poor interaction between the regional and alluvial aquifers was neglected.

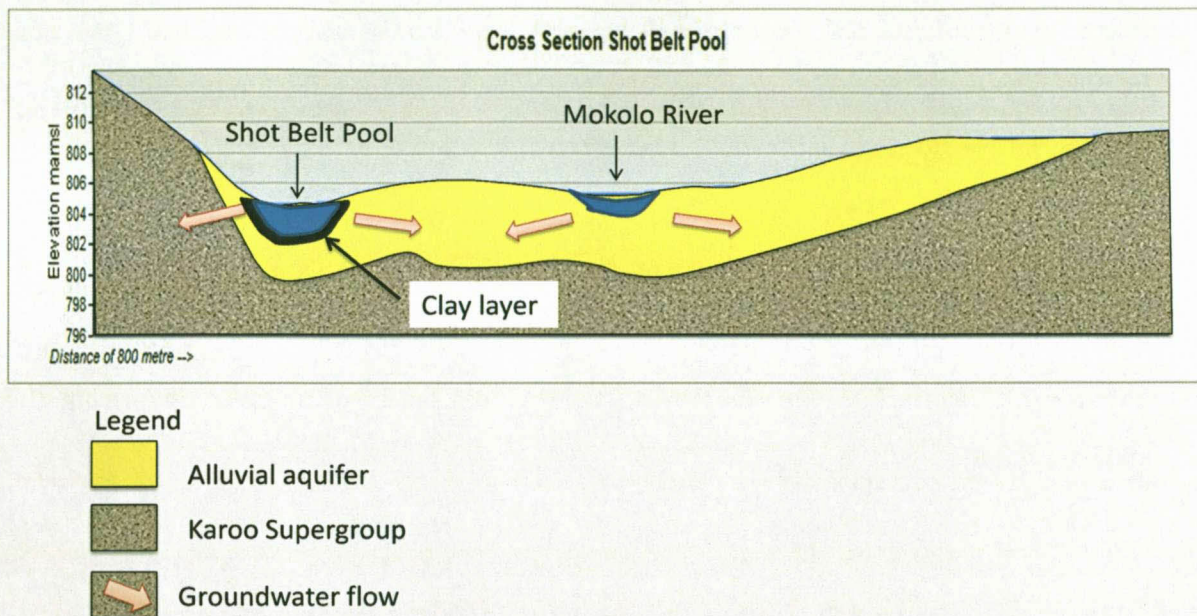


Figure 95: Mokolo River is in direct contact with the alluvial aquifer, and both the river and the pool lose water to the alluvial aquifer (which is same as Figure 93); (Department of WaterAffairs², South Africa, 2010)

3.5.10 Key issues that are worth noting from the Mokolo Case study

The following are some of the lessons from this case study:

- The groundwater recharge, occurrence and dynamics in the Mokolo catchment are controlled by geology, topography and structure with Waterberg Formations as the most productive aquifer. Most of the water use in agriculture, near the foothills of the mountains is sourced from this aquifer.
- Mokolo River and Shot Belt Pool are mostly losing water to the alluvial aquifer than the other way round
- Chemistry is crucial in confirming the existence of interaction between Shot Belt pool water and the alluvial aquifer
- Dam spills and releases have a significant influence on groundwater levels in the alluvial aquifer
- The water level in the Shot Belt Pool is primarily maintained by infrequent overspills from the Mokolo River in the vicinity of the pool in combination with low levels of exchange between the pool and the alluvial aquifer. Shallow clay layers with very low permeability significantly limit any potential exchange of water between the pool and

the alluvial aquifer. The water level gradients were consistently away from the pool throughout the monitoring period indicating losing (i.e. influent) conditions with regard to the pool (i.e. surface water replenishing groundwater).

- Additional pump testing in Mokolo should be undertaken and subsequently hydraulic parameters estimated to confirm or review the current properties (especially T values). It may also be important to check if there are seepages, along face rock or above impervious layers into the river from the regional fractured aquifer underlying the alluvial system.

3.6 The case study of the Krugersdrift Catchment

This case study is based on the research project commissioned by the Water Research Commission of South Africa in 2010 and undertaken by the University of the Free State led consortium. The project is aimed at assessment of surface water - groundwater interaction mechanisms and processes at local and catchment scale as well as collection of relevant data.

3.6.1 Location of the case study

The case study is located within the Krugersdrift catchment, downstream of Krugersdrift Dam along Modder River, in the Free State about 35 km north-east of the city of Bloemfontein. This case study will mainly be focused on observation boreholes at study site 1, site 2 and site 3 as indicated in Figure 96, to test if there is interaction in the alluvial setting along the river channel.

The delineated plan view of the broader area covering the study sites is also shown on Figure 97. The area is likely to be developed further by the Institute of Groundwater Studies over the years as a field laboratory for similar studies. It is clear from this figure that the major land-use practice in the area is irrigation as evidenced by coverage of centre pivot irrigation patches.



Figure 97: A plan-view of the Krugersdrift study area (Kotze, 2011)

3.6.2 Climate and vegetation

The study area is characterized by typical semi-arid climate with low and unpredictable seasonal average precipitation of 550 mm per annum. Most of the precipitation is experienced during the wet season (October to April) with the maximum rainfall period occurring from December to February. The potential evaporation transpiration rate in the Modder River quaternary catchment covering the study area is estimated to range from 1800 to 2000 mm. A variety of riparian vegetation types along the Modder River include grasses, trees, bushes, and shrubs (Tsokeli, 2005). The riparian vegetation cover besides evapotranspiration improves stability of the river bank (Figure 98).



Figure 98: Vegetation cover stabilizes the banks of the Krugersdrift (Tsokeli, 2005)

3.6.3 General topography of the area and river flow

The landscape of the study area lies between 1237 and 1255 mamsl elevation in the shallow open valley. As a result of the relatively flat topography and low gradient of the area, runoff is low (2.5 mm to 5 mm), and the flow of the river is also almost stagnant. This leads to damming of the river (Figure 99) with a broad weir at about 500 m downstream, hence the dam makes the river stage relatively higher compared to other regions.

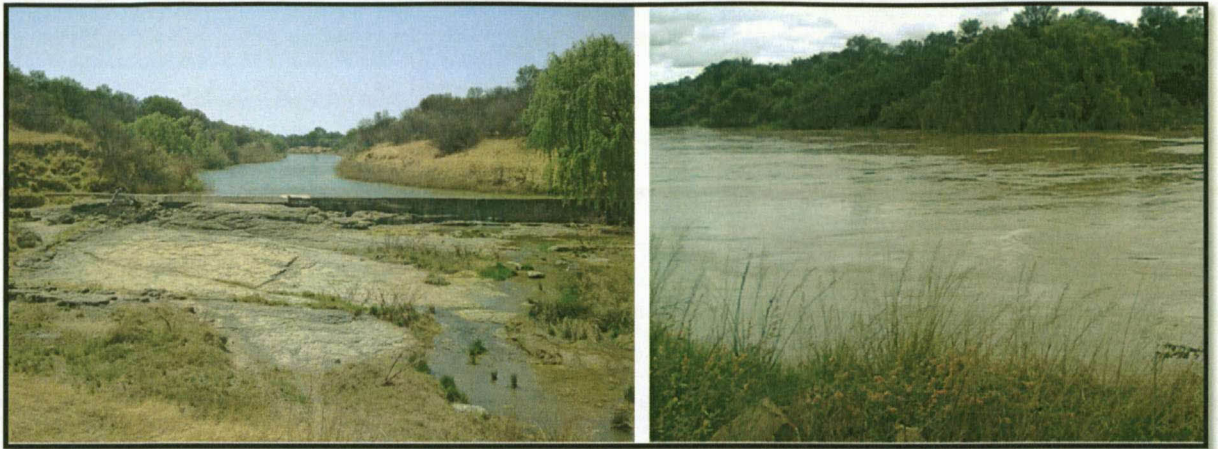


Figure 99: Krugersdrift stage during dry (left) and wet (right) period (Steyl *et al.*, 2011)

Downstream of the Krugersdrift Dam (Figure 100) a number of pans exist due to the relatively flat surroundings.



Figure 100: Krugersdrift dam (Google Earth, 2013)

These pans (Figure 101) are usually filled up during the summer heavy rainfall season and do not directly contribute to the river flow. The Modder River straddles the area flowing from east to west (see Figure 102 also).

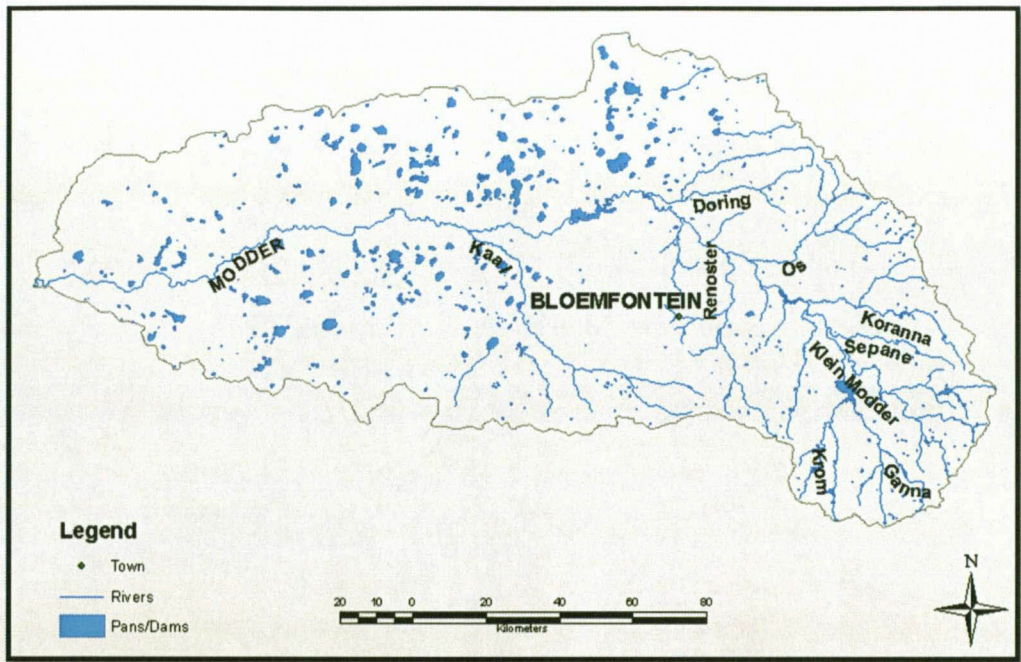


Figure 101: The Modder River drainage system and associated pans and dams (Steyl *et al.*, 2011)



Figure 102: The Krugersdrift intersects the study area and flows westwards (Kotze, 2011)

3.6.4 Monitoring borehole network and groundwater regime

3.6.4.1 Cluster of borehole network

A cluster or network of boreholes were drilled downstream of the dam within the study area about 3 km apart on either side of the river as shown on Figure 103, to investigate groundwater influence and for monitoring purposes. Site 1 monitoring network is located within the area that is characterized by calcrete outcrop. These boreholes were drilled to different depths to target different aquifer systems and to determine the flow direction or hydraulic gradient. Water was generally struck at about 12 m depth. However the water level is within the gravel layer (alluvial aquifer) for all these boreholes.

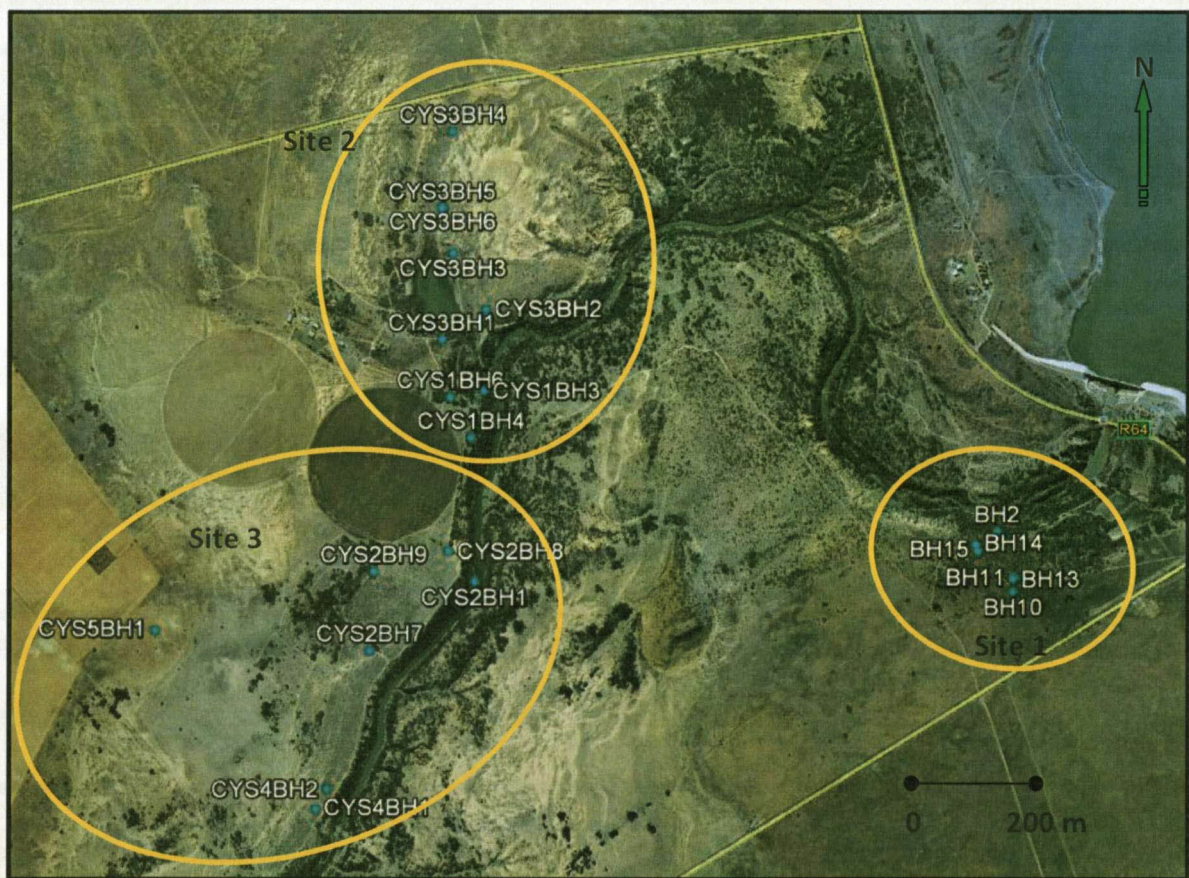


Figure 103: A cluster of boreholes in the Krugersdrift study area (adapted from Gomo, 2012)

The second cluster (site 2 monitoring network) was also drilled to varying depths to intersect different aquifer systems. However, these boreholes are far apart from each other, with some quite a distance (200-300 m) way from the river as indicated in Figure 103. The third cluster (site 3) has been drilled further downstream, mostly located within the riparian zone while at least 2 boreholes are sited further away from the river.

3.6.4.2 Groundwater occurrence

Recharge is typically enhanced by the high porosity calcrete that occurs in the area (Steyl et al, 2011) with borehole yields in the catchment ranging from 0.5 l/s to 2.0 l/s on average and where more than 10% of the recorded boreholes exhibit yields exceeding 5.0 l/s. Groundwater occurrence in the Eccca Group is principally associated with dolerite contact zones, joints and bedding planes and the Adelaide Subgroup of the Beaufort Group has been extensively intruded by dolerite sills and by dolerite dykes. Groundwater in these instances also occurs in joints and fractures on the contact zones, weathered dolerite zones, weathered and jointed sedimentary rocks and on bedding planes. Besides elevated nitrate levels recorded in some boreholes which are mainly attributed to agricultural practices, groundwater is generally of acceptable quality since the area is still relatively least affected by land based activities such as mining and industries.

3.6.5 Geology and structure

Figure 104 represents the geology of the Modder river catchment area.

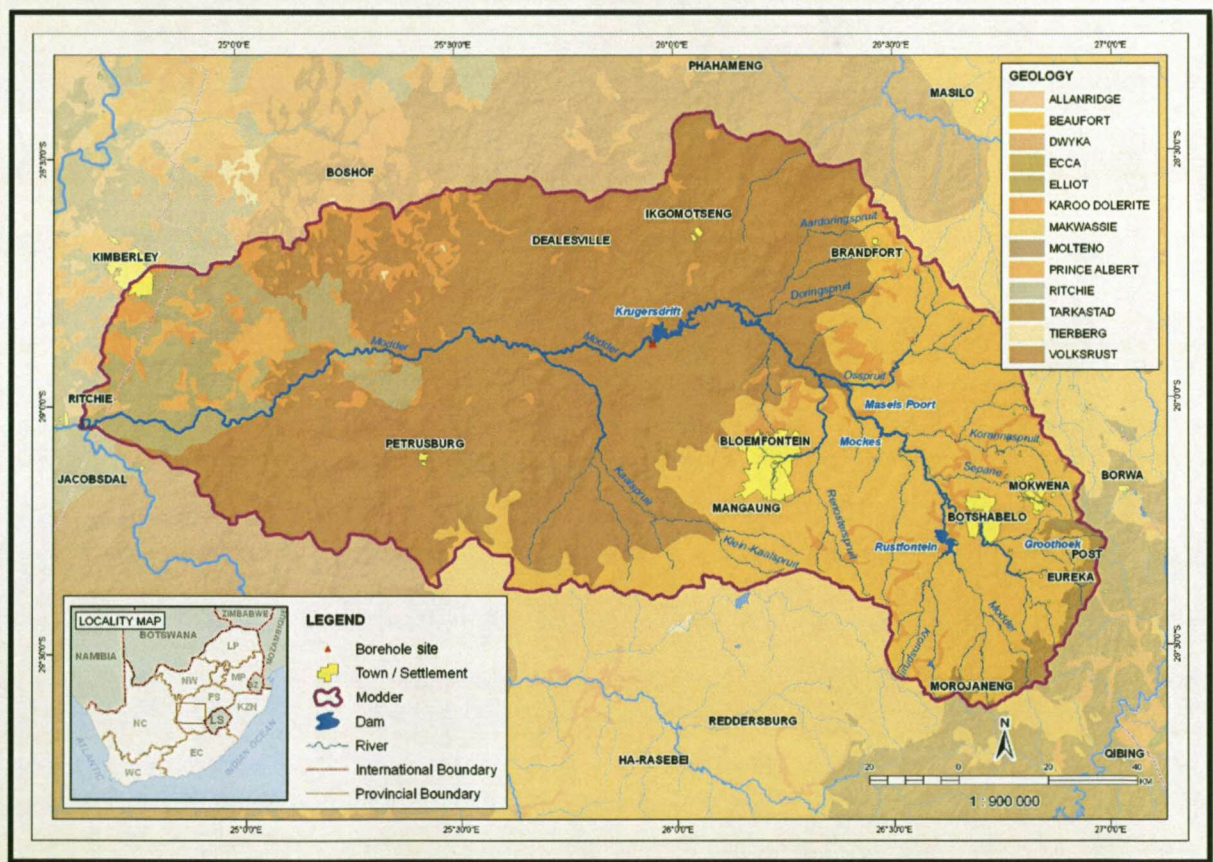


Figure 104: The Regional Geology of the Modder River Catchment

3.6.5.1 Geological profile

Some boreholes were drilled within the river bank while others were drilled inland about 200 m away to enable characterization of a more comprehensive conceptual model across the study site. Figure 105 depicts geological cuttings and logging of one the inland boreholes (BH 10) illustrating a typical geological profile in the study area.

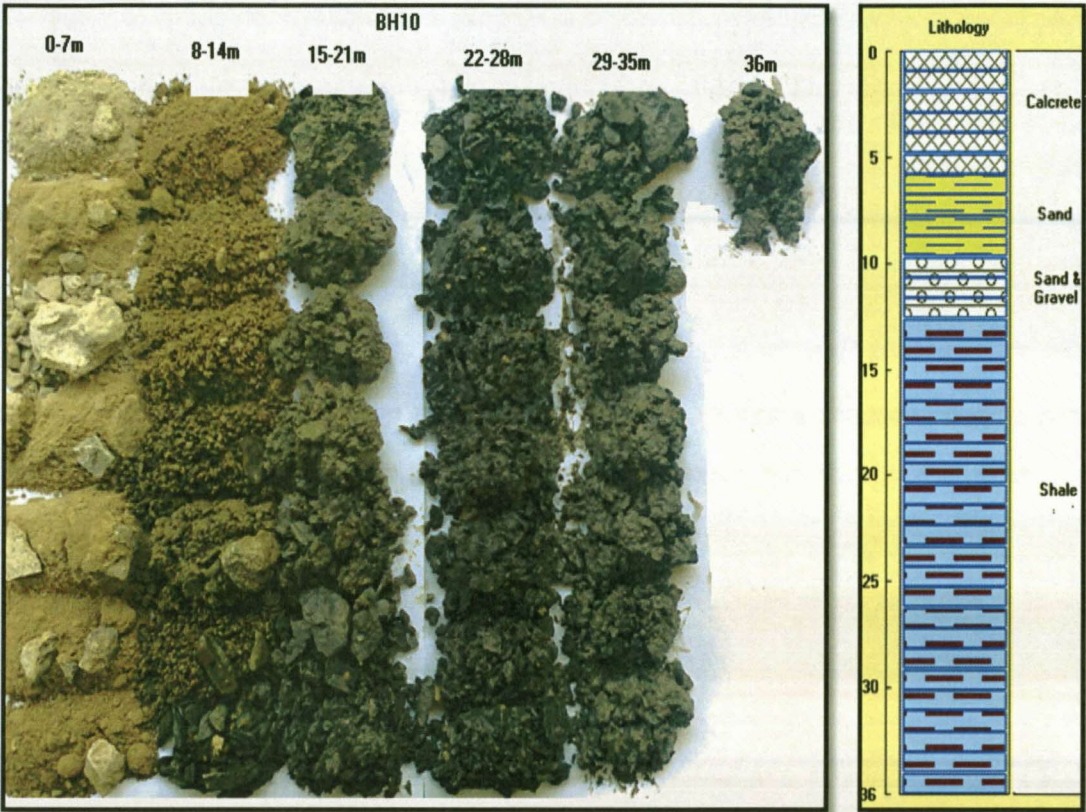


Figure 105: Geological samples and logging showing geological profile (Steyl *et al.*, 2011)

Figure 106 depicts a cross-sectional view of the positioning of the boreholes from the river landward and relative to the others in terms of elevations, and depth. Water strikes and the groundwater level as obtained in February 2011 are as indicated in the figure. It is quite clear from relative levels of groundwater and surface water that the Krugersdrift is a gaining stream fed by the sandy gravel layer (alluvial). Given average climate conditions in this area as well as the landscape, this could be a prevailing state unless flooding occurs that may reverse this scenario.

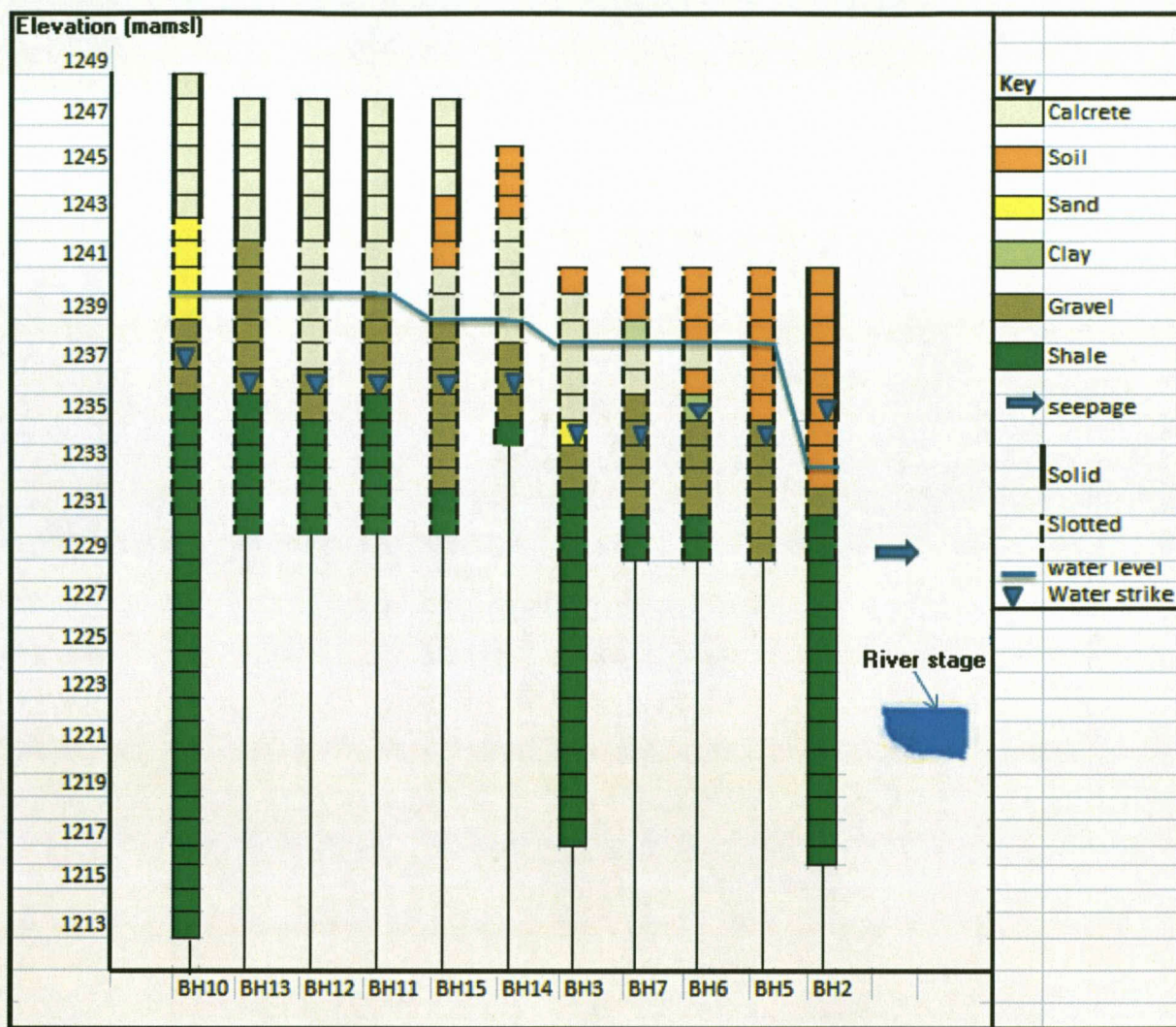


Figure 106: Cross-sectional view showing geology and hydrogeological features of the study area (Steyl et al., 2011)

Figure 107 represents the groundwater contour levels in site 1 boreholes. The contours indicate that the river has potential to gain water from the aquifer.

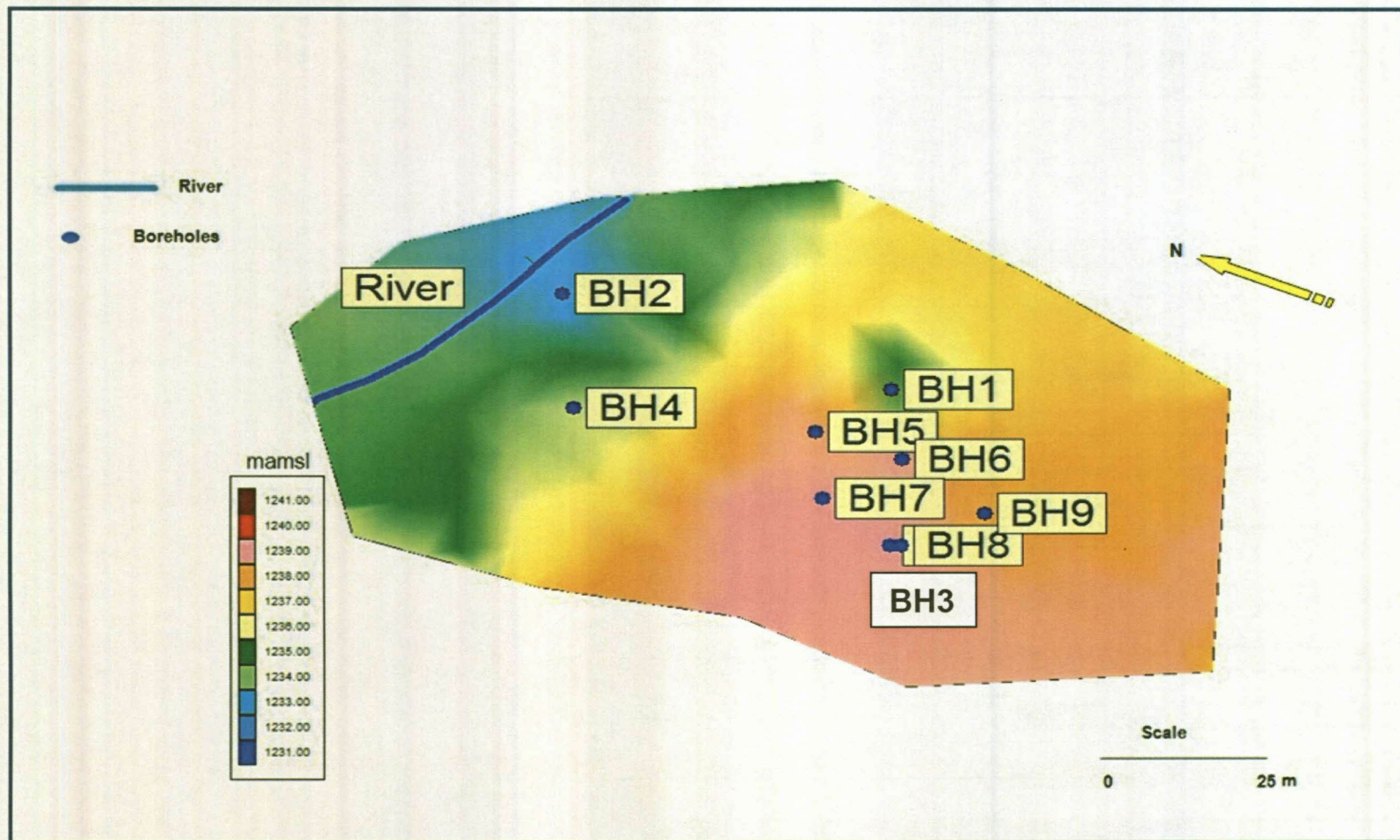


Figure 107: Groundwater level contours & relative borehole positions for site1 of the Krugersdrift study area (Source data - Steyl, 2012)

3.6.5.2 Conceptual model of the geology of Krugersdrift study area

Calcrete dominantly overlies the study area. The calcrete is underlain by a layer of gravel and sand material, comprising of rounded mudstone pebbles and rough gravel which includes quartz particles. This layer has both the highest porosity and the highest effective porosity compared to the whole geological profile indicating that the gravel formation is the zone of groundwater flow (Steyl *et al.*, 2011). Figure 108 is the conceptual model of the study area based on geological logs.

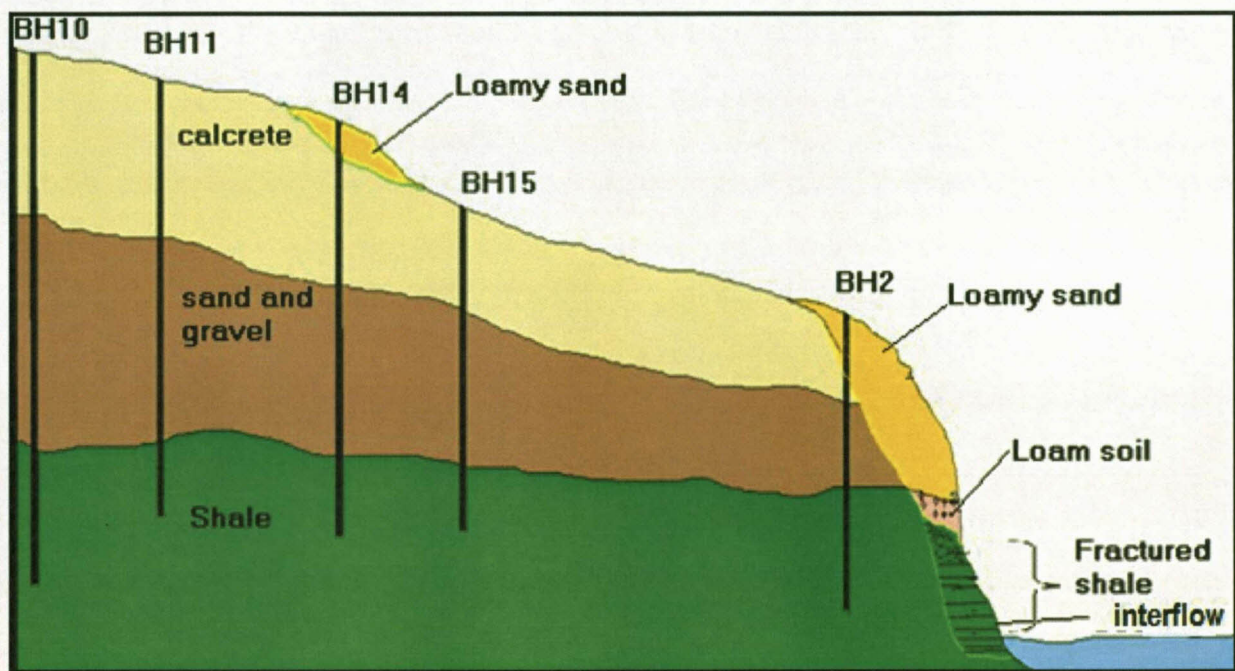


Figure 108: The geological conceptual model (cross section) of the Krugersdrift study area (Steyl *et al.*, 2011)

The shale-bed below the gravel layer probably impedes groundwater infiltration to layers below. Thus the flow topographically, in a sub-horizontal manner, follows geology down gradient towards the river then seeps through the fractured shale bed at the face of the succession as interflow as indicated in Figure 108.

3.6.6 Surface water - groundwater interaction in the alluvial setting

3.6.6.1 Isotope hydrology analysis approach

The entire isotope data points plot on the evaporation line as indicated in Figure 109. It is worth noting that “Boreholes” refers to groundwater samples from study site1, while “Boreholes_2” are groundwater samples from study site2 and site 3. Groundwater samples from the study site 1 are clustered in the relatively lighter area (depleted in heavy isotopes) and plot nearer the GMWL, while samples from the other 2 sites are spread across the evaporation line. Thus no evidence of interaction between groundwater and the surface water can be shown. The possible explanation may be that the river water is flowing relatively faster near the dam upstream compared to groundwater contribution and hence the isotopic signature the “mixed water” is not detectable.

However, some of groundwater samples from the other 2 sites plot in the same area as the weir and river water samples. Thus indicating that interaction between groundwater and surface water further downstream is occurring (i.e. near study site2).

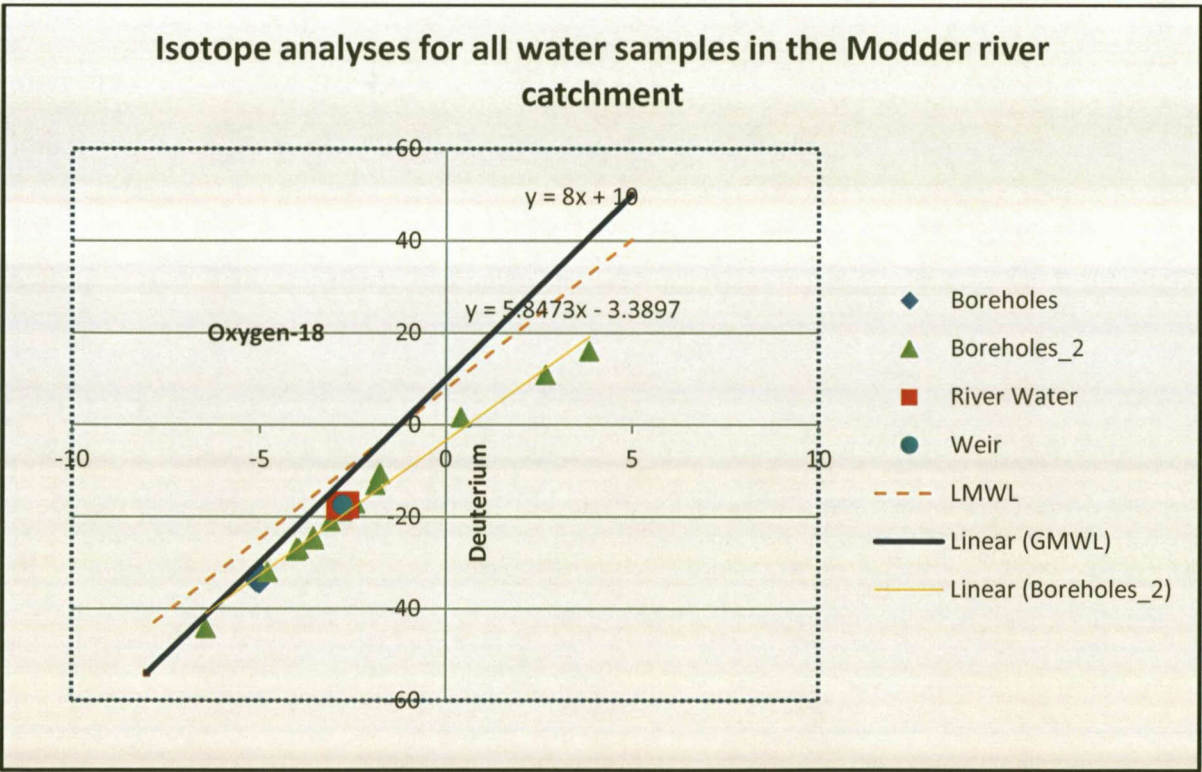


Figure 109: The stable environmental isotope data for Krugersdrift catchment (Source data - Steyl, 2012)

Figure 110 shows indeed that the same cluster of boreholes from site1 is quite different in isotope signature to the isotopes that were collected from the river and the weir.

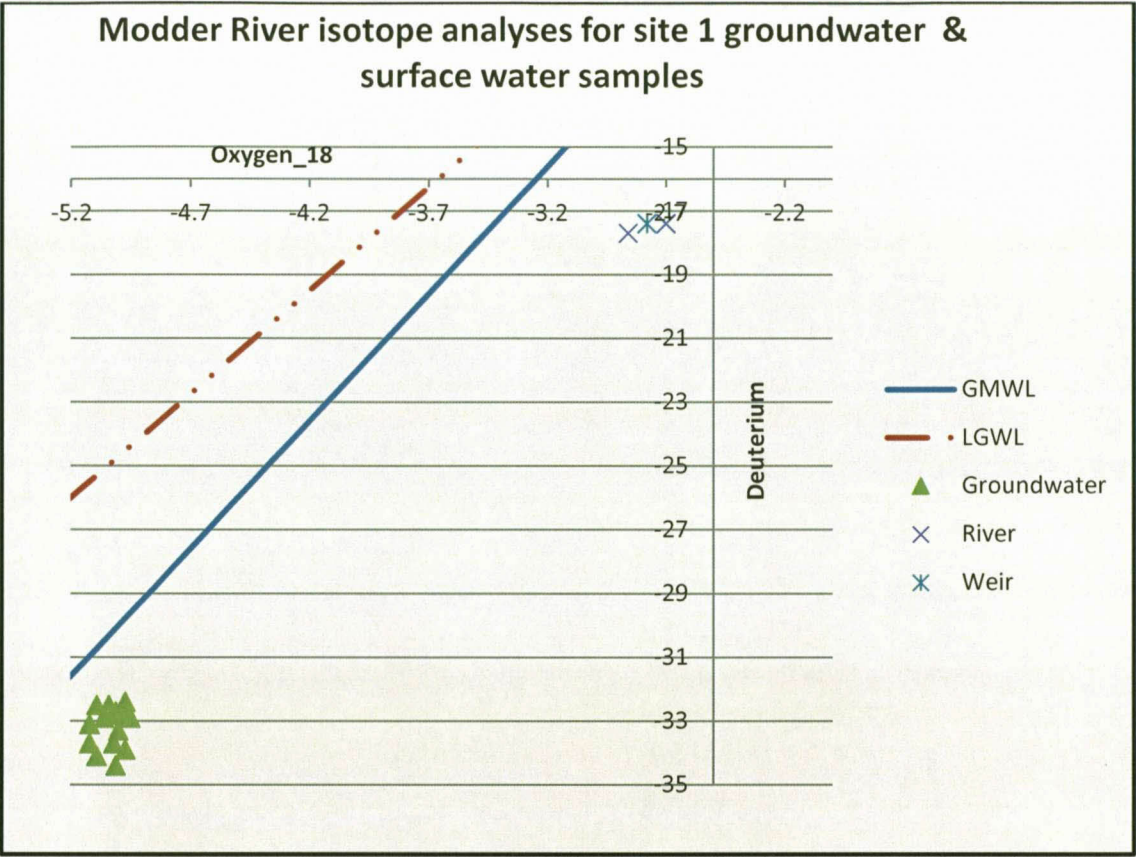


Figure 110: Isotope analysis for surface water and groundwater in site 1 (with those groundwater samples from site 2 excluded) for comparative purposes. (Source data - Steyl, 2012)

3.6.7 Chemistry

3.6.7.1 Ion balance error

The Ion Balance Error as shown in Figure 111 is less than 5% for the groundwater data points but one (i.e. BH3) which is -5.56%. The surface water sample's error is too high (9.73) for surface water sample but still acceptable. This might be due to error in sampling.

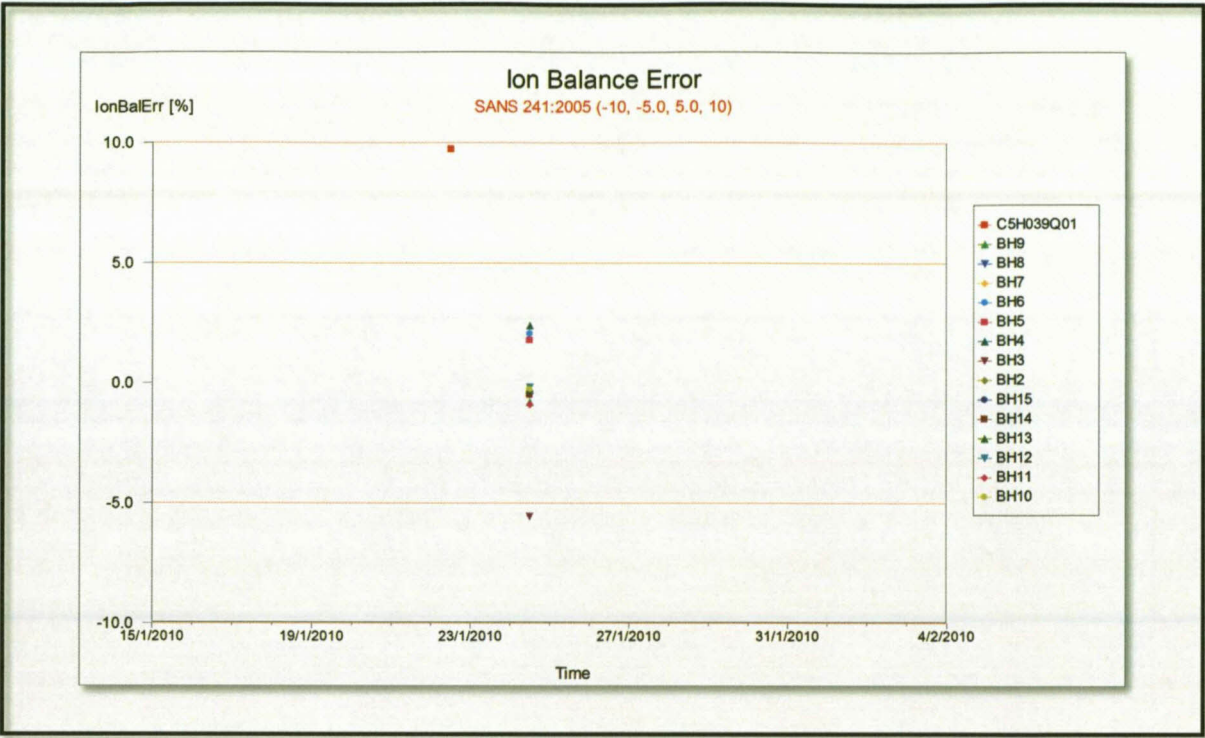


Figure 111: Ion Balance Error for the chemistry of the Krugersdrift samples (Source data - Steyl, 2012)

3.6.7.2 Chemistry data for the Krugersdrift study area

The Piper diagram (Figure 112) shows that the chemistry of both groundwater and surface water is similar. Thus indicting possible mixing and hence interaction between the two systems, though this is not very conclusive given the high ion error balance for the surface water sample. The Stiff diagram as indicated in Figure 113 is also plotted to show the chemical similarity, although the river sample is shown to be relatively enriched in Calcium and depleted in Chloride. Annexure 7 shows the chemistry data for the Krugersdrift study area from boreholes and seepages or interflow into the river collected in August 2011.

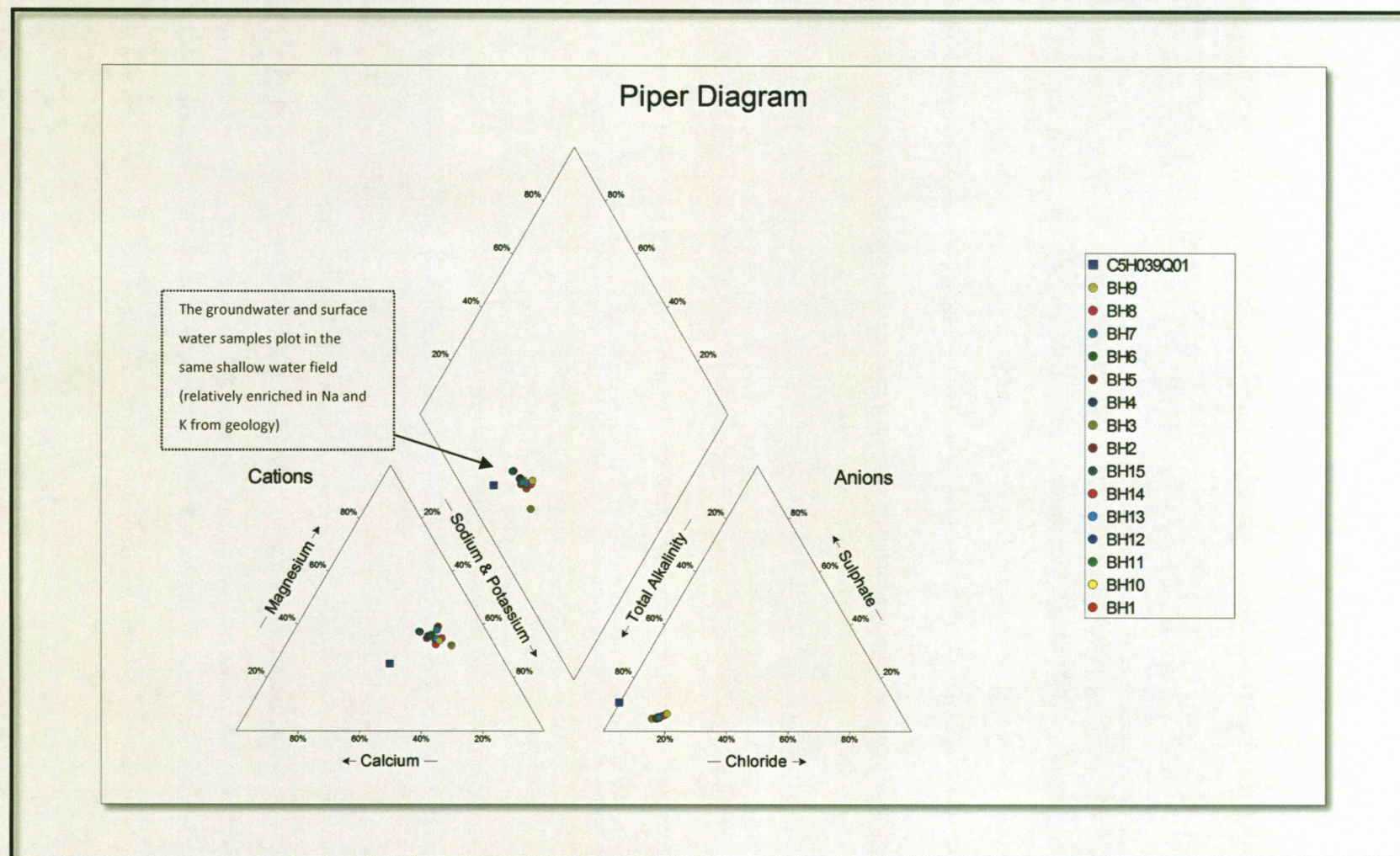


Figure 112: Chemistry of the water sample data for the Krugersdrift (Source data - Steyl, 2012)

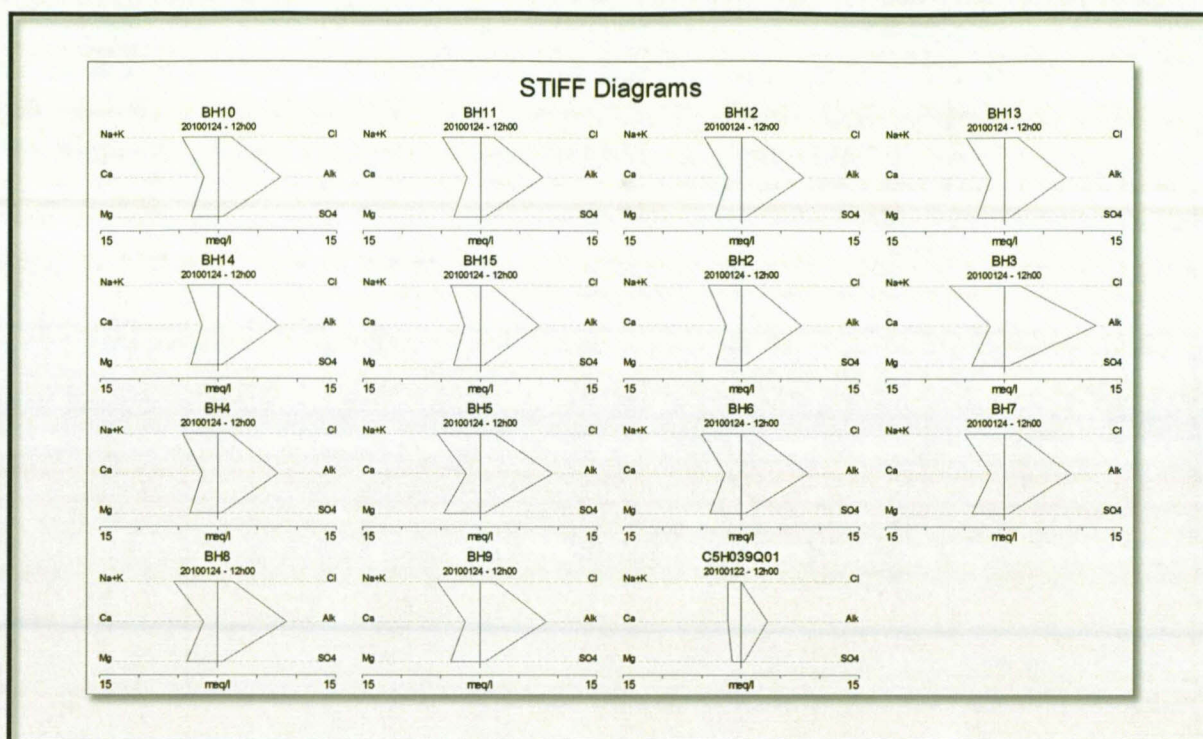


Figure 113: The Stiff diagram indicating the chemistry of the groundwater and the surface water sample
(Source data - Steyl, 2012)

3.6.8 Data requirements

Additional data for surface water, particularly chemistry and isotope data is necessary for improved assessment of the exchange between groundwater and surface water.

3.6.9 Lessons learnt from the Krugersdrift case study

- The study site needs more extensive monitoring on a regular basis and collection of data including that of dissolved oxygen and temperature of both groundwater and surface water samples. Otherwise, additional instrumentation such as seepage flow meters to measure fluxes between groundwater and surface water would be useful.
- Seemingly, based on observation of the geology and topography, the groundwater component that discharges into the river is both baseflow and interflow. Most probably markedly the latter as seepage along the edge of the river has been observed.

- The river may be subdivided into reaches or sections with appropriate instruments placed strategically and optimally at key observation points.
- The marked distinction in chemistry between surface water and groundwater could be due to the fact that the groundwater seepage is very little compared to dam releases thus diluting surface water that might have some chemical signature of groundwater. To remedy the situation sampling further downstream may be the way around this challenge.
- In this case study, isotope data indicated possible interaction to occur downstream.

4 : Case studies - Analyses and critique

4.1 Introduction and general comment

To guide or inform choice of appropriate methodology for assessment or quantification of surface water - groundwater interaction, associated assumptions should be taken into account as a prime test for relevance to the conceptual model of the area of interest. In other words the question that should come to mind is whether given the conceptual model of the study area, the assumptions related to the model or method would still be valid. The required amount of input data vary from one model to the next, with some methodologies or models being more data intensive (e.g. numerical models) than others.

In each of the case studies used in this research the approach followed in evaluation of water exchange between surface water and groundwater was markedly influenced and often to a larger extent limited by availability of, or in most cases incompleteness of the requisite set of data. Nevertheless, thanks to a number of generous individuals out there some data and a lot of information were made available to enable the assessment process. However, an overarching or major lesson that one could learn from this research is that despite availability of several methods for evaluation of surface water - groundwater interaction, some of those methods are unlikely to see the light of day in real life practical application due to paucity of necessary data among others. On the other hand, some methods such as hydrograph separation, chemistry analyses, or environmental isotope analyses will invariably be ready for use and applicable since required data are often readily available.

With regard to monitoring and data collection for the purpose of assessing surface water - groundwater interaction, additional parameters can be collected using existing field observation points or by installing supplementary instrumentation. For instance in a semi-arid environment with markedly fractured aquifer systems such as is the case in South Africa, Radon (^{222}Rn) data would be useful, yet this radioactive noble gas is seldom captured. When rainfall infiltrates and recharges the aquifer, the concentration of ^{222}Rn in groundwater increases due to radioactive decay of ^{226}Ra in soils, sediments and rocks.

However, when ^{222}Rn rich groundwater seeps into surface water the concentration of ^{222}Rn in surface water would rapidly decline due to short-life (of less than 4 days) thereof in surface water. Hence concentration of ^{222}Rn is typically higher in groundwater than in surface water. Therefore measurements of ^{222}Rn in surface water can be used to detect groundwater seepages. Other useful parameters that can be used in evaluation of water exchange between surface water and groundwater yet seldom captured such as temperature, can easily be collected by for instance, inserting probes among current instruments that are already setup.

Lastly, particularly for a semi-arid country such as South Africa that receives high solar radiation with most parts of the country's evaporation rates exceeding rainfall by 4 to 5 factors, measurements of transpiration by riparian vegetation is crucial. Seasonal riparian vegetation water use data can be generated using remote sensing technology (in conjunction with in-situ measurements for calibration) where riparian vegetation types are identified, canopy transpiration over time (i.e. during growing season) could be determined.

Now, this chapter seeks to make critical comment on the case studies that were further investigated with the intent to identify lessons that can be learnt or to recommend alternative approaches.

4.2 Richmond Farm Case Study: Estimation of baseflow

The results of earlier work by Groundwater Monitoring Services cc (2001) as well as ERM Southern Africa Pty Ltd & GMS cc (2004) on water balancing and model simulation of the unconfined weathered aquifer indicated that the river downstream is fed by the dam and by the weathered bedrock aquifer upstream, with net recharge to the river estimated at 326 l/s. This "baseflow" is an estimate for the entire catchment and happen to mostly entail interflow. Later investigations undertaken by van Tonder and Dennis (2005), as well as by Kotze et al.,(2006) dealt with the area at relatively higher resolution.

The methodological approach followed in the investigation and assessment for groundwater supply in the Richmond Farm is quite comprehensive and commendable. For instance, the geohydrological conceptual model of the of the study area was appropriately developed

with discernible aquifer systems identified and subsequently a long term pumping test undertaken to estimate sustainable yield as required. On one hand neither the environmental stable isotope analysis nor flow measurements established surface water - groundwater interaction. Hence it was concluded that the river did not gain any water from the aquifer. On the other hand, baseflow was estimated using hydrograph separation. This baseflow that was determined by hydrograph separation method is actually "a lumped parameter." In fact most of the baseflow reported is in essence interflow or mostly interflow. Hence baseflow in the stricter sense is the groundwater contribution to the river from the saturated zone. However, it does not mean that there is completely no groundwater contribution that occurred but rather that the amount is very little and hence relatively undetectable. It is also possible that the river water samples that were analysed for isotopic analysis were highly enriched due to evaporation and that masked the groundwater isotope signature despite some mixing, while the groundwater samples were extremely depleted. It should also be noted that the evaporation rate of the Richmond area (1 061 mm) is about twice that of rainfall (540 mm) on average, and more than 7 times thereof in winter (in June rainfall is about 7 mm while evaporation rate is 52 mm) based on data for a 70 year period.

The recommended approach would hence be similar to that by Kotze *et al.*, (2006), especially in term of first developing a conceptual geohydrological model based on available information and field observation. An alternative approach could be used instead of drilling all boreholes along the river channel within the riparian zone. Additional monitoring network of at least 2 or 3 boreholes should be drilled on either side of the river, in a triangular manner, to determine groundwater flow direction. The boreholes should be located at distances of 20 to 40 m away from the river to enable interception of water within and outside the riparian zone in various formations or layers within the site. A good example of an ideal network is as indicated as a section in Figure 114.

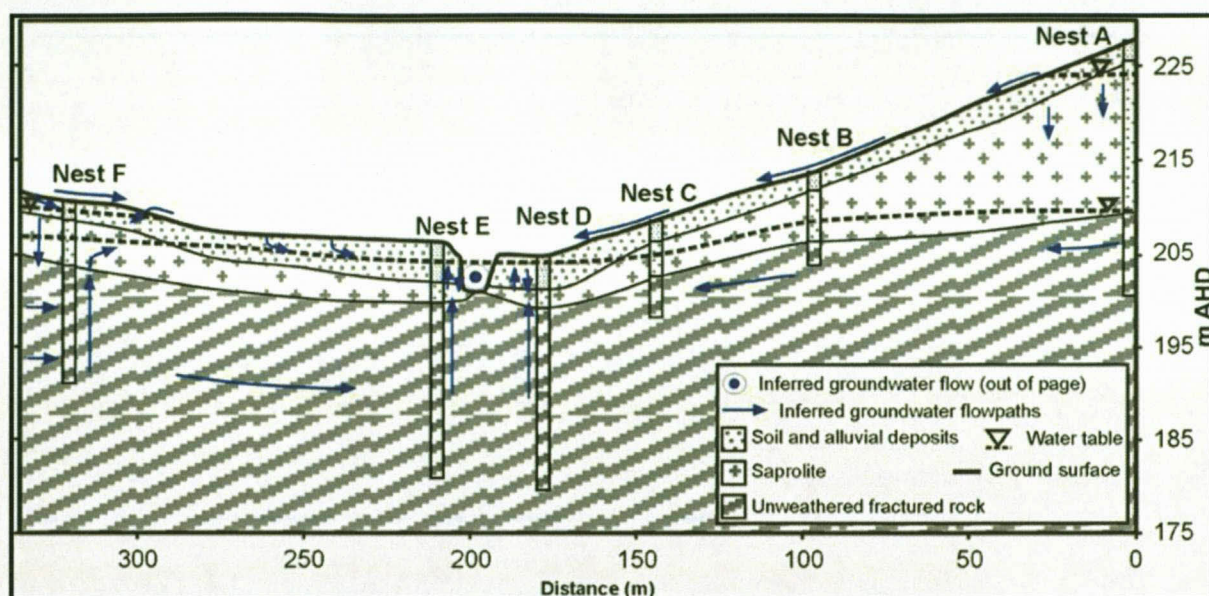


Figure 114: Groundwater monitoring network for evaluation of surface water - groundwater monitoring network (Adapted from Banks *et al.*, 2009)

These boreholes should be equipped with probes for capturing sets of data. The in-river monitoring observation positions should also be carefully paced away from the mouth of the dam; while seepage meters are strategically located in areas that are likely to be sources of groundwater/surface water leakages. Meteorological instruments such as the rain-gauge and evaporation pan should also be located or identified in the area.

4.3 The Weatherley Catchment Case Study: Baseflow estimation in the hillslope setting

The Weatherley catchment is a valuable field laboratory that has potential to benefit additional disciplines or fields that previously had not optimally utilised the services. Much of the work that has to date been carried out focussed on the hydrology of the hillslope and soil science. That enabled a better understanding of the hydrological conceptual model of the catchment, and that knowledge can be applied in similar catchments.

The stable isotope data-sets received from the University of KwaZulu-Natal were very helpful in the analysis and assessment of the interaction between surface water and

groundwater in the hillslope setting. Much of the groundwater contribution to surface water is interflow from perched groundwater formation or aquifer.

In terms of observation points the instrumentation that has been set up at Weatherley is World-class and enhanced investment should rather be in monitoring for additional parameters with existing equipment. For instance, it would have been more helpful if chemistry data such as macro elements and radon (^{222}Rn) could also be captured. It is also recommended that regular monitoring, at least on six monthly basis be done where groundwater levels from deep boreholes are measured.

4.4 Seekoei River Case Study: Analysis/evaluation of surface water - groundwater fluxes into pools

It was postulated that pools in the Seekoei area were sustained by upstream spring water and the groundwater flux from the channel aquifer. The sustained pool storage especially during extended periods of no river flow was reasonably attributed to effluent groundwater contribution. However, the sluggish and low groundwater flux into pools is most likely due to evapotranspiration by riparian vegetation.

A fair amount of data (isotope, hydrological and chemical) is availed for assessment of surface water - groundwater interaction in the Seekoei study area. However, it is recommended that monitoring on a regular basis be established at key strategic observation points in the catchment to at least address the following issues:

- Regular measurement of pool water levels to cover periods of high, low and no rainfall, and to confirm the assertion that groundwater flux sustains pool storage
- Capturing (as part of the monitoring) and subsequently analysing Radon (radioactive) data in addition to other chemistry data for detection of groundwater seepages
- The role of riparian vegetation in interception and water use that would otherwise be baseflow contribution to surface water needs attention. Notwithstanding calculations carried out by previous investigators to estimate water taken up by vegetation, which is commendable under the circumstances, seasonal monitoring data of riparian evapotranspiration is essential.

Researchers have already made preliminary investigations and captured once off data and time series data over the project life time. It is now the responsible monitoring agencies' turn to invest in the network and continue monitoring to sustain the process.

4.5 Mokolo River Case Study: Surface water - groundwater interaction in the Mokolo system

Given comprehensive work on surface water ground water interaction previously undertaken by researchers in the Mokolo area, the case study seemed at first to have been a done deal. However, on closer inspection a number of challenges and opportunities came to the fore. The first huddle was encountered when it became clear that the hydraulic parameters (e.g. Transmissivity values) as determined by earlier investigators (Vpond, 1987 & 1988; Water Geoscience Consulting, 2008 and Department of Water Affairs², 2010) were too high compared to those determined later in follow on research by VSA (2009); Dennis (2010) and Vermeulen et al.,(2010). This took a while to figure out since the gap between Transmissivity values proposed by earlier investigators (ranging from 591 to 1 835 m²/d) was too large compared to later investigations (0.1 to 500 m²/d).

That the aquifer types were named after the formation that underlies the alluvial aquifer under investigation caused earlier confusion during the initial stages of this research. For instance the upstream aquifer in the mountain area is referred to as the Waterberg flat mountain (with assigned T values of 1 489 m²/d and 1 835 m²/d by Department of Water Affairs² (2010) and Vpond (1987) respectively). However, on closer inspection, it turned out that in fact the investigators were referring all along to the alluvial aquifer! Indeed they stated that this is the alluvial aquifer which is expected to have a relatively higher T value. In other words, there is no contradiction between the earlier and later researchers because the former had determined the T values for the alluvial aquifer while the latter determined T values for the underlying aquifer whose trasmissivities are comparatively very low. However, the nomenclature or naming of the aquifers is still strange and confusing.

It is recommended that observation network for groundwater monitoring should be similar to that which was recommended for Richmond area. It would be much useful to understand how groundwater contribution from the hard rock aquifers on the sides of the channel and at distances away from the channel occurs. Otherwise, if boreholes, for practical and

accessibility purposes in terms of drilling, can only be drilled within the riparian zone (perhaps due to landscape challenges) then such monitoring boreholes should be nested, and drilled to various depths to reach particular aquifers of interest as indicated in Figure 115. It is important to ensure that one aquifer is sealed from the other above or below with bentonite where piezometers (made of PVC) are isolated for specific monitoring purposes.

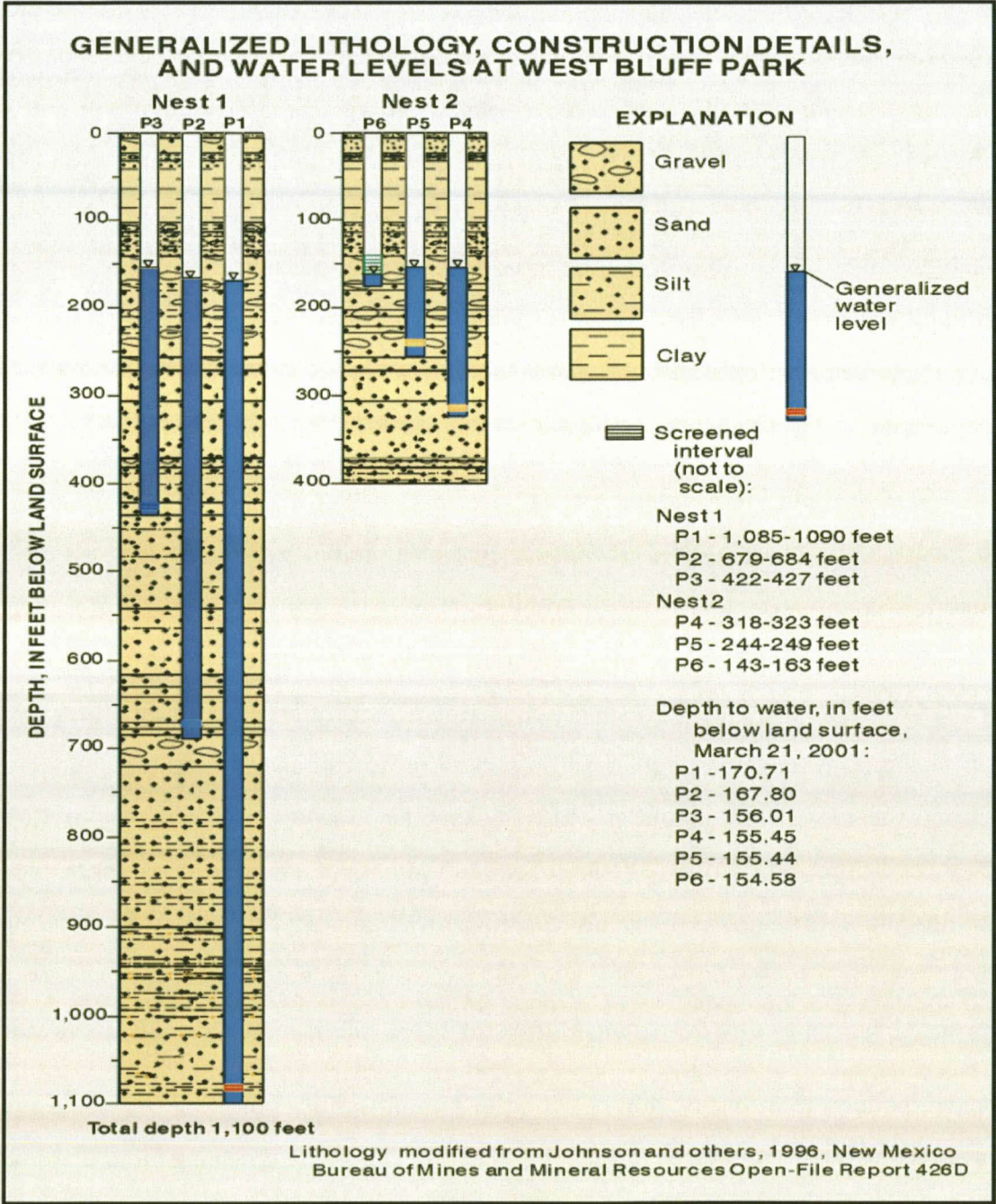


Figure 115: Example of nested boreholes drilled to different aquifers (USGS West Bluff Monitoring site)

4.6 Krugersdrift Case Study area – Investigation of surface water - groundwater interaction in the alluvial setting

The Krugersdrift case study area may become an ideal site for surface water - groundwater interaction in an alluvial setting provided an improvement in borehole positioning or construction is done. Various methodologies could be tested and applied at various scales and geohydrological settings in addition to the alluvial aquifer setting.

Data obtained for this case study was quite valuable for assessment purposes. However, time series of various parameters is necessary and regular monitoring is required to generate time series. Similarly to the Mokolo and Richmond case studies observation network should include appropriately drilled and nested boreholes, capturing of additional data including ^{222}Ra and temperature data time series should be collected.

5 : Classification systems and framework for surface water - groundwater interaction methodologies

5.1 Introduction

The choice of appropriate methodology for qualitative assessment or quantitative estimation of surface water - groundwater interaction should be preceded by a framework that guides the selection process. Hence, in this chapter classification systems are investigated to identify or inform development of a framework that informs the methodologies for surface water - groundwater interaction, that are appropriate for South African geohydrological conditions. The approach is to first consider current international trends followed by national initiatives, thus leading to development or evolution of suitable tools for local conditions.

5.2 Australia

In support of the National Water Initiative aimed at ensuring *“recognition of the connectivity between groundwater and surface water resources and for connected systems to be managed as a single resource,”* Brodie et al., (2007) on behalf of the Australian Government developed an adaptive management framework. The purpose thereof is to address water resource related challenges such as double counting in allocation of surface water and groundwater that are linked where the exchange is not taken into account, and water quality impacts due to interaction between groundwater and surface water. The framework is an adaptive process that entails a checklist of activities undertaken in an iterative manner with feedback loops as indicated in Figure 116, into the entire management process.

The component entitled “Identify Management Setting” is the first step and entails identifying existing planning and policy, key issues facing water management in the catchment, water use and water resource development. The next component of the

framework, “Assess Water Resources” involves acquisition of baseline information necessary to describe groundwater surface water systems and their interaction. The process includes collation and interpretation of geohydrological attributes, climate parameters, including time series records on water chemistry, flow volumes and water levels.

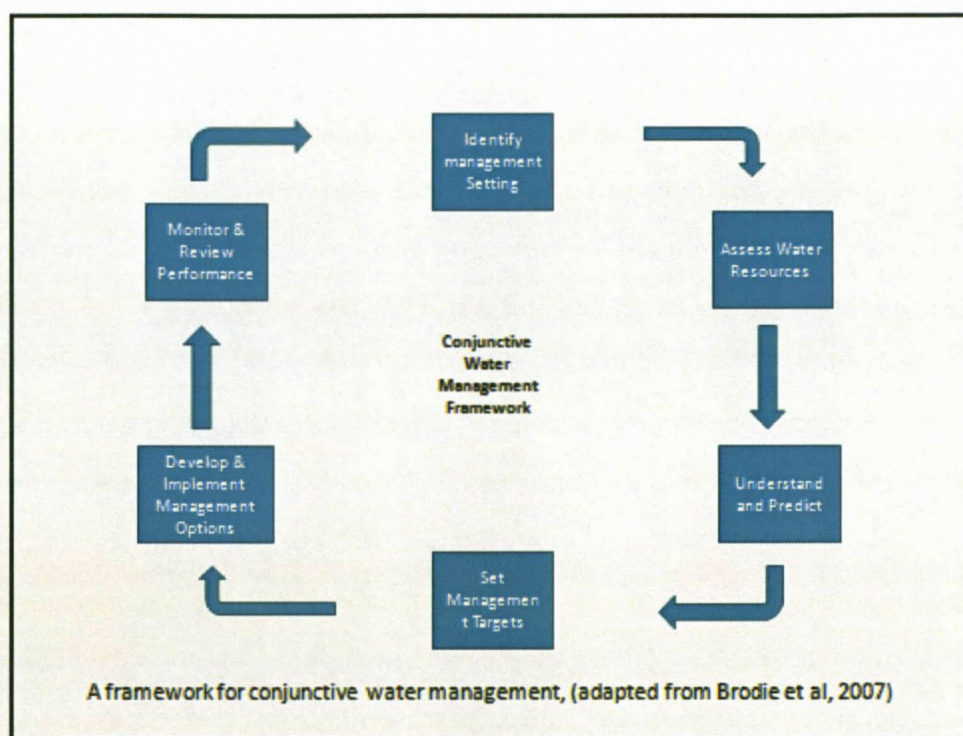


Figure 116: A framework for conjunctive water management (adapted from Brodie *et al.*, 2007)

This particular step, and the next on “understanding and predict,” have direct relevance to the research study. This is particularly regarding consideration of methods available to assess the nature and degree of the interaction between groundwater and surface water systems. In the next step, “Understanding and Predict,” the information available is used to develop a conceptual model for classification of the river – aquifer linkages, nature and geometry of the groundwater flow regime, water balance, magnitude and dynamics of water fluxes, and for prediction of how potential water flux may change due to external factors such as climate change or extraction, and for identification of key information gaps that need to be addressed. The rest of components are as follows:

- “Set Management Targets” which is the basis for planning and management decisions,
- “Develop and implement Management Actions” wherein implementation of Policy options that recognise or take advantage of the linkages between groundwater and

surface water resources could include strategies such as licensing and allocation, water trading or risk management approaches, buffer zones or planning rules.

- “Monitor and Review Performance”
- Defining “Roles and Responsibilities” and
- “Future Priorities”

Clearly, this framework is broader than just methodologies for assessing surface water - groundwater interaction. However, there are important elements herein that could directly inform the framework sought in this research study, especially since South Africa and Australia have comparable geohydrological conditions and also have similar water related challenges such as high climate variability and water scarcity.

In this regard, relevant and important elements of the Australian framework particularly those that are considered for the classification scheme will be considered in the development of the methodologies appropriate for South African conditions. The classification scheme is based on the position of the water-table, the dominant direction of seepage, the permeability or conductance, and the impact of connectivity on water management targets. In the South African context, some of the targets could entail water management issues that need to be addressed such as the setting of Ecological Reserve, or the sustainable yield limit placed on allocation of surface water and groundwater resources.

5.2.1 Categorisation of connectivity between groundwater and surface water systems

Classification of connectivity of the hydrological system in terms of the relative position of the water table seeks to establish whether the saturated zone of the aquifer is in direct hydraulic contact to the river or not. If directly connected then any withdrawal of water from either system would then impact the other. In this case Brodie et al., (2007) refer to the system as being contiguous (or connected as generally referred to in the literature). On the other hand in cases where groundwater is disconnected from surface water, the river is regarded as perched. It is worth noting that in South Africa we commonly talk of perched aquifers while in the Australian case classification (Figure 117) refers to a reversed situation

where the unsaturated zone below the river separates the deeper aquifer from a perched river.

Connected water resources may also be classified in terms of direction of seepage where effluent (or gaining) reaches receive discharges from the aquifer while influent (or losing) streams recharge the aquifer. Intermediate to these possibilities are fluctuating streams probably due to climatic (or seasonal) influence); or variably gaining (or losing) streams depending on the relative stage along the stretch of the stream. In this case Xu et al., (2002) refer to these as intermittent streams of riffle and pools.

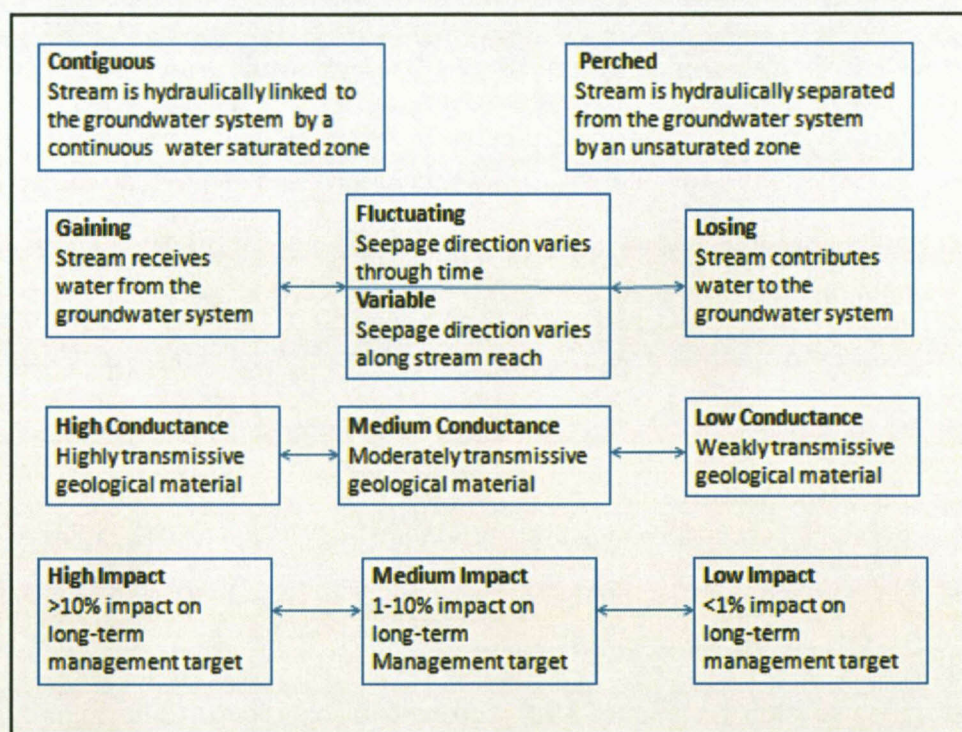


Figure 117: Categorisation of stream-aquifer connectivity (Brodie et al., 2007)

Classification may also be based on the degree of conductance (or permeability) of the geological formation, where the system is classed as being of high, medium or low conductance depending on the measure of the seepage flux. For instance high conductance geological formations would include alluvial or highly fractured rocks while mudstone or layers of low transmissivity values would be classed under low conductance category.

Connectivity can alternatively be classed in terms of potential impacts on the hydrological system and its usage. High, medium or low impact is attained depending on water quantity or quality (e.g. where nutrient-rich discharges may promote algal blooms), relative value of the resource (e.g. the only source of water or water for strategic use such as allocation for Eskom to generate electricity may be rated high impact regardless of available amount) or ecosystem dependency. It is commonly known that in South Africa the Reserve or water for ecological requirements is primed very high or the only right in legal terms, while the rest is allocated in terms of the License or under General Authorization subject to prioritization of water for strategic use and for meeting International Obligations.

5.2.2 Methods for assessing connectivity between groundwater and surface water systems

A wide range of methods available to assess the nature and degree of the interaction between groundwater and surface water are then proposed. These are conventional methods used World-wide such as

- Field observation to identify areas where interaction occurs (such as visual inspection of seepage flux),
- Ecological mapping (for instance near-stream presence of phreatophytic plants, which are deep-rooted can access groundwater thus indicating shallow water-table).
- Geohydrological / geophysics mapping (of occurrence and configuration of groundwater flow regime, or geological features that control seepage flux)
- Hydrograph separation techniques and hydrometric analysis (for determining gradient and estimating hydraulic properties)
- Isotope, chemical, temperature and tracer studies
- Hydrometry and water level gradients

5.2.3 Comparison of various assessment methods

Various assessment methods of surface water - groundwater interaction are compared in terms of spatial (local, intermediate or regional) scale, temporal (short -, medium – or long-term), costs associated with handling of data and in terms of accessibility of the technology. The advantages or benefits, disadvantages or constraints as well as the application of the method are also part of the criteria used in comparing various methodologies. The choice

of method depends on the purpose for which the measure or assessment is required as well as availability of data and other resources. The most cost-effective strategies typically involve a combination of broad scale mapping to identify sites where interaction is likely, followed by site-specific investigation to trace and quantify the flows (Ransley et al, 2007). Typically there are two fundamental types of methods used to assess connectivity – measurement techniques and modelling techniques.

5.2.4 GIS based methodology (mapping) to assess potential surface water - groundwater interaction

A GIS based method was developed to map the potential hydraulic connection between groundwater and surface water in a catchment taking hydrological and geohydrological factors into account. This method is a step preceding the quantitative assessment stage. The GIS based methodology, primarily focussed on the permeability of the geological material provides information for preliminary prioritisation of stream reaches that include identifying potential connected and disconnected systems.

The first step is to determine a stream-aquifer connectivity potential using a rating and ranking index for parameters including depth to groundwater table, geology, streambed characteristics and geomorphology. The following formula is used in a programme developed to derive the single connectivity index and then to map potential river-aquifer connectivity in an area of interest (see Figure 94):

$$\begin{aligned} \text{Connectivity Index} &= \text{Potential for groundwater-surface water connectivity} \\ &= (3 \times \text{depth to water table}) + (5 \times \text{stream bed sediments}) + (5 \times \text{aquifer material}) + (2 \times \text{geomorphology}) \end{aligned} \quad (5.1)$$

The higher the index obtained, the greater the potential for stream-aquifer connectivity. The index value output is categorised into low, medium and high connectivity potential classes based on the output classification classes in the connectivity model. These categories are then mapped to spatially represent the estimated potential connectivity along the river reaches as indicated in Figure 118.

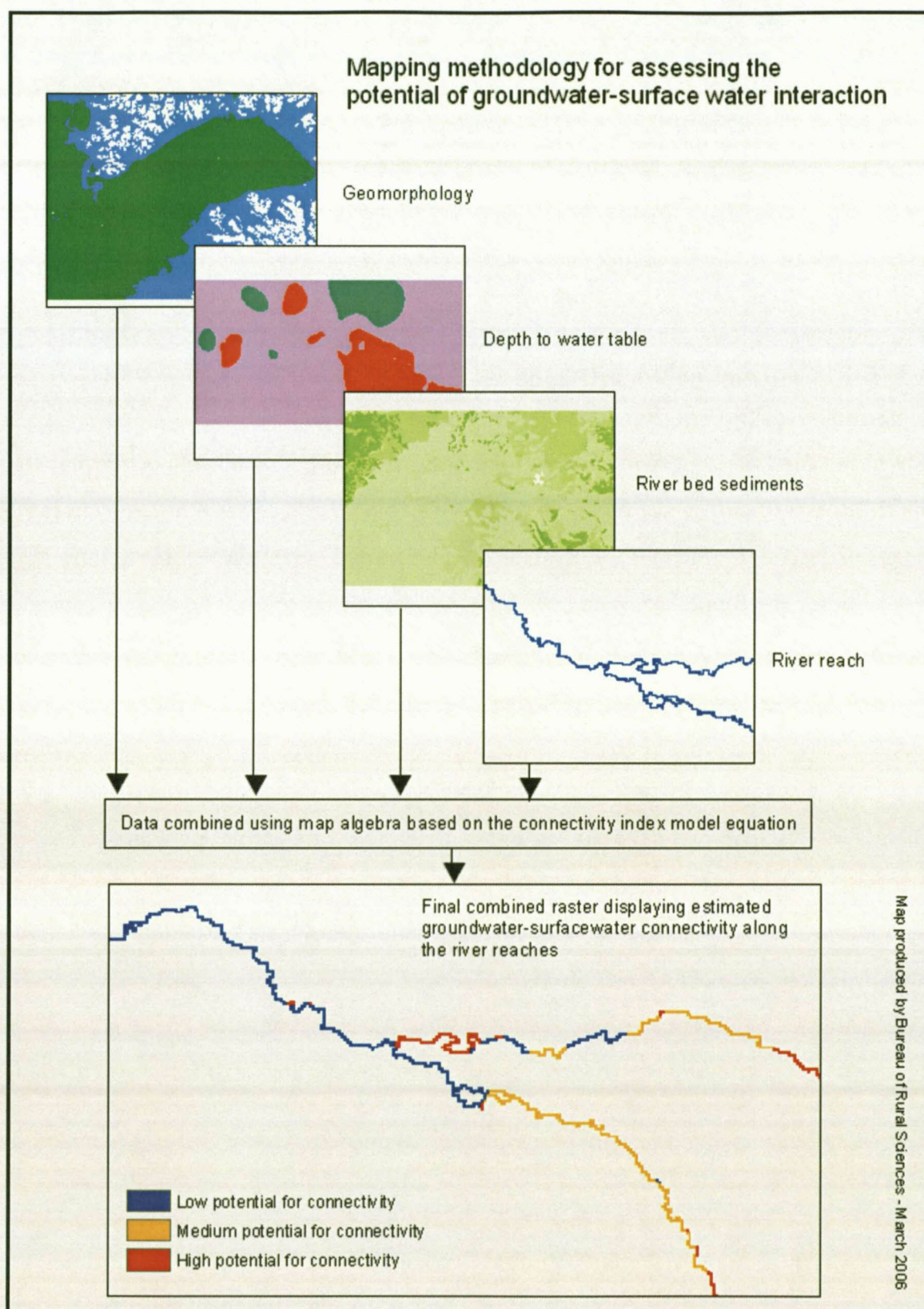


Figure 118: GIS based methodology for mapping stream-aquifer connectivity applied in the Border Rivers Catchment (Ransley *et al.*, 2007)

5.2.5 Conventional assessment models (conceptual, analytic and numerical):

A conceptual model is then developed based on data and information gathered as a basis for development of integrated surface water - groundwater interaction models (analytical or numerical models) to quantitatively evaluate the water exchange seepage flux.

5.2.6 Key aspects of the Australian framework

This was found to be one of the most comprehensive framework methodologies that were encountered during the investigation (in the literature), that together with some of the current locally developed methodologies can be a candidate framework.

The choice of methodology used in quantitative estimation of surface water - groundwater interaction is preceded and informed by a well structured and systematic series of steps within a bigger picture (broadly conjunctive water management framework). This approach wherein the connectivity of the hydrological system is classified in terms of selected parameters (such as the relative position of water table), followed by comparison of assessment methods to inform selection of the most appropriate one based on purpose and available resources is quite valuable.

Similarity in Geology between Australia and South Africa justifies consideration of this approach in this country where appropriate.

5.3 South Africa

A number of South African researchers have proposed methodologies for assessing surface water - groundwater interaction from various perspectives. Most practitioners often use hydrograph separation method as their point of departure while some prefer Darcy approach and a few others developed classification systems approach to select the method for estimation of baseflow or assessment of surface water - groundwater interaction.

5.3.1 A hydrogeomorphological approach to quantification of groundwater discharge to streams in South Africa – proposed by Xu et al., (2002)

The first step in the hydrogeomorphological approach entails construction of a conceptual model of the area in terms of geomorphologic features and hydrological regime including groundwater levels in boreholes and river stages. Then streams are classified to identify the surface water - groundwater interaction type.

5.3.1.1 Type 1: Constantly losing or gaining streams

The first category consists of upper catchment (mainly mountainous) areas that is characterised by streams without bank storage (e.g. braided rivers), with incised stream channels, which is also a regional recharge site for groundwater systems. Under this category, the surface water is constantly lost to the aquifer (i.e. losing stream) where the groundwater level constantly remain below the stream stage, or effluent conditions may prevail where the stream constantly gains water from the aquifer with the groundwater level perched above the stream stage. However, this type of interaction is dominantly characterised by interflow.

5.3.1.2 Type 2: Intermittent streams

The next category constitutes the middle courses (or intermittent) streams of pool and riffle sequences. The interaction in this category is seasonally controlled, where groundwater would typically discharge to streams during dry periods, and conversely lose water to aquifers during flooding events.

5.3.1.3 Type 3: Gaining streams (with or without storage)

The topographically flat or lower courses area which is also a regional groundwater discharge area, where streams such as meandering rivers have bank storage has groundwater level constantly perched above stream stage (typical for arid regions). Hence groundwater discharges often into the river (gaining stream).

5.3.1.4 Type 4: Interflow-dominant streams and special type

Characterised by interflow dominant typically in the upper catchment areas this type may be complicated by geology where the interaction between surface water and groundwater may be site specific.

The interaction type then informs the choice of appropriate parameter values for the equation used for estimation or quantification of baseflow in the next step. The equation is modified or adapted from Herold’s version of the hydrograph separation method. Basically groundwater contribution or baseflow is the summation of the decay of previous groundwater contribution ($Q_{gi-1} * D$) and rainfall-induced flow increment ($Q_{i-1} * I$):

$$Q_{gi} = Q_{gi-1} * D + Q_{i-1} * I = Q_{g0} * D^i + \left[\sum_{j=1}^i Q_{i-j} * D^{j-1} \right] * I \tag{5.2}$$

Where

Q_{gi} = the baseflow (is a function of groundwater contribution and rainfall induced flow);

Q_{g0} = is initial or average groundwater contribution estimated using the water table contour maps and the numerical simulation or hydrograph separation method;

I ($-1 \leq I \leq 1$) and D ($0 \leq D \leq 1$) are parameter values typically as indicated in Table 13 according to particular type of interaction;

Q_{i-j} = flow.

This method is based on the assumption that groundwater discharge fluctuation is the same as spring flows; the influence on in-stream materials and geometry is negligible; and that groundwater recharge events correlate directly to stream discharge peaks.

Recharge may be estimated using fractional characteristic method, cumulative rainfall departure method, chloride method, any appropriate or combination of various methods.

The hydrogeomorphic types are summarized in Figure 119.

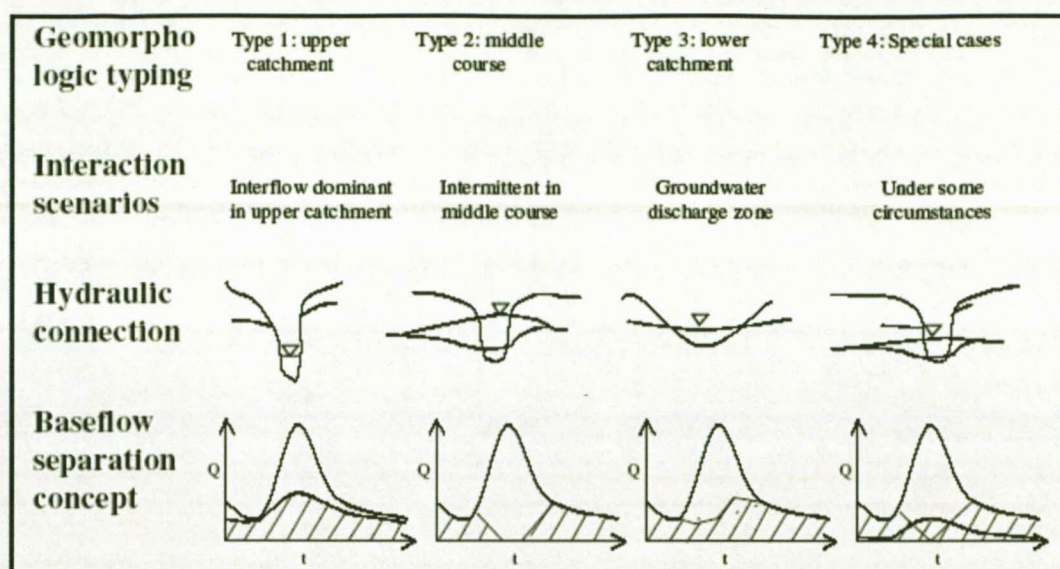


Figure 119: Hydrogeomorphic types, interaction scenarios and baseflow separation concept (Xu et al., 2002)

5.3.1.5 Part 1 - Numerical simulation to estimate initial groundwater contribution, Q_{g0} :

The area is discretised into finite difference blocks where the block size depends on resolution required. Then assign observed groundwater levels, otherwise, using values obtained through interpolation of the few available and the altitude of all observation points. This is especially applicable in semi-arid regions such as South Africa where groundwater levels often correlate to topography. Also assign other hydraulic characteristics such as transmissivity and storage values (high per block to ensure that water level does not change markedly during simulation), hydraulic conductivities as well as aquifer thicknesses to each block. Generate water-table level contour map to for steady state simulation, then perform transient flow simulation to calculate (Q_{g0}) fluxes between blocks. Known recharge may be used for calibration purposes.

5.3.1.6 Part 2 – Modified Herold’s hydrograph separation method to estimate baseflow:

On the basis of geohydromorphological classification to determine the interaction type, Table 14 can be used to estimate D and I parameter values, and having calculated the value of Q_{g0} in part 1, equation (5.2) and the simple excel spreadsheet programme available at the site developed by Xu et al.,(2002).

Table 14: Types of interaction between groundwater and rivers (adapted from Xu *et al.*, 2002)

Types of interaction between groundwater and rivers					
Type	Nature of interaction	Geomorphological type	Q_{g0}	D	I
1	Constantly losing streams (constantly gaining streams)	Upper catchment	-, (+)	1, (1)	0, (0)
2	Intermittent streams	Middle course	+, -	+	-
3	Gaining stream as groundwater level is above river stage (with storage)	Lower course	+, (+)	1, +, (<)	0.34 – 1, (0.01- 0.33)
4	Interflow dominant streams	Special type	0	0-0.5	+

This approach is an improved version of Herold’s method that appropriately takes account of the geologic, geomorphic and hydrological setting of the system. Most probably further refinement of this method where other key drivers such as climate, the role of riparian vegetation, land-use impacts and other are taken into account, could be pursued in future version hereof. Otherwise, this approach is recommended as one element in the toolbox of methodologies appropriate for South African conditions, especially where there is paucity of data and estimate of groundwater fluxes into streams are required.

5.3.2 Dennis and Witthueser’s Classification System for surface water - groundwater interaction

The objective of this classification system is to serve as a tool to guide the user on selection or choice of an analytical or numerical model that is appropriate for assessing surface water - groundwater interaction. The researchers started by reviewing current classification systems developed internationally and nationally, considering the pros and cons of each as part of the process and then subsequently proposed the new system.

Key features or geohydrological regimes that are taken into account include the presence of clogging layers in riverbeds, hydraulic gradient or relative stage of the river and level of groundwater and geohydrological characteristics of strata along the stretch of the river.

These features were then worked on current classification systems by Pitman and Vegter, as well as the approach used by the Environment Agency. A two tier systems was developed and is discussed in the following sections:

5.3.2.1 Primary classification

The first step entails geological classification of the river-aquifer setting with differentiation between the aquifer-type in which the river flows (i.e. porous, bedrock or valley train underlain by aquitard) as shown in Figure 120. Then subdivide the porous media into alluvial sediments and weathered hard rock in which case checking whether the layer at interface is impervious, semi-pervious or if the connectivity would be direct. If impervious layer is present then there would be no connectivity and hence no interaction, whereas in case of semi-pervious layer either analytical or numerical method would be applicable. For direct contact, also either the analytical or numerical method may be used but there would be no need for the leakance factor. The bedrock media includes hard rock, dolomites and hard pans. The interaction would either be absent, localised or continuous along the river stretch. If localised then a numerical model is used while for continuous either analytical or numerical model is used. The first tier, primary classification basically identifies whether surface-groundwater interactions must be considered and narrows down the applicable mathematical models for particular hydrological settings.

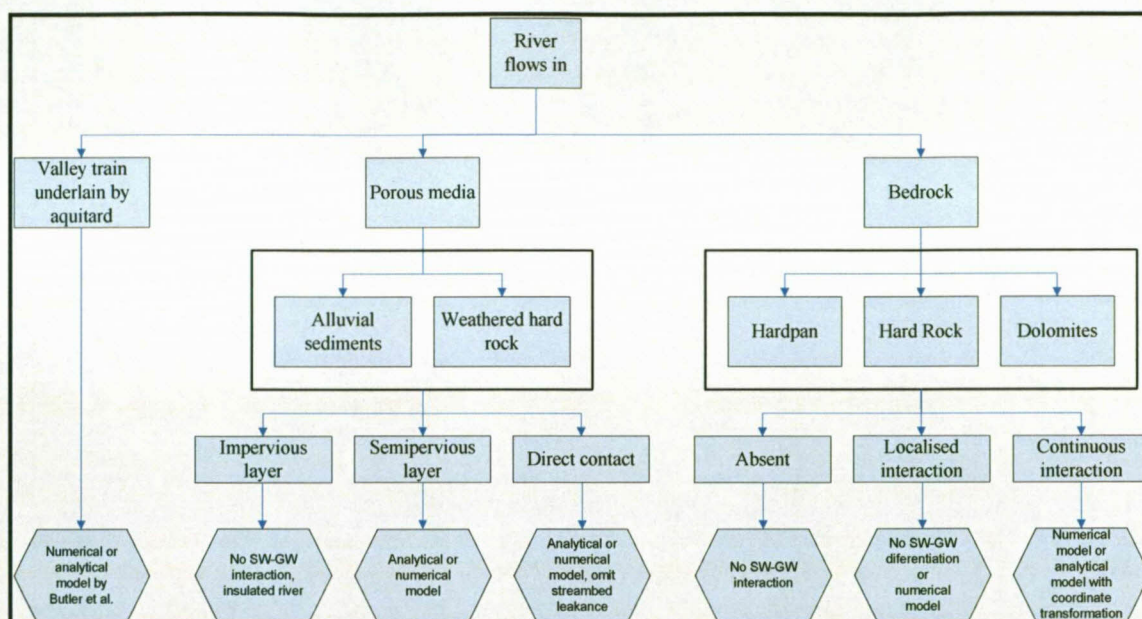


Figure 120: Primary geological classification scheme for surface water - groundwater interaction assessment (Dennis and Witthueser, 2007); same as Figure 88

5.3.2.2 Secondary classification

Then the second tier then entails classification of the interaction according to the prevailing hydraulic gradient and consideration of potential clogging that may impede water exchange fluxes (Figure 121). Basically, depending on whether the river is gaining (effluent), losing (influent), seasonal alternating gaining or losing conditions, or through-flow. For effluent conditions numerical model or analytical model by Chen and Yin is used while for alternating conditions a numerical model or analytical model is applicable. In case of through-flow only a numerical model is recommended. In case of a losing river a further subdivision into either perched river or connected river. For the latter case either a numerical or analytical model is used whereas for a perched river a numerical model or analytical model by Bower would be applicable.

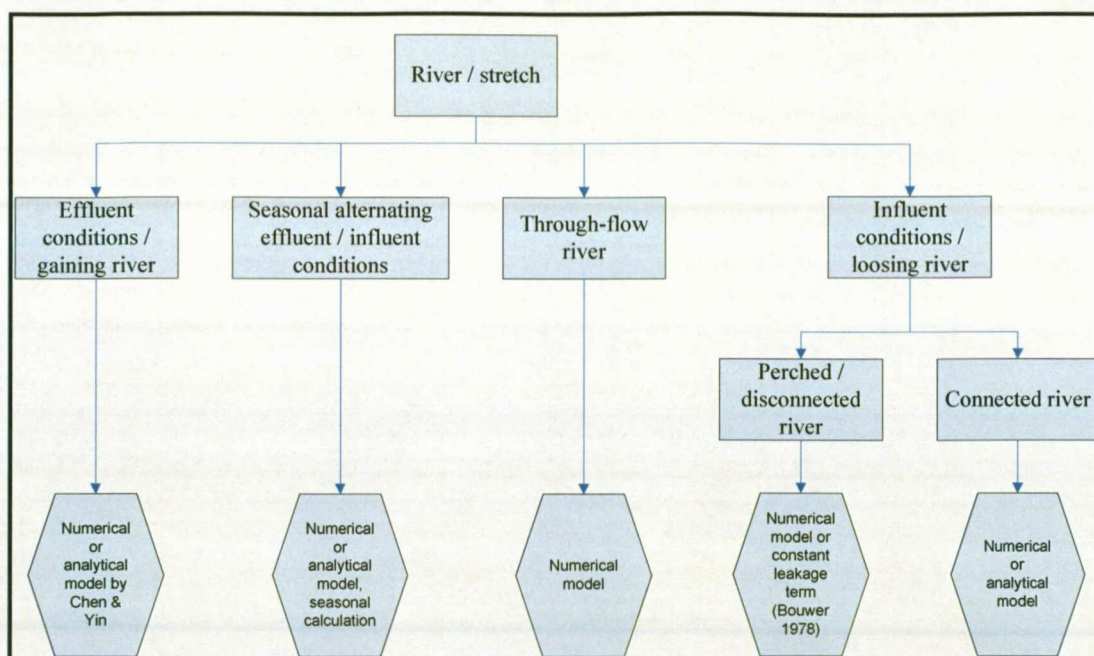


Figure 121: Secondary hydraulic classification scheme for surface water - groundwater interaction assessment (Dennis and Witthueser, 2007); same as Figure 89

5.3.3 A software program for simulation of surface water - groundwater interaction by Sami (2006)

South Africa's Department of Water Affairs undertook a project that was part of the Phase 2 programme to review methods on quantification of surface water - groundwater interactions. Sami (2006) was then tasked to develop a generic algorithm that can be applied to estimate surface water - groundwater interaction nationally at a quaternary catchment scale. Microsoft EXCEL spreadsheet based software was then designed to simulate the potential impacts of groundwater abstraction on baseflow for incorporation into the Water Resources Yield Model (utilised in the planning processes by the Department of Water Affairs) in which then, only surface water abstractions were considered. Modifications were also made to the software to enable incorporation of the model into WR2005 (i.e. the 2005 version of the Surface Water Resources of South Africa system), and the algorithms are coded into the Pitman model in WR2005 in order to simulate baseflow at national scale.

5.3.3.1 Equations and parameters used in the model development

The derivation of equations is not done in this research investigation but reference (Sami, 2006) is provided for further details in this regard.

5.3.3.2 Model assumptions

The proposed model for surface water - groundwater interaction is based on the following assumptions:

- Baseflow depletion due to groundwater abstraction and groundwater outflow from the catchment is calculated using a Darcian approach, (i.e. flow in a porous media or primary aquifer is assumed in the application of this model).
- The model is not validated for a fractured/secondary or karstified aquifer system.
- The baseflow depletion calculation assumes that water is withdrawn only from the regional aquifer, but not from the perched aquifers.
- Assessment of the impact of a single groundwater abstraction point on baseflow is not applicable because the baseflow depletion calculation uses the weighted mean distance of abstraction points from the main channel.
- The hydrogeological parameters of the model are determined with water balance approaches and averaged over a Quaternary catchment scale.

5.3.3.3 Estimation of soil moisture

The catchment soil moisture time series (Pitman S) generated by the WRSM 2000 (i.e. the revised version of the Pitman model (i.e. a monthly time-step rainfall-runoff model – representation thereof is attached in Annexure) utilised in Surface Water Resources of South Africa 2000) is used in the methodology to calculate a time series of recharge.

Interflow (Q_g) is generated using the Pitman algorithm; and then the soil moisture (S) calculated from the following equation:

$$Q_g = FT \left(\frac{S-SL}{ST-SL} \right)^{POW} \tag{5.1}$$

Where,

S = variable of Pitman S (subsurface moisture storage in mm) for each month

- POW = parameter of the ratio of actual soil storage to storage capacity
- SL = parameter of minimum soil moisture storage below which there is no runoff
- ST = parameter of maximum soil moisture storage
- FT = parameter of the maximum baseflow expressed as a depth

Parameters for SL, ST, FT and POW are obtained from WR90 (the 1990 version of Surface Water Resources of South Africa study).

5.3.3.4 Calculation of recharge

Once soil moisture is obtained, monthly recharge is calculated using the following equation:

$$RE = \left(\frac{S - SL}{ST - SL} \right)^{GPOW}$$

Where,

- RE = variable of potential recharge (mm)
- GW = parameter of maximum recharge in mm at maximum soil moisture (ST)
- S = Input data of soil moisture in mm
- SL = Parameter of soil moisture threshold below which there is no recharge
- GPOW = Parameter of the storage-recharge relationship

The SL parameter controls the soil moisture threshold below which there is no recharge. The author also indicated that parameters for GW and GPOW could either be calibrated to achieve a fit with long term mean annual recharge measurements obtained from other methods, or initially parameters similar to POW and FT could be selected, since the parameters have similar bases. He also indicated that GPOW would lie between 1 and 3. Recharge from soil moisture (RE) is then added to a percolating storage zone defined by a parameter PMAX (mm), where its transmission to the aquifer is attenuated by:

$$PERC = RE_x \times \left(\frac{P}{PMAX} \right)^{PPOW} \times \frac{RE_x}{RE}$$

Where:

PERC = variable of percolation from the percolating store to the aquifer storage

RE_x = variable of the moving average of recharge (RE) for x months (1-120 months)

P = variable of percolating storage

PPOW = parameter of the relationship between storage and percolation (<1)

PMAX = maximum capacity of percolating storage

RE = mean monthly recharge

PMAX can be calculated as the mean water strike depth multiplied by storativity.

5.3.3.5 Estimation of the evapotranspiration

The monthly evapotranspiration is calculated as the product of mean annual evaporation, monthly distribution and crop factor. Rainfall is then subtracted from evapotranspiration to obtain evapotranspiration demand from groundwater. When rainfall exceeds evapotranspiration demand evapotranspiration from groundwater is defaulted to 0, since it would then be assumed that the evapotranspiration demand will be met from soil moisture storage.

Evapotranspiration demand is multiplied by an aquifer storage factor to allow evaporation to decrease as groundwater storage is depleted. Evapotranspiration occurs at the maximum rate when groundwater storage is at or above total aquifer storage (TAS) and declines towards 0 as groundwater storage drops to a level below the stream channel, defined by a parameter of static water level.

Evapotranspiration from groundwater is therefore calculated from the following equation:

$$[(MAE \times MDIST \times CROP) - RAIN] \times AREA \times \left(\frac{STORE - SWL}{TAS - SWL} \right)$$

Where:

MAE	=	mean annual evaporation
MDIST	=	monthly distribution fraction of evaporation
CROP	=	monthly A pan crop factor for appropriate Acocks vegetation cover
RAIN	=	input data of monthly rainfall
AREA	=	riverine area where evapotranspiration from groundwater can take place
SWL	=	parameter of static water level
TAS	=	total aquifer storage from GRA II (Groundwater Resource Assessment, Phase 2)

5.3.3.6 Estimation of groundwater contribution to surface water or baseflow

Groundwater baseflow (GWBaseflow) and transmission losses are calculated using a non-linear equation to account for the effects of hydraulic resistance:

$$GWBaseflow = (1 - e^{(HEAD \times BPOW)}) \times BFMAX$$

Where,

BFMAX	=	parameter of the maximum rate of groundwater baseflow
BPOW	=	relationship between head difference and baseflow

$$HEAD = STORE - SWL - \left(\frac{RUNOFF}{CATCHMENT}\right)$$

Where,

RUNOFF	=	Input of streamflow
CATCHMENT	=	Catchment area

The parameters BFMAX and SWL are calibrated by verifying that groundwater baseflow approximately equals total streamflow during the driest period on record. Where no interactions occur, BFMAX is set to 0.

This equation allows large increases in baseflow or transmission losses for small head changes when the head difference between surface and groundwater is small, but causes baseflow and transmission losses to approach the maximum value of BFMAX as the head difference becomes larger. As the head difference increases, the exchange of water thereby increases at an increasingly smaller rate.

5.3.3.7 Structure of the model

Figure 122 represents the structure of the model.

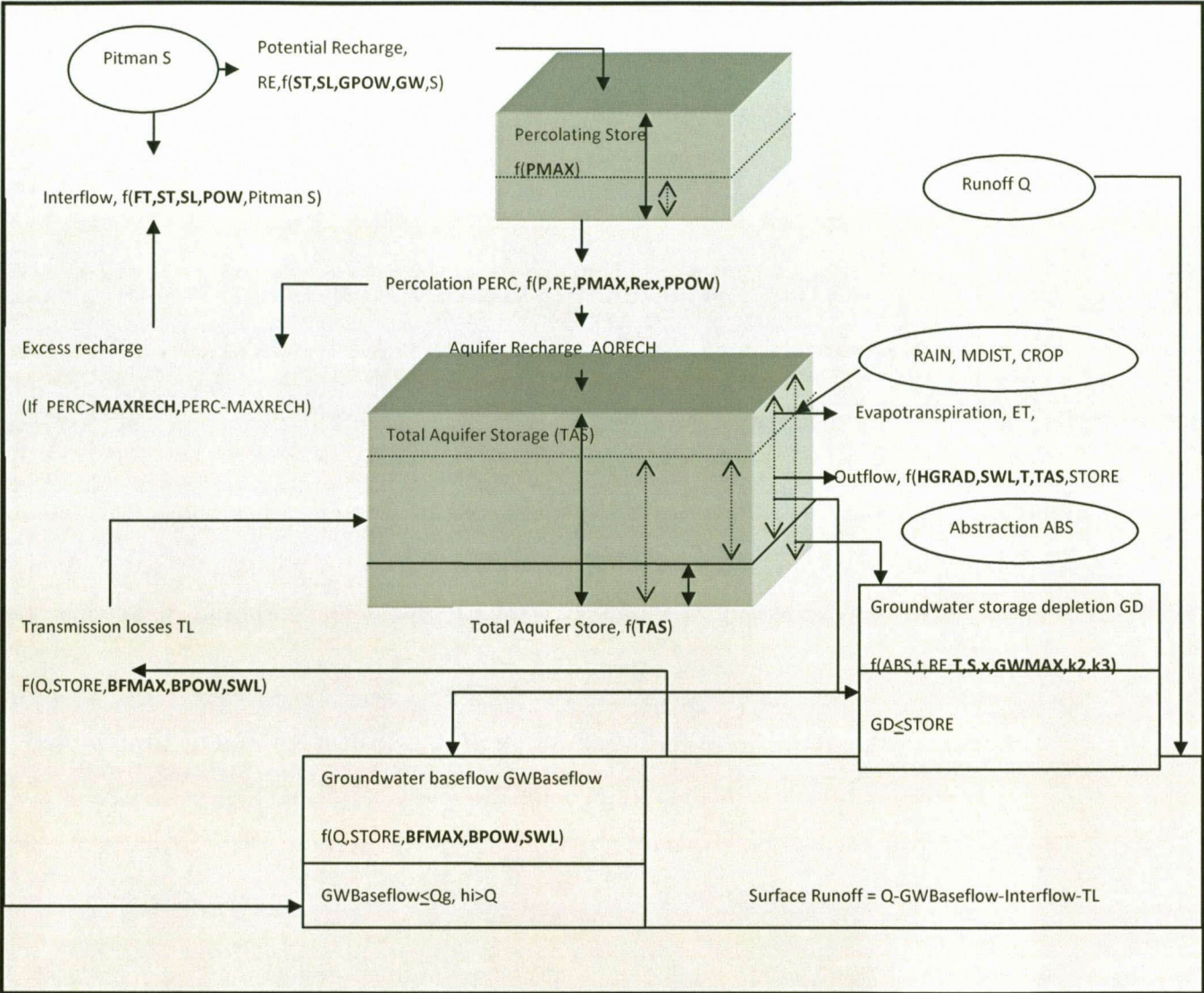


Figure 122: Structure of the model software (Sami, 2006)

5.3.4 Remarks regarding the application of the model

This model by Sami (2006) is still work in progress and is not recommended for use at this stage unless the following concerns are addressed. Firstly, most of the South African aquifers are secondary or fractured type and yet as indicated among the model assumptions, the model has not been validated for fractured aquifers. The other reason is that this model was initially successfully tested in the Schoonspruit Catchment (Mare et al, 2007), but the test could not successfully be replicated in follow-on tests using different datasets or study areas. The plausible explanation is that since some parameters were

determined by water balance and calibration approach; the results were case area specific and thus could not generically be applicable to other areas. Also, a number of surface water - groundwater interaction case areas are either characterised by perched aquifers or rivers, yet one of the assumptions is that groundwater withdrawal is only valid for regional and not perched aquifer. Otherwise this methodology has potential to make valuable contribution.

5.4 Framework for assessment of surface water - groundwater interaction

In this section a framework for the purpose of informing appropriate methodology for assessment or quantification of surface water - groundwater interaction is developed. Alternatively, in the researcher prefers to use only either the numerical or analytical method to evaluate surface water - groundwater interaction, then the tables (i.e. Table1 under Section 2.2.14; Table 2A and 2B under Section 2.4) that were proposed earlier may be utilised for decision making in this regard.

5.4.1 Development of the conceptual model of the study area and establishment of whether the water exchange is likely to occur or not

The first step is to develop the conceptual model of the system or the study area. Key questions could include the following:

- How can the study area broadly be described given the baseline and available desktop information? (This also typically includes previous investigators' reports)
- What is the hydrological setting of the area in terms of water flow regime, water levels and climate based on available data and information?
- What important features can broadly be observed (topographical, geomorphic, structural, geological, etc) in the study area?

Hence this step should include capturing (geophysical mapping) and collation of baseline data and information (e.g. geo-structural features, landscape, hydrometric data or water levels for determination of the hydraulic gradient and groundwater flow direction relative to that of surface water) and flows. Use the analysed data and information to broadly describe the study area (refer to Figure 123). Then description of the geohydrological setting

(including geology, structure and prominent features) and mapping of the area follows on. For instance, regional changes in groundwater level in space and historical surface water flows during wet and extended drought periods would be very useful in this process. Synthesis of the information gathered at this stage should lead to a broad description of the area.

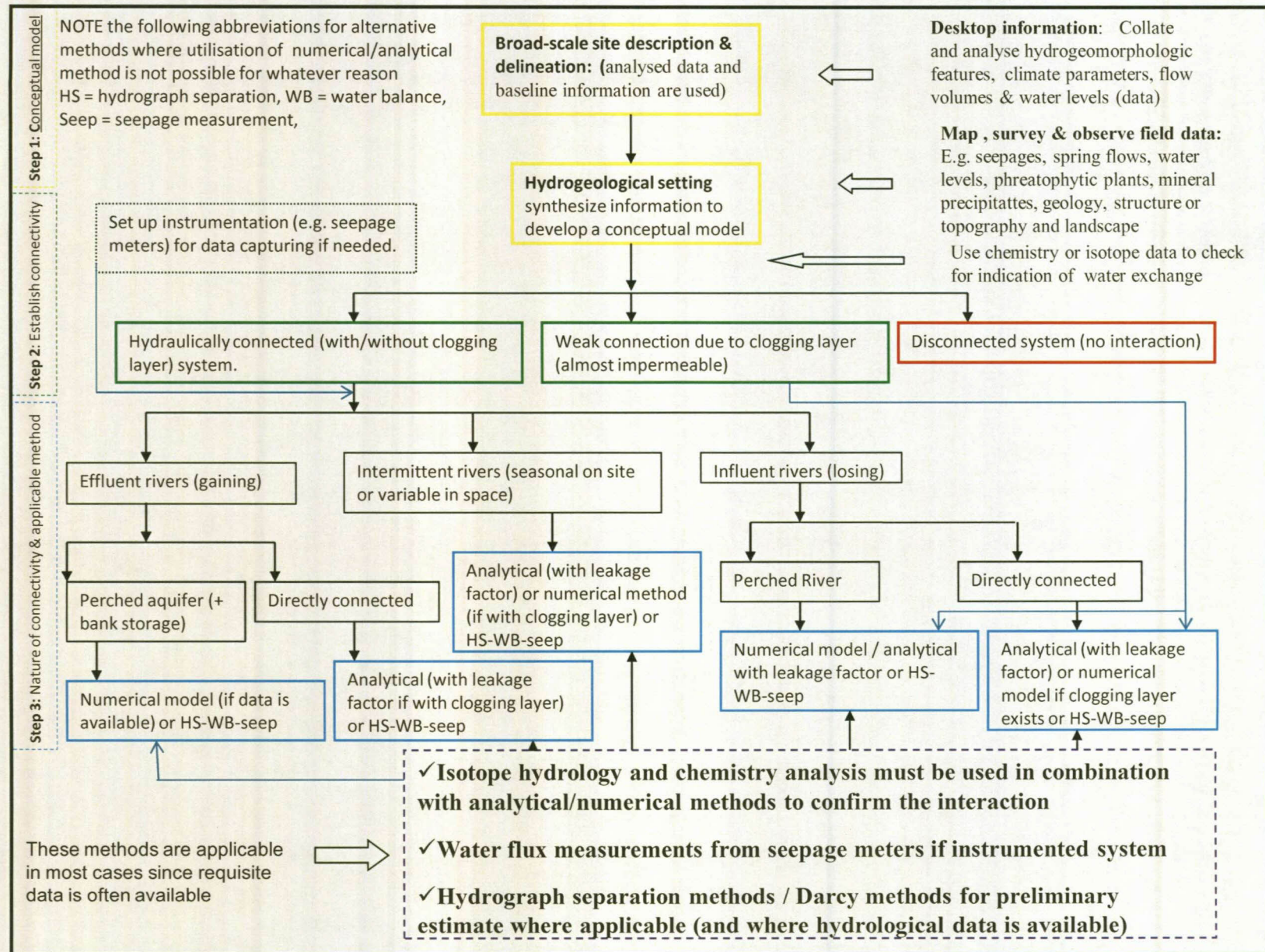


Figure 123: The framework for assessment of surface water - groundwater interaction

Then a site visit to the area to perform to observe and capture information or use visual field (e.g. river water and groundwater level measurements and relative positions of boreholes in relation to the river within the riparian zone to determine potential groundwater flow direction) indicators. The activities would typically entail a general observation of the landscape/topography, the riparian vegetation including phreaphytic plants (i.e. deep rooted plant that gets water from the saturated zone and exist due to presence of shallow groundwater), observing and identifying location of potential seepages areas along the river (either by visual inspection of seepage meter measurements if these are installed). Utilise the information to determine the hydrological setting of the area and a conceptual model of the system.

5.4.2 The interface or connectivity between surface water and groundwater

The next question is whether there are any indications of interconnectivity between surface water and groundwater. In other words, does the conceptual model indicate that surface water - groundwater interaction is likely to occur, is occurring or has occurred? The evidence or proof of interaction can be signified or shown by a qualitative method of assessing water exchange such as seepage flux observation in the field, chemical analyses, isotope hydrology techniques or tracer testing methods. On the other hand, quantitative approaches of assessing the interaction such as seepage flux measurements, hydrograph separation method, numerical modelling or water balance techniques are used to either estimate baseflow or recharge in the case of losing streams. If field mapping or visual inspection for seepage flux does not yield any evidence of interaction then chemical analyses or environmental stable isotope hydrology method may be used to confirm this conclusion or establish connectivity. Otherwise, the assessment of the interaction does not need to be undertaken any further beyond this stage.

5.4.3 Gaining, losing or intermittent rivers and applicable methods

The river could be effluent (gaining), influent (losing) or intermittent (where the conditions are either seasonally gaining/losing at a point or variable in space along the river stem in which case the river gains at certain places and lose at others).

For effluent conditions, the aquifer may either be perched or directly connected as indicated in Figure 123. A perched aquifer consists of an unconfined body of water that is separated above the regional water table by an unsaturated zone, at times enveloped by a lens of semi-permeable material as well. Trickles of water may be discharged into underlying layers through the unsaturated zone and part thereof eventually reaching the river. The numerical method may be used depending on availability of data. Otherwise, the analytical model used should take account of a leakage factor that takes account of the thickness and hydraulic conductivity of the intervening materials. Either Stang and Hunt or Chen and Yen's solution would be suitable. In cases where even the analytical method would not do, for whatever reason, the hydrograph separation method, water balance methods and seepage measurements may be used to quantify the interaction. Additionally, the chemistry data, isotope data, radon data, temperature data or artificial tracers should be used as indicators of the interaction for confirmation purposes or otherwise.

For the direct connectivity, either a numerical model or analytical model should be suitable. However, in all cases assumptions for an analytical model used should be considered against the geohydrological setting, including hydraulic characteristics. Again if neither the numerical nor the analytical method applies in a particular case, alternative methods of estimating baseflow in quantitative terms (e.g. seepage flux measurements) should be used.

For influent conditions where the river is in direct contact with the aquifer, either a numerical or analytical model may be used to assess the interaction. Otherwise, if paucity of data or any reason makes it difficult to use these models, then seepage measurements, hydrograph separation or water balance methods should be used together with chemistry, isotope hydrology or tracer methods.

Under influent conditions rivers may lose water to aquifers, in which case the river is either directly connected to the aquifer or perched. For directly connected systems, either the analytical model with a leakage factor may be used. Otherwise, a numerical model may be used applicable if a clogging layer exists. In cases where data is inadequate or not available alternative methods become relevant. If influent rivers are perched then the river and the aquifer would either be disconnected or weakly connected. In the latter case an analytical or numerical model may be used to assess the interaction, while in the former case there

would be no need for further assessment. It should also be noted that there is a limit to which recharge from the stream would be induced. In a perched stream, when the depth to the water table below the stream is more than 5 times the maximum depth of water in the stream, then further drawdown from pumping will not induce seepage from the stream (Pattle Delamore Partners, 2000). However these empirical factors, originally based on the net flow analysis by Hunt (2007), should be used with caution, bearing in mind the basic assumption that allows for it. For instance, in cases of preferential flow path in the unsaturated zone separating the 2 systems then this argument would no longer be valid.

Intermittent rivers either lose or gain water in spatial or temporal domain. During a wet or dry season the river may on a particular site gain or lose water respectively (seasonal intermittent rivers), while the river may gain water from the aquifer at what site along the stream and lose water to the aquifer at another part of the reach along the stretch (spatially variable) once more in this case the analytical or numerical models may be used to assess the interaction between groundwater and surface water.

It is noteworthy that the chemistry or isotope hydrology method should be used either as substitute to relatively data intensive methods such as the numerical methods (especially where data is inadequate) or in combination therewith to confirm the interaction. Where interaction hot spots were mapped or identified, seepage meters should be installed to measure baseflow estimates. The captured data may be used later for management purposes.

In conclusion of this chapter, the proposed framework is deemed generic and may be used to decide on the appropriate methodological approach under consideration.

5.5 Application of the developed Framework using one of the case studies - The Mokolo catchment case study

Part of the Mokolo Catchment along the river stretch is used to illustrate how the process or approach to surface water - groundwater interaction should be handled leading to the selection of the appropriate method or model.

5.5.1 The broad description of the study:

The first step involves collection and collation of data and information as well as desktop analysis and interpretation. In this regard, the geology, structure, topography and climate of the study area have already been described Chapter3, sub-section 3.5.2 to 3.5.4. Geohydrological setting and conceptual model of the entire system has also been described in this case study are before. At this stage a site visit to carry out field observations and collect data on evidence of seepages and other 'tell-tales' such as aquatic plants, or mineral precipitates along areas where leakages possibly occurred before would be appropriate following the desktop study.

5.5.2 The nature of connectivity between surface water and groundwater

It should suffice to re-state that the groundwater level in the alluvial aquifer (about 9 m thick thinning downstream) that is being tested for interaction with the Mokolo River is influenced by the dam water release, hence by the fluctuation of water level in the river (as evidenced by Figure 124, same as Figure 82 in section 3.5). Thus, there is an indication that the surface water and groundwater in the alluvial aquifer are directly linked and interact.

It is also clear that most of the time the water level in the river is higher than that in the aquifer which implies that the river is losing water to the aquifer. This point is also confirmed by the fact that the pool also loses water to the aquifer as indicated before in this case study. In this particular case the framework indicates that, since the river is losing water to the aquifer and directly connected the analytical model is applicable. There is no evidence of a clogging layer and hence a leakage factor may not be necessary at this stage. According to a decision Table 1 the appropriate analytical model would be Theis/Glover & Balmer, or Jenkins/Wallace et al., or DiMatteo & Dragoni. It should also be emphasised that this situation applies mainly at that particular area of interest and at the time when the observation is made.

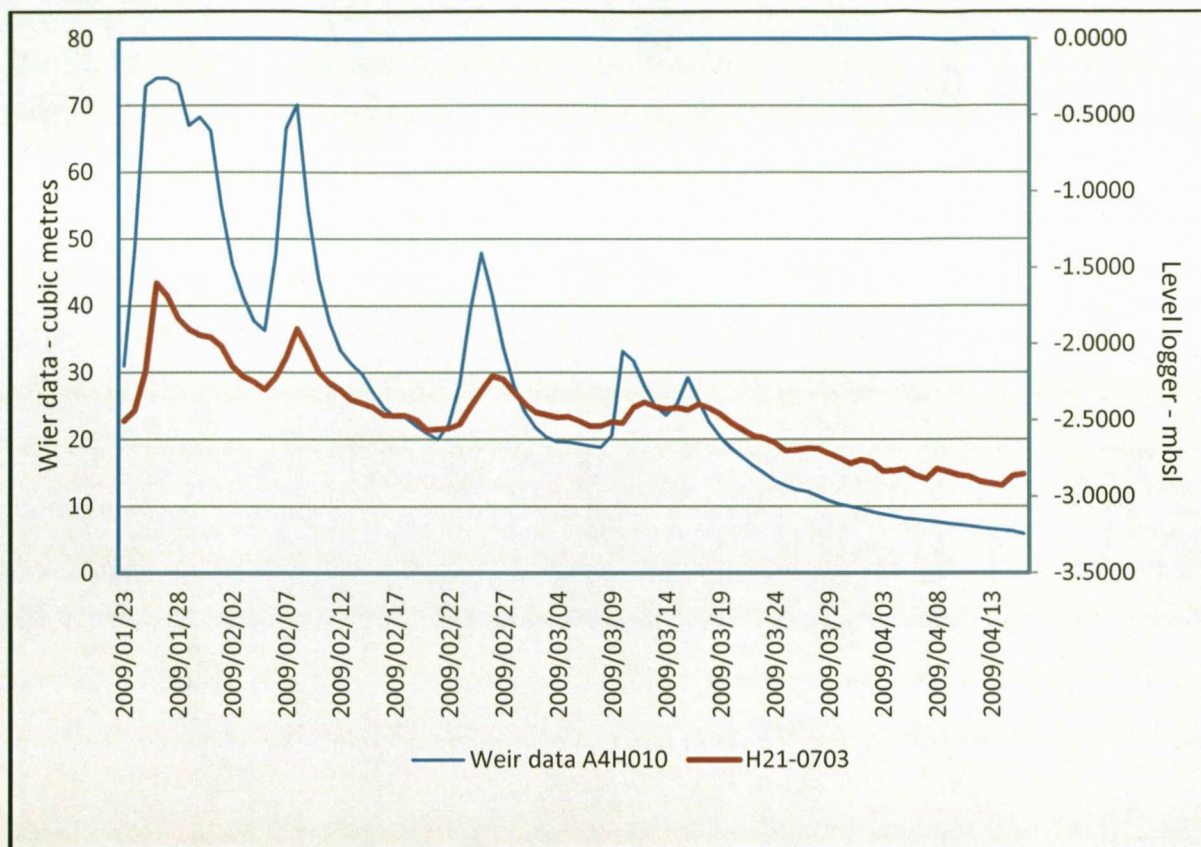


Figure 124: Water level fluctuation in the alluvial borehole (H21-0703) and weir data at observation point (A4H010) by Dept of Water Affairs², (2010) – same as Figure 82.

However, this fluctuation occurs seasonally as indicated in the figure, and hence we are dealing here with the intermittent connectivity between surface water and groundwater where depending on whether dam water has been released or if it rained or not the river either gains water from or loses water to the aquifer. Hence according to the framework an analytical method is applicable here as well. However, if the interaction between the pool and the alluvial is considered, use of either an analytical model that takes a leakage factor into account or a numerical model that considers a clogging layer is used.

5.5.3 The chemistry and isotope data analyses and interpretation as alternative methods and to confirm the interaction

The chemistry and isotope data analyses, as before in the case study, which need not be repeated, did confirm the interaction between Mokolo river water and the groundwater of the alluvial aquifer. Sub-section 3.5.6.4 in Chapter 3 has relevant details in this regard.

6 Conclusions

6.1 Findings and knowledge contributions

The literature study revealed that despite numerous methods and approaches utilised Worldwide for assessing and in some instances quantifying surface water - groundwater interaction, there is no one size fits all type of toolbox that can be used and provide precise answers. However for any method to be applicable to assessment of the interaction a number of factors, the conceptual model of the system and availability of data, are vital. South Africa is a relatively dry country with various types of aquifers ranging in complexity from intervening clogging layers of limited extent to directly connected or disconnected (i.e. poorly connected) connected systems. Hence, the choice of a particular method requires conceptual understanding of the system and use of some framework.

The hypotheses that in spite of being treated as separate entities, surface water and groundwater interact to a varied degree and at various spatial and temporal scales was a reasonable assertion as confirmed through the literature review and case studies. In fact in some cases the interaction between the two systems is undetectable or poorly detectable yet real. Reconsideration of the objectives that were originally set, following the unfortunate unavailability of data and subsequent decision to focus on investigating and ultimately identifying methodologies that are appropriate for South African conditions was also a reasonably good move. This research investigation has the following contributions to make to the water sector.

Firstly, for assessment of surface water - groundwater interaction, estimation of baseflow contribution to surface water or even just to establish whether water exchange occurred (or occurs) between the two systems (say for resource protection purposes); the proposed approach is to use more than one methodology in order to make conclusive statements about the interaction. It is common practice for most practitioners to start and end with the hydrograph separation method when dealing with baseflow estimations. It is however, crucial to use at least an additional method in order to confirm whether interaction is occurring, has occurred or not. In addition, the advantage of utilising multi techniques is that the confidence level of the results is enhanced or that uncertainties are reduced.

The other contribution is that decision tables have been developed to guide choice of appropriate model to use given associated assumptions for such a model and the conceptual model of the study area or the system in question. Again in practice it is not uncommon for one to only be guided by what dataset is available and mostly what methodology one is comfortable with. This practice is not prohibitive but what is most important is for one to be aware of assumptions associated with the model of choice and hence be prepared to live with the consequences albeit unintended.

The developed framework for guiding the process of undertaking surface water - groundwater interaction is also a contribution from this research investigation. The structure is not markedly different from those by previous researchers but there is also some additional proposal for alternatives to analytical and numerical models. This is important because the reality is that adequate data are seldom available and not always accessible. Among all methods the chemistry is currently under-utilised; especially the radioactive radon noble gas that can be valuable in assessments of surface water - groundwater interaction. Hence a further proposal that "new data" (i.e. data that is not conventionally collected) should be part of regular monitoring data.

The other lesson learnt from this investigation that is a contribution to science is that one size fits all approach should be avoided. Each of the 5 case studies used is distinct in terms geological, structural and climatic controls and associated water flow dynamics. For instance ordinarily it would be easy to underestimate the role of hillslope hydrological processes in surface water - groundwater interaction. However, on closer inspection, particularly in a dry country such as South Africa with perched aquifer systems, understanding these processes with respect to baseflow is very important.

Finally, the surface water - groundwater interaction is a very complex process that is not static but dynamic in time and space. Upon working on an area and then finding out that for instance the river reach (i.e. particular stretch or section) at that particular moment is either losing water to or gaining water from the aquifer, one should be cautious of the dynamic nature of the process. It is possible that the type of interaction is seasonal (i.e. depends on whether it is a raining or dry season), or that downstream the situation is the

other way round (i.e. the type of interaction varies in space along the same river. This is also the case in the Mokolo River example.

6.2 Conclusions

In conclusion, the framework developed in this investigation to inform decision processes in selection of appropriate methodologies for evaluation of surface water - groundwater interaction under South Africa's geological and hydrological conditions is useful. There are several methods and classification systems available in the literature nationally and globally on how to deal with surface water - groundwater interaction. The variation in scale (i.e. local, catchment or regional), complexity (i.e. level of user friendliness) and costs (i.e. time and data requirements) among available options can be overwhelming. The usefulness of each particular model, system or methodology depends on what the user need (i.e. the purpose of the investigation) and on the conceptual model of the study area. The objectives of this investigation, namely to develop the framework for assessment of surface water - groundwater interaction as well as identification of appropriate methodologies for South African conditions were achieved.

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8. Annexure

8.1 Annexure 1: Monthly rainfall for Richmond area from Nov 2004 to Jan 2005

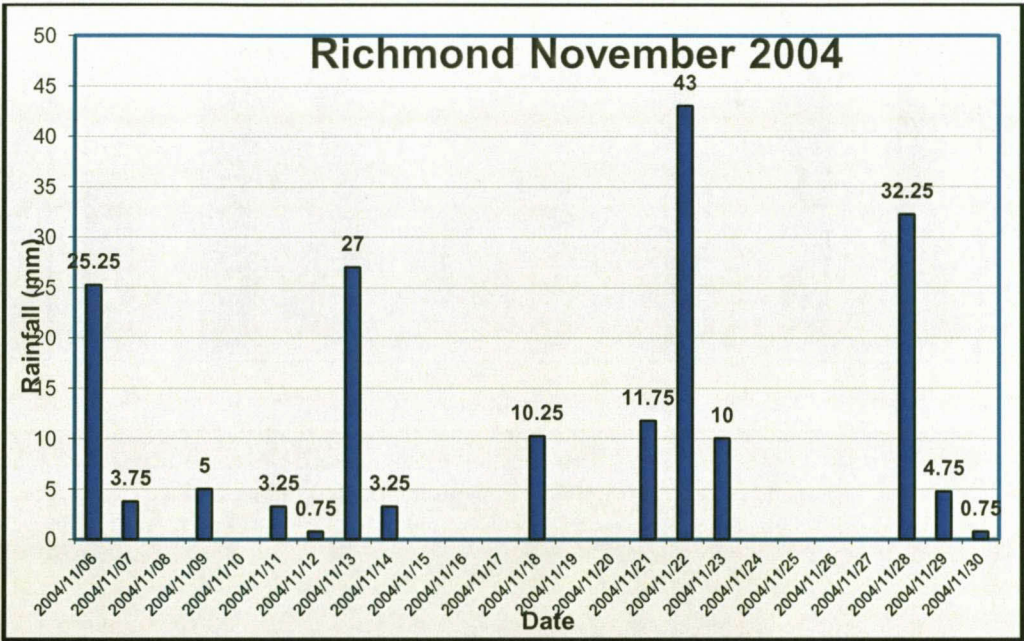


Figure 125: Average rainfall in Richmond area in November 2004

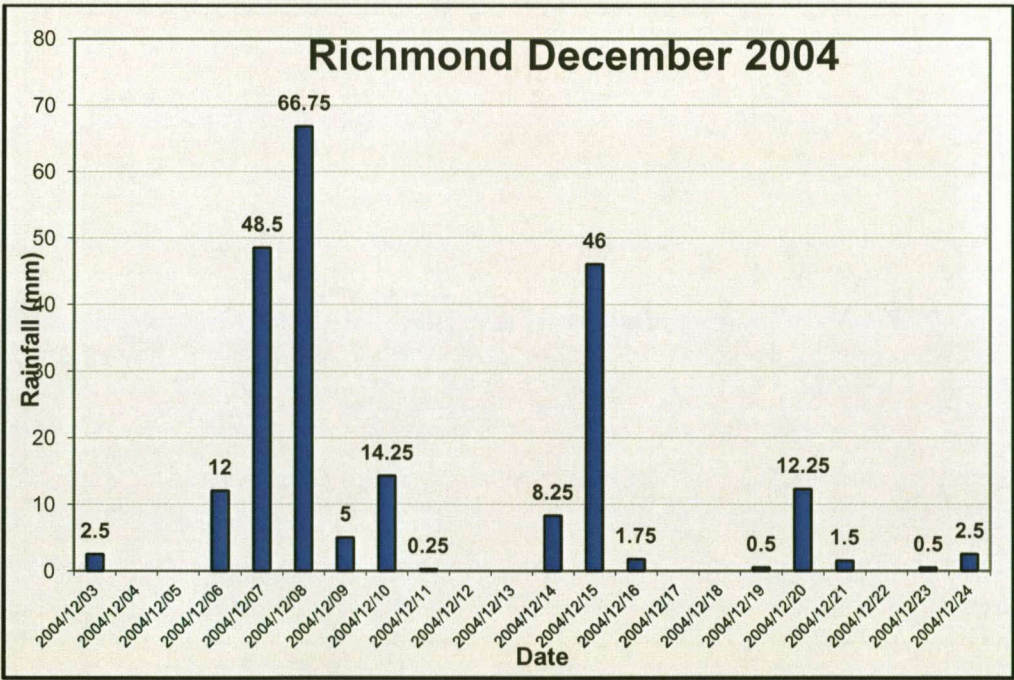


Figure 126: Average rainfall in Richmond area in December 2004

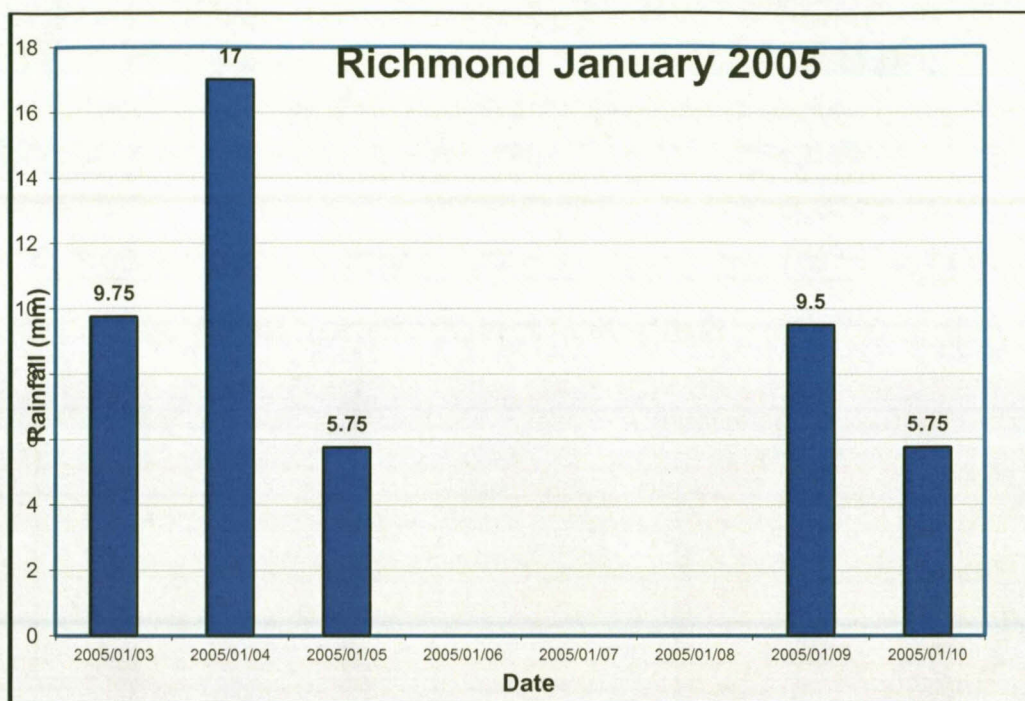


Figure 127: Average rainfall in Richmond area in January 2005

8.2 Annexure 2 Drawdown curves for pump tested boreholes in the Richmond area

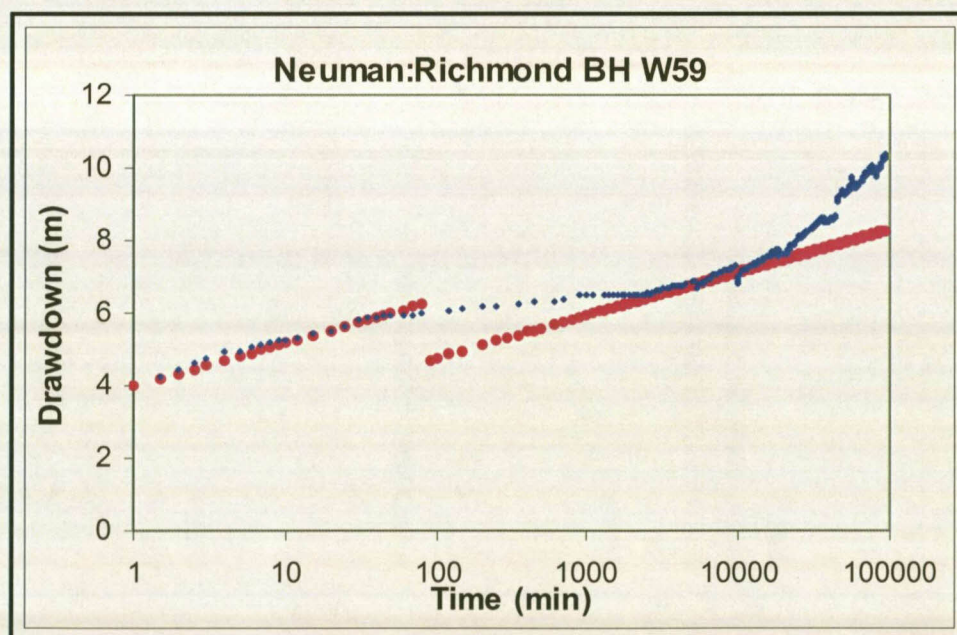


Figure 128: Drawdown curve for borehole RMGW59 which shows a water table aquifer and then a zone with much smaller T-value after 19 000 minutes (van Tonder and Dennis, 2005)

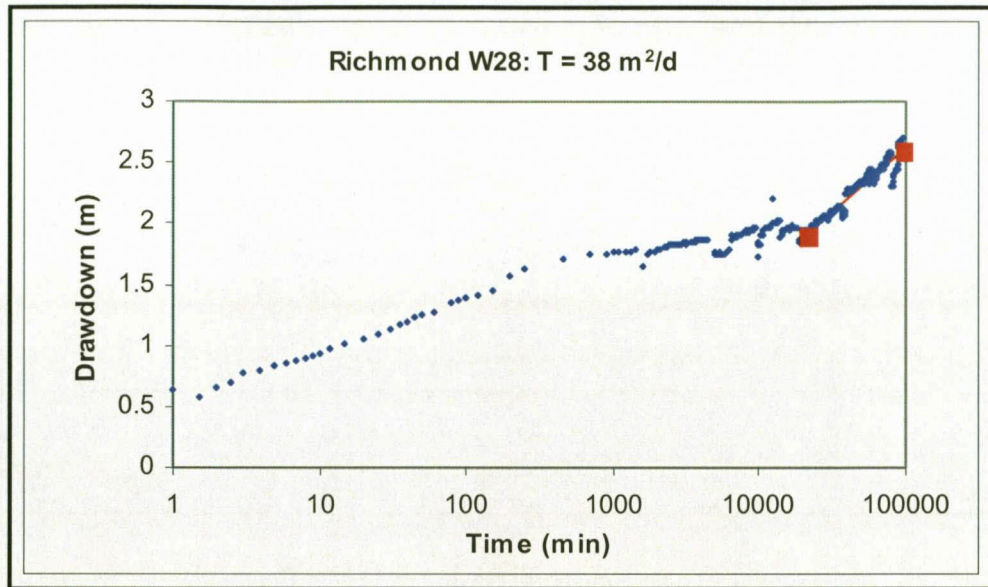


Figure 129: Drawdown for Richmond borehole RMGW28 (van Tonder and Dennis, 2005)

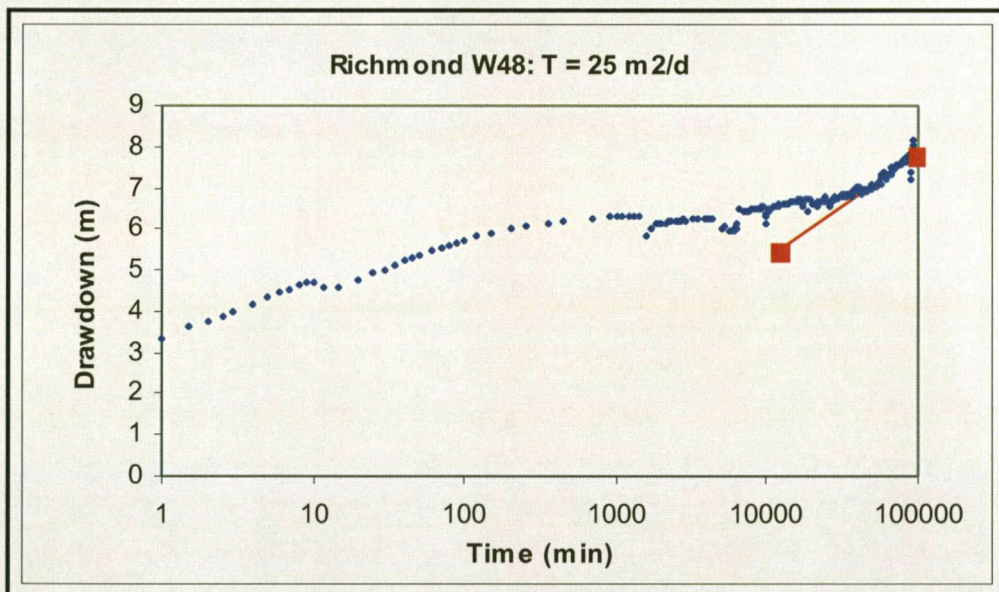


Figure 130: Drawdown for Richmond borehole RMGW48 (van Tonder and Dennis, 2005)

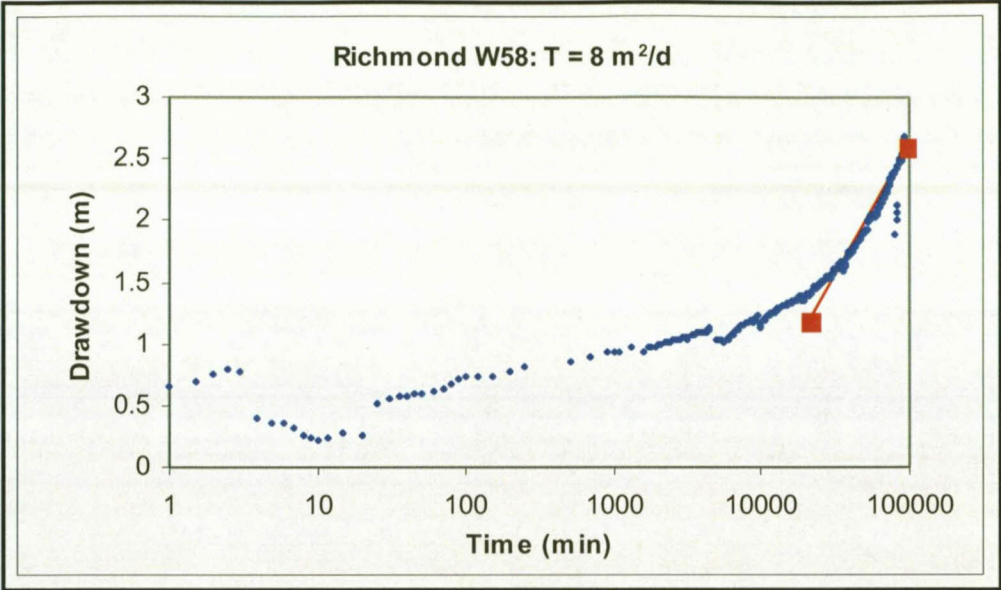


Figure 131: Drawdown for Richmond borehole RMGW58 (van Tonder and Dennis, 2005)

8.3 Annexure 3 Richmond daily rainfall data for 10 years

Table 15: Daily rainfall data for Richmond area from 1998 to 2007

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
01-Jan	11.8	0.4	22.0	24.0	1.2	0.0	0.8	1.2	37.6	0.8
02-Jan	0.4	2.0	11.0	3.0	10.8	0.4	3.6	4.4	0.4	0
03-Jan	0.0	0.4	2.0	0.0	0.2	5.4	0.2	21.8	0.0	0
04-Jan	0.2	0.6	3.6	0.0	0.2	7.4	0.0	0.8	0.0	16.4
05-Jan	16.2	0.6	0.2	0.0	0	6.4	9.4	0.0	0.0	0
06-Jan	11.0	13.6	0.2	0.0	4.8	4.0	14.0	0.0	2.0	4.2
07-Jan	7.2	3.8	44.0	0.0	1	0.6	28.8	11.0	88.0	11.2
08-Jan	25.4	0.4	1.8	20.0	0.4	30.0	10.2	6.6	29.0	0.2
09-Jan	6.8	0.0	0.2	6.0	0	0.2	0.2	8.0	2.2	1.2
10-Jan	0.0	0.6	12.8	0.0	0	11.6	13.6	4.0	0.2	0
11-Jan	0.2	12.6	19.2	0.0	1	15.0	0.0	0.0	1.0	9
12-Jan	0.2	3.2	19.8	0.0	0	1.0	5.8	4.6	0.0	1
13-Jan	9.6	0.0	4.2	25.2	0	0.0	4.0	1.8	0.0	1
14-Jan	11.2	0.0	16.4	2.8	0	1.0	10.4	31.8	0.2	3.8
15-Jan	20.2	0.0	13.0	0.0	0.2	0.2	0.2	5.2	0.0	1.6
16-Jan	0.0	11.0	23.0	0.0	1.4	0.0	0.4	0.2	0.0	3.6
17-Jan	2.8	5.6	5.8	0.0	0	0.0	0.0	8.4	16.6	17.8
18-Jan	0.2	12.4	0.2	11.2	0	5.0	0.0	0.0	22.6	0.2
19-Jan	3.8	0.2	0.0	4.6	0	4.8	0.0	1.0	6.2	0
20-Jan	0.2	21.8	0.0	6.8	0	0.2	26.0	73.8	0.4	0
21-Jan	0.0	4.8	0.0	0.4	2.8	4.8	1.6	26.4	13.0	5.8
22-Jan	0.0	11.2	0.0	0.0	0.0	0.0	12.0	11.4	2.6	0.4
23-Jan	2.0	1.2	0.0	1.0	47.6	0.0	29.0	8.6	24.6	0.2
24-Jan	4.6	0.0	3.6	6.8	13.8	0.0	0.0	0.2	0.6	0.6
25-Jan	1.4	0.0	5.8	2.4	2.0	14.4	0.2	3.0	10.0	5.4

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
28-Jan	5.4	0.0	6.4	38.8	0.0	0.0	19.0	8.0	11.0	0.2
29-Jan	6.8	41.2	5.8	14.4	3.2	0.0	12.0	0.2	1.2	7.2
30-Jan	15.0	6.4	0.0	2.2	43.4	7.4	12.6	12.6	17.2	34.6
31-Jan	0.2	0.2	0.0	0.4	3.6	0.0	18.2	1.6	13.0	0.2
01-Feb	0.0	7.0	17.6	0.4	0.2	0.0	0.0	0.0	2.4	11.4
02-Feb	0.4	27.8	23.2	0.0	0.0	0.0	0.0	0.2	0.0	0
03-Feb	0.4	9.0	2.8	6.2	0.2	0.2	0.0	0.0	24.8	0
04-Feb	0.2	8.0	0.4	2.0	0.0	1.2	0.0	3.6	0.4	1
05-Feb	0.0	7.0	6.4	0.0	0.0	1.0	1.0	0.2	0.2	4
06-Feb	3.6	0.6	0.8	0.0	16.6	23.4	0.6	0.0	1.4	0.2
07-Feb	0.4	0.4	0.0	15.0	10.6	0.0	0.4	3.6	0.6	1.2
08-Feb	2.6	0.0	0.0	8.2	0.0	0.2	0.0	0.0	25.0	0.2
09-Feb	7.4	0.0	0.0	3.2	0.0	0.0	0.0	3.4	32.4	17.4
10-Feb	3.6	0.0	0.2	0.2	0.2	28.0	0.0	7.0	2.4	17.2
11-Feb	19.0	0.0	0.0	0.0	1.0	4.4	0.0	0.0	0.0	0.2
12-Feb	26.2	3.2	18.2	0.4	12.4	0.6	26.2	4.4	2.2	0.2
13-Feb	1.0	5.8	4.8	40.8	0.2	0.0	20.2	2.6	0.4	0
14-Feb	0.8	6.0	7.0	12.6	0.0	10.4	15.6	5.4	0.2	0
15-Feb	10.4	0.2	3.2	3.2	0.6	0.0	0.2	0.0	0.0	0.2
16-Feb	51.0	0.0	0.2	0.2	1.2	0.0	0.0	0.2	3.0	0.4
17-Feb	30.6	4.6	0.0	3.0	0.4	7.0	29.4	0.2	1.4	0.2
18-Feb	0.4	59.2	0.0	5.6	0.0	0.4	0.2	75.0	11.0	1
19-Feb	1.6	41.4	0.0	16.2	0.0	0.0	20.0	0.0	0.0	0.2
20-Feb	0.0	0.2	44.8	1.2	10.8	0.0	22.8	10.2	2.8	0
21-Feb	8.0	10.4	3.2	0.0	0.4	3.8	0.6	0.2	6.6	0
22-Feb	24.4	0.2	58.2	1.6	0.0	12.0	1.0	0.0	0.2	0

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
23-Feb	8.6	13.4	4.4	21.8	0.0	2.2	0.2	0.4	5.6	1.4
24-Feb	0.2	0.0	0.2	2.4	1.0	2.6	0.2	0.0	0.0	0.4
25-Feb	0.0	1.8	0.0	0.0	0.2	0.0	0.0	4.8	0.0	0
26-Feb	4.6	0.0	22.4	0.4	1.2	0.4	7.0	0.2	0.0	0
27-Feb	2.6	0.0	0.8	0.6	6.6	0.8	37.4	3.0	0.2	6.4
28-Feb	6.8	0.0	0.0	0.0	10.4	2.6	4.0	0.0	8.0	3.2
01-Mar	17.0	0.4	9.2	0.4	2.8	0.2	11.0	27.4	0.2	13.2
02-Mar	2.0	0.8	20.4	0.0	12.4	1.8	0.0	0.4	2.8	0.8
03-Mar	0.0	12.0	6.4	0.0	0.6	0.2	10.6	0.2	4.4	7.4
04-Mar	0.0	0.0	52.2	0.0	0.0	1.8	0.0	0.0	3.0	4.8
05-Mar	2.8	0.0	0.0	0.0	14.4	19.8	0.0	0.0	0.2	4.2
06-Mar	5.8	0.0	0.2	0.4	1.6	18.2	14.3	0.0	0.4	0.2
07-Mar	3.6	0.0	5.4	2.4	8.0	0.2	0.6	7.4	0.0	0.6
08-Mar	29.8	5.2	3.8	2.2	1.4	0.2	0.0	3.6	0.0	0
09-Mar	0.4	4.0	0.4	1.0	15.2	0.0	0.0	0.2	0.0	0.2
10-Mar	0.0	4.0	2.0	47.2	0.0	0.0	8.6	3.8	0.0	0
11-Mar	0.2	3.8	2.2	3.4	6.8	0.0	0.0	0.2	0.0	16
12-Mar	0.0	3.6	6.8	0.6	0.8	0.0	4.2	0.0	8.4	0.4
13-Mar	0.0	0.4	2.6	1.2	4.0	15.6	4.0	0.0	0.4	0.4
14-Mar	0.4	0.0	0.8	0.2	0.0	0.0	0.0	3.6	0.2	0
15-Mar	9.4	0.0	0.0	0.0	0.0	11.4	0.0	5.8	0.2	1.2
16-Mar	0.0	0.0	1.2	2.8	0.0	0.0	0.0	3.4	8.2	0.2
17-Mar	0.0	19.0	0.8	29.6	0.0	0.4	0.0	0.0	0.2	0.2
18-Mar	3.8	0.4	11.2	24.6	19.4	0.0	0.0	0.2	9.8	1
19-Mar	0.2	0.0	8.8	0.2	3.2	0.0	0.0	0.0	0.6	0.2
20-Mar	0.0	10.4	0.8	3.0	0.0	4.4	0.2	0.0	7.6	0.2

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
22-Mar	2.6	14.4	1.2	7.8	0.0	14.8	0.0	0.0	0.2	0.2
23-Mar	4.2	2.0	0.4	17.6	0.2	9.0	0.0	0.0	0.0	0
24-Mar	2.8	3.0	0.2	1.2	0.0	0.4	23.8	0.0	0.0	0
25-Mar	25.4	0.2	0.0	3.4	8.4	0.0	0.0	0.0	1.6	0
26-Mar	5.8	17.4	25.4	0.2	1.2	0.2	0.0	0.0	0.0	3
27-Mar	6.0	0.0	10.2	0.0	1.2	0.0	0.0	0.0	12.8	2.8
28-Mar	3.0	0.4	21.6	0.0	0.2	0.0	0.0	0.0	0.0	0.8
29-Mar	0.0	0.2	4.2	0.0	0.0	0.0	0.0	0.0	0.0	0.8
30-Mar	0.0	0.0	3.2	0.0	0.0	0.0	0.0	0.0	0.4	0.2
31-Mar	2.0	0.2	1.0	4.4	0.0	0.0	0.0	0.0	0.0	0.4
01-Apr	1.4	0.0	2.6	3.0	0.2	0.0	2.4	0.0	0.0	0
02-Apr	0.2	0.0	0.2	0.6	0.0	0.0	0.0	0.0	0.0	0
03-Apr	0.2	0.0	0.0	7.2	0.0	2.4	2.2	0.0	0.0	0
04-Apr	0.0	0.0	19.4	0.8	0.0	0.0	0.2	0.0	0.0	0
05-Apr	0.0	0.0	20.2	0.0	1.2	16.4	4.2	0.0	0.0	27.6
06-Apr	0.0	0.0	7.2	0.0	2.0	0.2	0.0	0.0	0.0	20.8
07-Apr	0.0	0.0	0.0	3.8	0.0	0.4	9.2	0.0	6.8	3.2
08-Apr	0.6	1.4	0.0	8.6	0.0	0.0	0.0	0.0	5.8	5.2
09-Apr	0.0	1.0	0.2	0.6	5.0	0.0	0.2	0.0	1.2	0
10-Apr	2.6	3.4	0.0	32.0	33.4	15.0	8.0	0.0	7.0	0
11-Apr	0.0	0.2	0.4	0.4	1.2	4.8	0.2	0.0	8.8	1.6
12-Apr	0.0	0.0	0.0	0.2	0.0	1.0	0.0	0.0	0.0	1.6
13-Apr	0.6	0.2	25.6	0.0	0.0	0.0	0.0	0.0	0.4	0
14-Apr	3.0	0.0	1.2	2.8	0.0	4.0	0.0	0.0	0.4	0
15-Apr	3.0	3.4	0.0	0.0	0.0	1.4	0.0	0.0	1.2	0
16-Apr	0.0	3.4	0.0	11.4	0.0	0.2	2.2	0.0	0.0	1.8

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
19-Apr	0.0	0.0	0.0	0.2	0.0	0.0	0.0	6.8	1.0	5.6
20-Apr	0.0	0.0	2.4	0.0	0.0	0.0	13.4	0.0	0.2	0.2
21-Apr	0.0	0.0	3.8	0.0	0.0	0.0	14.0	0.0	0.0	1.6
22-Apr	21.8	0.0	0.0	0.0	0.0	0.0	0.2	0.2	0.2	0.2
23-Apr	20.6	0.0	0.0	12.6	0.0	0.0	0.0	3.4	7.0	0
24-Apr	0.0	0.0	1.2	0.0	0.0	13.2	0.0	1.0	2.6	0
25-Apr	0.0	0.0	0.0	0.0	0.0	3.8	0.0	0.6	0.0	0
26-Apr	0.0	0.0	0.0	0.0	0.2	0.0	0.4	0.0	0.2	0
27-Apr	0.0	0.0	6.8	2.4	0.0	0.0	0.8	0.2	0.2	0.2
28-Apr	0.0	0.8	0.4	0.2	0.0	0.0	0.0	0.2	0.2	0
29-Apr	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.4	0
30-Apr	0.0	0.0	1.4	4.2	9.4	0.0	0.0	0.0	17.4	0
01-May	0.0	0.0	0.2	0.0	0.2	0.0	0.0	0.2	2.4	0
02-May	0.0	0.0	1.8	0.2	0.0	0.2	0.4	0.0	0.0	0
03-May	0.0	0.0	7.8	1.0	0.0	0.0	3.6	0.0	3.2	0
04-May	0.0	0.0	29.4	0.6	0.0	0.0	0.6	6.4	0.0	0
05-May	0.0	0.2	4.0	0.0	2.4	0.0	0.2	0.2	0.2	0
06-May	0.0	0.0	0.2	0.0	1.6	0.2	0.0	0.0	0.0	0
07-May	7.0	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0
08-May	0.0	3.8	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0
09-May	1.2	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0
10-May	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.4	0
11-May	0.0	0.0	0.0	0.2	0.0	0.0	0.0	0.0	0.2	0
12-May	0.0	0.0	3.4	0.0	0.0	9.0	0.0	0.0	0.0	0
13-May	0.0	0.0	0.6	0.0	0.0	3.6	0.0	0.0	0.0	0
14-May	0.0	0.0	0.0	0.0	0.0	6.0	0.0	0.0	0.0	0

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
15-May	0.4	0.6	0.4	0.0	0.0	2.0	0.0	1.4	0.0	0
16-May	0.6	2.2	0.0	0.0	0.0	0.8	0.0	0.0	0.0	1.2
17-May	0.0	0.4	0.0	0.0	0.0	0.0	0.0	0.0	1.8	1.2
18-May	0.0	1.8	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.2
19-May	0.0	0.2	26.6	0.0	0.0	1.2	0.0	0.0	0.0	0
20-May	0.0	0.0	2.2	0.0	0.0	1.4	0.2	0.0	1.8	2.8
21-May	0.0	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.2	0
22-May	0.0	0.2	0.0	0.0	2.0	0.0	0.0	0.0	0.0	0.4
23-May	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.8	0.2	0
24-May	0.0	0.0	0.0	0.0	0.0	0.0	0.0	2.4	4.4	0.4
25-May	0.0	0.0	10.0	0.2	0.0	0.0	0.0	0.2	0.2	0
26-May	0.0	0.0	0.4	0.0	0.0	0.4	0.0	0.4	0.0	0
27-May	0.0	0.0	7.4	0.0	0.0	0.0	0.0	0.0	0.0	0
28-May	15.4	0.0	0.4	0.0	0.0	0.2	0.0	4.4	0.0	0.2
29-May	0.2	0.0	0.2	0.0	0.0	0.0	0.0	1.4	0.0	0
30-May	0.0	0.0	0.0	0.2	31.2	0.0	0.0	0.0	0.0	0
31-May	0.0	0.2	0.0	0.2	0.0	0.0	0.0	0.2	0.0	0
01-Jun	0.0	0.0	0.0	0.2	0.2	0.0	0.0	0.0	0.2	0
02-Jun	0.6	0.2	0.0	0.0	0.2	0.0	0.0	0.0	0.4	0
03-Jun	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.0	0.2	0.2
04-Jun	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0
05-Jun	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	4.4
06-Jun	0.0	1.0	0.4	0.0	0.0	3.6	0.0	0.2	0.0	1.4
07-Jun	0.0	0.0	0.0	0.0	0.2	1.6	0.0	1.2	0.0	0.2
08-Jun	0.0	0.0	0.0	0.0	0.0	0.2	0.0	0.0	0.0	0
09-Jun	0.0	0.0	0.0	0.0	0.0	0.2	0.0	0.0	0.0	0

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
10-Jun	0.0	0.0	0.0	0.0	0.0	0.2	0.0	0.0	0.0	0.2
11-Jun	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0
12-Jun	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.4	0.0	0
13-Jun	0.0	0.0	0.0	0.0	2.2	0.2	0.0	0.0	0.0	0
14-Jun	0.0	0.0	0.0	0.0	9.0	0.0	0.0	0.0	0.0	0.4
15-Jun	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0
16-Jun	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0
17-Jun	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0
18-Jun	0.0	0.0	3.2	0.0	0.0	0.4	0.0	0.0	0.0	0
19-Jun	0.0	0.0	0.4	0.0	0.0	0.0	0.0	0.0	0.0	0
20-Jun	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.2
21-Jun	0.0	0.2	0.0	0.0	0.2	0.0	0.0	0.0	0.0	0
22-Jun	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.0	3.0	0
23-Jun	0.0	0.0	8.2	0.0	1.4	0.0	0.0	0.4	0.2	0
24-Jun	0.0	0.0	0.4	0.0	0.0	0.0	0.0	0.0	0.0	0
25-Jun	0.0	0.0	0.4	0.0	0.0	0.0	0.0	0.0	0.0	0
26-Jun	0.0	0.2	0.0	0.6	0.0	0.0	0.0	0.0	0.0	12.4
27-Jun	0.0	0.0	0.0	0.2	0.0	0.2	0.0	0.0	0.2	3.2
28-Jun	0.0	0.0	0.0	0.4	0.0	0.2	9.8	0.0	0.0	0
29-Jun	0.0	0.0	0.0	1.4	0.0	0.0	0.0	0.0	0.0	0.4
30-Jun	0.0	0.0	0.0	0.2	0.0	0.0	0.2	1.6	0.2	0.2
01-Jul	0.0	0.0	0.2	0.0	0.2	0.0	0.0	0.0	0.2	0
02-Jul	0.0	0.0	0.0	0.0	0.0	0.0	4.0	0.2	0.0	0.4
03-Jul	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.2
04-Jul	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.4	0.2	0
05-Jul	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.2

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
06-Jul	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0
07-Jul	0.0	0.0	0.0	0.0	0.0	0.2	0.2	0.0	0.0	0
08-Jul	0.0	0.0	0.0	0.0	0.0	0.2	0.0	0.0	0.0	0
09-Jul	0.0	0.0	0.0	0.0	0.8	0.0	0.0	0.0	0.0	0
10-Jul	0.0	0.0	0.0	0.4	0.0	0.0	0.0	0.0	0.0	0
11-Jul	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0
12-Jul	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0
13-Jul	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0
14-Jul	0.0	0.0	0.0	0.0	0.0	0.0	3.6	0.6	0.2	0
15-Jul	0.0	0.0	0.4	0.0	1.2	0.0	0.0	0.0	0.0	0
16-Jul	0.0	0.0	0.4	0.0	0.0	0.0	0.2	0.0	0.0	0
17-Jul	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.8	0
18-Jul	0.0	0.0	0.0	0.0	13.4	0.0	0.0	0.0	0.0	0.2
19-Jul	0.0	0.0	0.0	0.0	7.4	0.0	0.0	0.0	0.0	0
20-Jul	0.0	0.0	0.0	0.0	6.8	0.0	0.0	1.0	0.0	0
21-Jul	10.2	0.0	0.4	0.0	9.8	0.0	0.0	0.0	0.0	0
22-Jul	0.0	0.0	0.0	0.0	0.4	0.0	0.0	0.0	1.4	0
23-Jul	0.0	0.0	0.0	4.4	0.0	0.0	0.0	0.0	0.0	0
24-Jul	0.4	1.0	0.0	1.0	0.0	0.0	0.0	0.0	0.0	0
25-Jul	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0
26-Jul	0.0	0.0	2.2	0.2	0.0	0.0	2.6	0.0	0.0	0
27-Jul	0.0	3.6	0.0	0.0	0.0	0.0	0.6	0.2	0.0	3.8
28-Jul	0.0	0.2	0.0	0.0	0.0	0.0	24.6	0.0	0.0	0
29-Jul	1.4	0.0	0.0	0.0	0.0	0.0	15.0	0.2	0.0	0
30-Jul	0.2	0.0	0.0	0.2	0.0	0.0	0.4	0.0	0.0	0
31-Jul	0.0	0.2	0.0	0.0	0.0	0.0	0.4	0.2	0.0	0

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
01-Aug	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.2	6.2	0
02-Aug	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	52.8	0
03-Aug	0.0	0.0	0.0	0.0	0.6	0.0	0.0	0.0	2.4	0
04-Aug	0.0	0.0	0.0	0.0	30.4	0.0	0.0	0.0	0.0	0
05-Aug	0.4	0.0	0.0	0.0	11.8	0.0	0.0	0.0	0.0	0.2
06-Aug	0.2	0.0	0.0	0.2	0.4	0.0	0.0	0.0	0.0	0
07-Aug	0.0	0.0	0.0	0.0	0.0	2.8	0.0	0.0	0.0	8
08-Aug	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.2
09-Aug	0.2	0.2	0.0	0.0	0.0	0.0	0.0	29.2	1.6	0.2
10-Aug	0.0	0.0	0.0	0.0	0.0	0.0	0.0	7.4	0.0	0
11-Aug	0.0	0.0	0.0	0.0	0.0	1.2	0.0	0.2	0.0	0
12-Aug	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.6	0.0	0
13-Aug	0.0	0.0	0.0	0.0	0.0	9.8	0	0.0	0.0	0
14-Aug	0.0	0.0	0.0	0.0	15.2	0.4	0	0.0	0.0	0
15-Aug	0.0	0.0	0.0	0.0	7.0	7.4	0	0.0	29.4	0
16-Aug	0.0	0.0	0.0	0.0	0.2	0.0	0	0.0	14.4	1.8
17-Aug	0.0	0.0	0.0	0.0	11.4	0.0	0.4	1.2	0.2	0
18-Aug	0.2	0.0	0.0	0.0	7.8	0.0	0.2	0.0	0.0	0
19-Aug	0.0	0.0	0.0	0.0	0.0	0.0	0	0.0	0.0	0
20-Aug	0.0	0.0	0.0	0.0	0.4	0.0	0	0.0	0.0	0
21-Aug	0.2	0.0	0.0	0.0	0.4	0.2	2.8	0.2	0.2	0
22-Aug	19.2	0.0	0.0	0.0	0.0	0.0	0.8	0.2	0.0	0
23-Aug	7.4	0.0	0.0	0.0	0.0	0.0	0.2	0.2	0.0	0
24-Aug	0.8	0.0	0.0	0.0	0.0	0.0	0.4	0.0	4.0	1
25-Aug	0.2	0.0	0.0	0.0	20.0	0.0	0	0.0	3.6	0
26-Aug	0.0	0.2	0.0	0.2	0.0	0.0	7.4	0.0	0.0	0

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
27-Aug	0.2	0.0	0.0	0.0	5.0	0.0	0.2	0.0	0.2	0.2
28-Aug	0.0	0.0	0.0	0.0	1.2	1.6	0	0.0	0.0	0
29-Aug	0.0	0.0	0.0	3.0	0.4	0.0	0.4	0.0	0.0	0.2
30-Aug	0.0	0.0	0.0	2.4	0.2	0.0	0	0.0	0.0	0
31-Aug	0.0	0.0	0.0	4.2	0.0	0.0	0	0.0	0.0	0
01-Sep	0.0	3.8	0.0	0.0	0.0	0.0	0	3.0	0.0	0
02-Sep	0.0	0.0	0.0	0.0	0.0	0.0	0	0.0	0.0	0
03-Sep	0.0	0.6	0.0	0.0	0.4	0.0	0.6	0.0	0.0	0
04-Sep	0.0	2.0	0.0	0.0	3.2	0.0	0	0.6	0.0	0
05-Sep	0.0	0.0	0.0	0.0	4.4	0.0	0.2	0.2	0.0	0
06-Sep	1.8	0.0	0.0	0.0	3.2	4.8	3.2	3.2	0.0	0
07-Sep	0.2	0.0	0.0	0.0	0.2	9.0	16	1.2	0.0	0
08-Sep	0.0	0.0	0.0	0.0	0.2	0.6	0.4	0.0	0.0	2.6
09-Sep	0.0	0.0	0.0	0.0	7.0	0.0	0.2	0.0	0.0	0.2
10-Sep	0.0	0.4	0.0	0.0	20.6	0.0	0	0.0	0.0	0.2
11-Sep	0.0	0.0	0.0	0.6	0.2	0.0	0	0.2	0.0	0
12-Sep	0.0	0.0	0.0	0.6	0.0	0.0	0.2	0.0	0.2	0
13-Sep	0.2	0.0	0.0	23.8	0.0	0.0	0	0.0	0.2	0.4
14-Sep	0.0	0.0	0.0	1.2	0.0	0.8	0.2	0.0	0.0	3.2
15-Sep	0.0	0.0	1.2	0.0	0.0	14.2	0	0.0	0.0	0
16-Sep	0.0	0.0	7.6	1.4	0.0	1.0	0	0.2	0.0	0
17-Sep	11.4	0.0	0.2	1.0	0.0	0.4	0.2	0.0	5.8	0
18-Sep	2.2	0.0	16.6	7.0	0.0	18.8	0	0.0	1.8	0
19-Sep	0.0	0.0	9.6	5.2	0.0	6.0	0.2	0.0	0.0	0
20-Sep	0.0	0.6	22.6	0.6	0.0	0.0	0	0.2	0.0	1
21-Sep	0.0	1.2	1.0	2.4	0.4	0.0	0	2.0	0.0	0

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
22-Sep	0.0	0.2	0.0	5.4	1.8	2.0	0	0.4	0.0	0
23-Sep	0.0	0.6	0.0	1.4	0.0	5.6	0.2	0.2	0.0	0
24-Sep	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.2
25-Sep	2.2	0.0	0.0	0.0	0.2	0.2	1.8	0.0	7.2	0
26-Sep	1.2	0.0	0.0	0.0	1.2	0.0	72.0	0.2	58.2	1.6
27-Sep	0.2	0.6	0.0	0.4	0.0	0.0	5.2	0.6	0.0	0.2
28-Sep	6.6	3.8	0.0	0.0	1.6	0.0	0.2	1.0	0.2	0.6
29-Sep	0.4	0.0	0.0	0.0	1.6	0.0	0.0	0.0	0.0	1.2
30-Sep	0.2	0.0	0.0	0.0	0.0	0.2	0.0	0.0	0.0	1
01-Oct	5.2	0.0	1.2	0.0	0.0	0.2	0.2	0.0	0.0	0
02-Oct	5.2	1.6	0.0	0.0	0.0	0.0	0.0	0.0	0.8	0.6
03-Oct	0.0	3.0	0.0	0.2	0.0	0.0	0.0	0.0	25.2	0.2
04-Oct	0.0	0.6	0.0	0.0	0.0	0.0	0.0	1.4	10.8	0
05-Oct	0.0	0.2	1.4	5.2	0.0	0.0	0.0	9.2	1.0	4.4
06-Oct	0.0	0.0	1.6	2.4	0.0	0.0	0.0	4.2	0.8	4.4
07-Oct	1.4	3.6	0.2	0.8	0.0	0.0	0.0	5.4	0.4	1.8
08-Oct	0.0	0.0	0.2	2.2	0.0	0.4	0.0	0.0	4.0	16
09-Oct	0.2	0.0	0.4	0.2	0.0	1.4	0.0	0.0	19.6	12.6
10-Oct	9.2	0.0	0.2	0.0	0.0	0.6	0.0	0.0	0.8	4.8
11-Oct	6.2	2.8	4.6	0.0	0.0	0.2	0.0	0.0	0.0	0.2
12-Oct	0.2	0.0	1.0	0.0	0.0	0.0	0.0	0.0	0.0	0.2
13-Oct	0.0	0.0	0.0	1.4	0.0	0.0	0.0	0.0	8.0	8.4
14-Oct	0.0	0.0	0.0	0.2	0.4	0.8	0.0	0.0	6.8	0.2
15-Oct	12.8	0.0	0.0	0.4	0.0	2.2	2.2	8.6	20.8	3
16-Oct	11.6	1.0	0.0	6.2	0.0	2.4	0.0	33.2	13.2	0.4
17-Oct	0.2	2.2	0.0	0.0	0.0	0.0	0.0	0.0	2.6	2.8

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
18-Oct	0.0	4.4	0.2	3.0	0.0	0.2	13.4	8.4	0.0	5.4
19-Oct	0.0	14.6	0.0	1.8	0.8	0.8	2.0	6.6	0.0	0.4
20-Oct	0.0	0.6	3.2	3.0	0.2	0.0	0.6	0.0	0.0	0
21-Oct	0.6	0.2	1.4	14.2	0.0	0.0	0.0	0.2	0.0	12
22-Oct	0.0	0.0	0.6	0.4	0.0	0.2	3.8	0.0	0.0	2.4
23-Oct	0.2	13.2	6.8	3.4	0.2	0.0	13.8	0.0	0.0	6.4
24-Oct	1.2	15.4	2.6	23.6	0.6	0.0	0.6	0.0	0.0	0
25-Oct	0.8	12.8	0.4	9.6	0.0	0.0	0.0	0.0	0.0	0
26-Oct	0.2	3.6	1.4	0.0	0.0	0.2	14.6	0.0	0.0	4
27-Oct	5.6	1.6	0.4	0.0	11.6	0.0	2.2	0.0	0.0	0.8
28-Oct	6.8	0.0	0.0	0.4	0.2	0.0	2.0	16.2	0.0	1.2
29-Oct	0.0	0.0	0.2	1.4	1.4	0.0	0.0	0.2	0.0	11.2
30-Oct	0.0	0.6	4.6	1.2	0.0	11.6	0.6	0.2	0.0	0.2
31-Oct	0.8	0.0	0.2	6.8	0.0	1.4	0.0	0.4	0.0	21.2
01-Nov	0.0	0.0	0.0	0.2	4.0	1.8	0.0	0.0	25.6	18
02-Nov	0.2	0.0	1.2	0.0	0.0	2.2	0.0	0.0	3.6	0.2
03-Nov	0.0	0.0	0.0	7.8	0.6	1.0	0.6	0.0	0.0	0.2
04-Nov	0.0	0.0	5.0	2.0	0.0	9.0	0.6	6.4	0.0	0
05-Nov	0.2	0.0	0.0	0.0	14.4	0.4	0.0	11.6	3.0	16
06-Nov	0.0	0.0	0.0	0.0	0.2	6.8	38.2	18.8	1.6	3.6
07-Nov	0.0	0.0	3.0	0.0	0.0	12.8	0.0	3.8	0.0	0
08-Nov	15.6	1.4	8.8	13.0	0.0	1.0	3.6	0.8	0.0	0
09-Nov	0.4	0.6	67.8	13.8	1.2	0.0	0.0	0.0	1.2	0.2
10-Nov	11.6	0.2	0.2	14.8	0.6	0.0	0.0	1.0	0.0	0.2
11-Nov	0.0	0.0	0.2	1.2	0.0	0.0	0.0	6.6	2.2	0
12-Nov	0.2	4.0	0.0	10.6	0.0	0.0	0.0	1.8	7.6	0

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
13-Nov	0.0	5.6	4.6	0.2	29.6	4.6	25.6	20.6	0.0	0
14-Nov	0.0	1.4	8.6	8.8	10.4	0.0	6.4	0.4	0.0	0.2
15-Nov	0.0	9.2	0.8	14.8	0.2	0.0	0.2	0.0	0.0	8
16-Nov	0.4	16.0	0.0	14.4	0.0	9.2	0.2	4.8	23.0	10.4
17-Nov	15.6	0.2	0.0	1.6	0.4	1.0	0.0	0.0	27.4	0.2
18-Nov	5.6	0.2	0.0	12.6	0.0	0.0	12.0	5.4	5.4	0
19-Nov	16.8	0.0	10.6	12.2	0.2	0.0	14.4	0.0	0.0	1.4
20-Nov	0.2	4.2	2.8	0.0	0.2	3.4	14.6	0.4	12.2	0
21-Nov	0.2	0.2	6.4	0.0	0.0	0.8	16.8	0.6	6.0	4.0
22-Nov	0.2	0.4	0.0	0.0	0.0	2.2	6.4	1.0	1.0	10.0
23-Nov	0.2	0.0	0.0	1.6	4.8	0.2	7.2	0.2	0.0	0.0
24-Nov	12.2	8.6	3.2	1.8	1.0	19.8	0.2	0.0	3.0	0.0
25-Nov	36.4	4.8	6.4	15.0	0.0	1.2	0.0	17.2	3.2	8.0
26-Nov	21.0	24.6	0.2	8.4	0.0	10.0	0.0	8.4	0.6	0.0
27-Nov	28.2	0.6	2.8	0.2	0.0	0.0	0.0	0.4	7.2	0.0
28-Nov	0.0	10.2	0.2	0.2	0.0	0.2	1.8	0.4	1.0	2.8
29-Nov	2.0	3.6	16.0	1.6	0.6	0.6	0.4	0.8	0.0	6.2
30-Nov	27.2	7.6	3.8	41.6	0.0	5.6	1.2	21.8	1.6	11
01-Dec	23.6	4.0	1.0	0.8	0.0	1.6	2.0	0.0	3.6	1
02-Dec	20.6	7.2	0.0	19.4	6.2	0.2	0.6	0.0	0.4	0.8
03-Dec	1.6	6.0	9.0	0.2	30.0	0.0	0.0	0.0	0.0	4.2
04-Dec	0.2	0.8	10.8	7.2	0.2	0.0	0.0	0.2	1.8	0.8
05-Dec	0.2	5.6	0.2	1.8	0.0	0.4	0.0	9.4	0.0	15.2
06-Dec	0.0	1.2	4.0	0.2	0.0	0.0	18.6	0.8	0.0	12.8
07-Dec	7.8	0.2	0.0	0	0.2	2.6	38.2	0.6	0.0	0.8
08-Dec	4.0	17.4	12.4	0	51.6	3.2	6.2	1.2	3.8	14

Date	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
09-Dec	0.4	6.8	9.2	0	0.4	1.2	0.0	0.0	13.6	3.6
10-Dec	44.4	5.2	0.4	21.6	0.2	0.4	0.0	2.4	17.8	1
11-Dec	12.2	10.0	22.2	3.8	1.4	3.6	1.0	0.2	0.0	7.6
12-Dec	9.4	0.2	21.2	0.6	3.8	0.0	0.0	0.0	0.0	1
13-Dec	0.4	0.2	0.0	29.8	14.4	0.2	0.2	0.0	0.0	0
14-Dec	5.4	0.0	0.0	0.2	0.4	0.3	0.0	0.0	0.4	1.4
15-Dec	7.0	8.2	0.4	0	1.6	0.0	5.8	0.0	0.0	0.2
16-Dec	0.0	7.0	0.0	12.2	2.8	0.0	6.6	0.6	4.0	3.2
17-Dec	0.0	1.8	0.2	0	0.2	0.0	0.2	3.4	7.2	9
18-Dec	0.0	0.0	0.8	0	2.4	0.0	6.6	2.2	1.4	0
19-Dec	3.2	2.6	0.0	9	2.0	0.0	12.2	0.6	0.4	0
20-Dec	6.4	62.0	0.0	0	2.6	0.2	0.8	1.6	3.2	0.4
21-Dec	0.2	5.0	29.4	0	0.4	0.0	6.2	17.4	6.4	5.2
22-Dec	1.6	32.6	0.0	3.8	36.8	0.9	22.8	0.6	0.2	0.2
23-Dec	1.0	0.0	0.4	12.8	1.0	14.7	12.8	0.0	61.6	10.2
24-Dec	4.2	19.8	0.4	0.4	0.0	1.5	2.8	3.2	12.6	9.2
25-Dec	20.2	7.4	31.2	19.6	0.0	23.3	4.0	1.4	10.4	7.2
26-Dec	0.4	7.6	5.4	2	0.6	0.0	0.8	0.2	22.2	1.8
27-Dec	0.8	3.2	0.2	0	0.2	0.0	2.0	0.2	1.6	0.2
28-Dec	3.2	32.8	0.2	0	12.4	0.0	3.0	0.8	0.6	0.2
29-Dec	2.6	0.4	4.4	21.2	1.6	0.3	48.8	3.0	3.4	0
30-Dec	14.6	0.0	0.0	0.8		0.0	0.2	0.0	6.0	0.2
31-Dec	52.6	0.0	0.0	35.8		0.0	1.6	2.6	3.2	0.2

8.4 Annexure 4: Seekoei area – Monthly rainfall data (Oct 2006 – Sept 2006) (Source data – van Tonder, 2012)

Area	Oct-05	Nov-05	Dec-05	Jan-06	Feb-06	Mar-06	Apr-06	May-06	Jun-06	Jul-06	Aug-06	Sep-06	Total
Richmond	18	13	22	56	165.5	51	52	26.5	7	14.3	80	1.5	506.8
Hanover	22	18	14	58	222	48	52.5	35	4	16.5	75	2	567
Colesberg	12.5	25	0	123.5	125	15.5	77	41	0	21.5	89	0.2	530.2
													534.6667

8.5 Annexure 5: Seekoei chemistry data (Source data – van Tonder, 2012)

SiteName	DateTimeMeas	pH	EC mS/m	TDS mg/l	Ca mg/l	Mg mg/l	Na mg/l	K mg/l	MALK	Cl mg/l	SO4 mg/l	NO3-N mg/l	Al mg/l	N_Amonia mg/l
EWR-1	2006/03/15 12:00	8.35	109.10	967.62	42.30	62.70	189.60	9.50	437.00	133.00	93.00	0.16	0.12	0.10
EWR-1	2006/05/15 12:00	8.15	161.70	1489.57	61.10	101.60	303.70	5.79	488.00	284.00	245.00	0.31	-1.00	0.04
EWR-1	2006/06/15 12:00	8.01	236.00	2581.75	98.40	168.50	497.80	4.70	-1.00	651.00	475.00	0.06	-1.00	0.25
EWR-1	2006/08/15 12:00	8.17	235.10	2224.44	79.90	148.40	458.20	2.71	-1.00	528.00	444.00	0.12	-1.00	0.08
EWR-1	2006/09/15 12:00	7.94	199.10	1765.00	60.90	109.40	346.90	1.66	-1.00	360.00	327.00	0.02	0.02	0.02
EWR-1	2006/11/15 12:00	8.31	271.40	1943.73	70.10	110.80	385.10	2.20	133.00	497.00	278.00	0.17	-1.00	0.04
EWR-2	2006/03/15 12:00	6.9	27.0	205.5	23.5	13.30	17.0	6.9	167.0	8.90	2.60	0.120	0.1	0.10
EWR-2	2006/05/15 12:00	7.3	26.0	250.8	28.1	17.50	19.9	6.7	-1.00	8.90	2.50	0.130	-1.0	0.04
EWR-2	2006/06/15 12:00	7.7	26.2	277.1	31.0	20.5	21.9	7.2	-1.00	12.00	3.40	0.110	-1.0	0.06
EWR-2	2006/08/15 12:00	7.7	34.7	347.1	36.1	26.1	26.6	7.0	-1.00	11.26	5.82	0.180	-1.0	0.09
EWR-2	2006/09/15 12:00	8.3	45.1	365.1	35.3	25.7	38.6	6.0	-1.00	27.10	24.90	0.023	0.0	0.04
EWR-2	2006/11/15 12:00	7.7	59.2	449.5	40.2	33.3	49.1	6.9	-1.00	39.90	9.43	0.190	-1.0	0.04
EWR-3	2006/03/15 12:00	8.2	61.4	450.6	46.8	30.2	51.5	5.6	245.0	36.00	35.00	0.070	0.1	0.06
EWR-3	2006/05/15 12:00	7.8	40.4	390.7	38.7	27.4	45.2	4.1	210.0	31.00	34.00	0.160	-1.0	0.08
EWR-3	2006/06/15 12:00	8.5	60.7	635.2	57.9	47.4	77.1	3.8	-1.00	64.00	57.30	0.090	-1.0	0.04
EWR-3	2006/08/15 12:00	8.8	80.3	767.3	55.2	60.2	102.1	3.4	-1.00	87.60	85.10	0.1	-	0.08

SiteName	DateTimeMeas	pH	EC mS/m	TDS mg/l	Ca mg/l	Mg mg/l	Na mg/l	K mg/l	MALK	Cl mg/l	SO4 mg/l	NO3-N mg/l	Al mg/l	N_Amonia mg/l
EWR-3	2006/09/15 12:00	8.5	87.2	718.5	49.1	54.3	94.7	2.6	-1.00	83.00	75.20	0.0	0.014	0.02
EWR-3	2006/11/15 12:00	8.6	101.9	763.0	44.2	57.9	112.4	3.4	-1.00	106.80	58.20	0.2	- 1.000	0.03
EWR-4	2006/03/15 12:00	8.4	50.5	369.5	38.9	24.3	41.4	5.7	200.0	29.00	30.0	0.2	0.088	0.09
EWR-4	2006/05/15 12:00	8.4	38.5	365.5	36.4	26.1	40.4	4.4	198.0	28.00	32.0	0.3	- 1.000	0.08
EWR-4	2006/06/15 12:00	8.5	85.4	613.9	57.3	45.5	74.0	4.1	-1.00	62.00	55.2	0.1	- 1.000	0.05
EWR-4	2006/08/15 12:00	8.9	74.6	719.3	52.4	54.7	93.6	3.3	-1.00	79.90	76.1	0.1	- 1.000	0.06
EWR-4	2006/09/15 12:00	8.6	88.5	701.1	48.2	53.7	92.9	2.8	-1.00	80.00	74.0	0.0	0.0	0.01
EWR-4	2006/11/15 12:00	8.6	104.8	733.6	42.6	54.4	106.3	3.5	-1.00	107.40	62.8	0.2	-1.0	0.04
EWR-6	2006/03/15 12:00	-1.0	-1.0	-1.0	-1.0	-1.0	-1.0	-1.0	108.0	-1.00	-1.00	-1.0	-1.0	-1.00
EWR-6	2006/05/15 12:00	8.5	35.5	311.2	31.8	21.0	34.7	4.4	-1.00	24.00	28.0	0.1	-1.0	0.06
EWR-6	2006/06/15 12:00	8.3	53.3	568.6	52.9	41.9	69.5	4.2	-1.00	56.00	52.4	0.1	-1.0	0.05
EWR-6	2006/08/15 12:00	8.9	85.9	750.9	54.0	57.0	100.2	3.4	-1.00	84.60	77.4	0.1	-1.0	0.05
EWR-6	2006/09/15 12:00	8.6	93.7	725.5	45.4	54.5	95.9	2.8	-1.00	86.00	82.0	0.0	0.0	0.00
EWR-6	2006/11/15 12:00	8.9	106.0	741.6	41.4	57.3	116.0	3.7	-1.00	115.00	69.8	0.3	-1.0	0.04
Fontein-1	2006/03/15 12:00	7.9	55.5	455.1	60.4	38.1	22.9	0.5	280.0	14.36	38.0	0.6	-1.0	0.09
Fontein-1	2006/05/15 12:00	-1.0	-1.0	-1.0	-1.0	-1.0	-1.0	-1.0	167.0	-1.00	-1.0	-1.0	-1.0	-1.00
Fontein-1	2006/06/15 12:00	-1.0	-1.0	-1.0	-1.0	-1.0	-1.0	-1.0	-1.00	-1.00	-1.0	-1.0	-1.0	-1.00
Fontein-1	2006/08/15 12:00	7.3	48.7	-1.0	-1.0	-1.0	-1.0	-1.0	-1.00	-1.00	-1.0	-1.0	-1.0	-1.00

SiteName	DateTimeMeas	pH	EC mS/m	TDS mg/l	Ca mg/l	Mg mg/l	Na mg/l	K mg/l	MALK	Cl mg/l	SO4 mg/l	NO3-N mg/l	Al mg/l	N_Amonia mg/l
Fontein-2	2006/09/15 12:00	6.9	34.3	453.3	59.8	38.6	24.2	0.5	-1.00	11.1	39.0	2.0	-1.0	0.003
Fontein-2	2006/11/15 12:00	6.9	54.9	-1.0	-1.0	-1.0	-1.0	-1.0	-1.00	-1.0	-1.0	-1.0	-1.0	-1
D3H015Q01	2006/05/11 12:00	7.9	29.6	182.619	19.246	10.942	13.998	4.234	82.68	14.8	17.351	0.04	-1	0.02
D3H015Q01	2006/06/01 12:00	8.2	54	415.101	38.989	26.887	37.739	4.627	200.7	33.585	27.604	0.04	-1	0.049
D3H015Q01	2006/07/05 12:00	8.4	78.5	604.239	50.876	41.197	61.04	3.567	278.1	54.92	52.715	0.04	-1	0.02
D3H015Q01	2006/08/02 12:00	8.5	85.1	766.962	54.107	53.414	87.117	3.764	353.7	71.282	65.085	0.04	-1	0.047
D3H015Q01	2006/08/30 12:00	8.5	98.2	844.456	50.252	55.837	103.44	3.46	381.1	80.703	84.908	0.04	-1	0.059
D3H015Q01	2006/09/28 12:00	8.5	93.8	799.722	52.44	57.683	101.7	3.917	356.5	86.239	84.618	0.04	-1	0.02
D3H015Q01	2006/11/01 12:00	8.5	105.6	807.716	39.232	56.229	113.31	3.667	356.9	91.946	67.133	0.04	-1	0.02

8.6 Annexure 6: Mokolo Catchment – Monthly rainfall from Oct 1978 to Sept 2000

Year	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	June	July	Aug	Sept	Average
1978/79	57.9	91.9	58.4	51.1	47.2	80	28	11	0	3	43	0	471.5
1979/80	47.5	74	29.5	117	185	41	22.5	0	0	0	0	11	527.5
1980/81	20	144	74	169	71	34	43	4.8	1.5	0	14	6	581.3
1981/82	12.3	105	38.5	85.5	128	14	33.5	1	0	2	0	15	434.8
1982/83	96.5	57	62	63.5	30.9	49	40	5	10	0	16	2	431.9
1983/84	7	88	88.3	70.5	27.5	84.9	0	0	28	22.5	0	14.7	431.4
1984/85	118.5	57	94.8	150.5	96	117.5	0	10.6	0	0	16	6.6	667.5
1985/86	48	11	149.7	29.1	64.4	13	90	0	0	0	0	5	410.2
1986/87	125.3	67.9	80.4	102.9	73.6	36.4	6.1	0	0	0	3.2	52.9	548.7
1987/88	22.3	62.2	103.3	33.5	126.8	133.5	31.3	0	2.3	0	3.3	21.7	540.2
1988/89	101.9	43.1	77.8	100.9	134.9	135.1	46.6	8	15.1	0	13	0	676.4
1989/90	20	76.1	107.6	46	97.2	83.3	41.8	9.3	0	0	2.3	1.2	484.8
1990/91	23.1	30.9	68.7	149.6	79.2	122.4	5.1	22	2	0	0	43.3	546.3
1991/92	28.2	62.4	96.3	110.8	69	66.2	9.2	0	0	0	0	0	442.1
1992/93	36.5	69.6	90.2	39.3	90.2	102.8	11	0	0	4	1.8	0	445.4
1993/94	47.3	74.8	132.6	116.2	126.8	54	7.1	0	0	0	0	0	558.8
1994/95	18.2	57	63.7	148.5	88.6	108.5	82	30.5	0	0	0	0	597
1995/96	47	69.7	245.5	224.8	102.7	20.4	2.6	64	1	1	0	0	778.7
1996/97	52.8	139.1	46.6	149.6	50.3	139.2	17.2	39	0	0	0	16	649.8
1997/98	26.2	104.8	56.4	62.1	12.5	96.2	2	0	0	0	0	4.3	364.5
1998/99	32.8	206.7	187.1	43	10	62.9	5	53	2	1.6	2	5	611.1
1999/00	29.4	135.2	126.4	177.1	169.1	78.8	114.5	8.4	34	0	1.4	0	874.3

8.7 Annexure 7: Chemistry data (in mg/l) for the Krugersdrift study area - Aug 2011; after Gomo (2011)

Site number	pH	EC	Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	PAIk	MAIk	F ⁻	Cl ⁻	NO ₂ ⁻ (N)	Br ⁻	NO ₃ ⁻ (N)	PO ₄ ⁻	SO ₄ ²⁻	Al ³⁺
BH1	7.53	107	44.5	48.3	122.3	5.6	0	443	0.55	86.2	-0.10	-0.04	0.57	-1.00	37.24	20.4
BH2	7.72	84.8	34.5	41.9	89.0	5.1	0	376	0.52	54.0	-0.01	0.20	0.52	-0.10	21.32	21.7
BH3	7.5	108	36.3	51.3	132.0	5.6	0	477	0.38	71.8	-0.10	0.34	-0.50	-1.00	28.16	23.6
BH4	7.46	87.7	36.7	41.4	94.4	5.3	0	392	0.55	56.0	-0.01	0.26	0.43	-0.10	22.06	21.8
BH5	7.25	168	53.6	107.2	172.2	7.3	0	613	0.51	206.0	-0.10	0.21	-0.50	-1.00	82.34	23.2
BH6	7.3	102	45.2	53.0	114.5	6.1	0	449	0.40	70.7	-0.10	0.30	-0.50	-1.00	28.01	23.7
BH7	7.31	109	43.5	63.5	123.2	6.6	0	470	0.25	84.2	-0.10	0.67	-0.50	-1.00	31.53	23.9
BH8	7.54	124	39.8	59.2	143.9	6.0	0	509	0.56	97.5	-0.10	0.57	-0.50	-1.00	53.30	23.0
BH9	7.36	165	56.4	82.8	209.6	6.6	0	539	0.73	204.5	-0.10	1.40	-0.50	-1.00	105.04	20.3
BH10	7.33	91.9	41.9	47.3	108.9	6.1	0	407	0.62	59.0	-0.01	0.26	0.32	-0.10	25.81	22.2
BH11	7.37	93.7	35.7	42.7	108.7	5.7	0	410	0.50	63.0	-0.01	0.29	0.31	-0.10	25.58	21.5
BH12	7.51	93.6	36.9	44.1	106.9	5.6	0	408	0.46	65.0	-0.01	0.25	0.36	-0.10	25.82	21.8
BH13	7.36	92.5	36.0	41.6	103.9	5.5	0	411	0.43	59.0	-0.01	0.21	0.30	-0.10	24.29	21.1
BH14	7.4	86.6	45.5	45.9	86.4	5.0	0	388	0.35	53.0	-0.01	0.16	0.59	-0.10	20.64	21.6
BH15	7.44	84.8	41.0	41.7	82.9	5.0	0	383	0.33	51.0	-0.01	0.19	0.55	-0.10	22.15	21.3
SP	7.52	135	54.4	74.2	139.2	5.7	0	533	0.26	106.1	-0.10	-0.40	0.11	-1.00	74.83	21.8
R1	7.41	21.8	14.6	4.3	18.2	5.3	0	76	0.11	15.0	-0.01	-0.04	0.85	-0.10	10.24	5.3
R2	7.22	22.3	14.6	4.2	19.1	5.2	0	75	0.08	16.3	-0.01	0.08	1.00	-0.10	10.74	5.2

Summary

Winter (1999) succinctly, made a profound statement to the effect that understanding the basic principles of the interaction between surface water and groundwater is needed for effective management of water resources. Hence, the research investigation was aimed at identifying appropriate methodologies for assessment of surface water - groundwater interaction, thus enhancing the understanding thereof. The methodology used entailed a review of national and international literature on related previous and current models, systems and methods used in assessment and quantification of water exchange between groundwater and surface water. This was then followed by relevant case study analyses where distinct areas were chosen based on availability of relevant data and information by previous investigators. The findings were that various methods and classification systems are widely available but the applicability thereof under the South African conditions depends on the conceptual understanding of the area or system under investigation, availability of data and the basic assumptions associated with the particular model or method. The surface water - groundwater interaction cannot easily be quantified with confidence without requisite data available. The other finding is that use of multiple techniques to reduce uncertainties and to confirm or verify the existence or non-existence of the interaction is essential. Preferably, at least one method should be utilised to trace flow or qualitatively establish the water exchange while the alternative method is used for quantitative estimation of the interaction between surface water and groundwater. Some of the products emerging from this research investigation include decision tables for choosing applicable analytical method, applicable numerical method and the framework for guiding the selection of appropriate methodologies for assessing or quantifying the interaction between surface water and groundwater. Knowledge generated is applicable to water resource management, resource protection, water allocation and monitoring.

Opsomming

Winter (1999) het 'n diepgaande verklaring gemaak tot die effek dat die begrip van die basiese beginsels van die interaksie tussen oppervlak-en grondwater is wat nodig is vir die doeltreffende bestuur van waterhulpbronne. Dus, is die navorsingsondersoek gemik op die identifisering van toepaslike metodologieë vir die assessering van oppervlak water grondwater interaksie, en om die begrip daarvan te verbeter. Die metodologie wat gebruik word behels 'n oorsig van die nasionale en internasionale literatuur oor verwante vorige en huidige modelle, stelsels en metodes wat gebruik word in die bepaling en kwantifisering van water uitruil tussen grondwater en oppervlakwater. Dan was die regte gevallestudie ontleding gemaak, en duidelike gebiede gekies, en op die beskikbaarheid van data en inligting deur vorige navorsers gebaseer. Die bevindinge was dat verskeie metodes en klassifikasie stelsels is wyd beskikbaar, maar die toepaslikheid daarvan onder die Suid-Afrikaanse toestande hang af van die konseptuele begrip van die gebied of stelsel wat ondersoek word, beskikbaarheid van data en die basiese aannames wat verband hou met die spesifieke model of metode. Die oppervlak water grondwater interaksie kan nie met selfvertroue gekwantifiseer word sonder die nodige data. Die ander bevinding is dat die gebruik van verskeie tegnieke om onsekerhede te verminder en om te bevestig of te verifieer die bestaan of nie-bestaan van die interaksie, is noodsaaklik. Moet verkieslik ten minste een metode gebruik word op te spoor vloei of kwalitatief die water ruil vestig terwyl die alternatiewe metode word gebruik vir kwantitatiewe skatting van die interaksie tussen oppervlak-en grondwater. Die produkte wat na vore kom uit hierdie navorsingsondersoek sluit besluit tafels vir die keuse van toepaslike analitiese metode, toepaslike numeriese metode, en die raamwerk vir die begeleiding van die seleksie van toepaslike metodologieë vir die beoordeling of die kwantifisering van die interaksie tussen oppervlak-en grondwater. Kennis wat deur hierdie navorsing gegenereer is op waterhulpbronbestuur, beskerming van die hulpbronne, water toekenning en monitering van toepassing.

