

Application of the Mixing Cell Model to the quantification of groundwater – surface water interaction

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by

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July 2013

Declaration

To the best of my knowledge the submitted dissertation does not contain any material which has been previously published or submitted by one other than myself except where due reference has been given.

I, Amy Jane Matthews, declare that the dissertation hereby handed in for the qualification Magister Scientiae in the Faculty of Natural and Agricultural Sciences, Institute for Groundwater Studies 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 at/in another University/faculty. I further declare that all sources cited or quoted have been acknowledged by means of a list of references. Furthermore, I concede the copyright of the dissertation in favour of the University of the Free State.

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“Learn from yesterday, live for today, hope for tomorrow. The important thing is not to stop questioning.”

- Albert Einstein

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Chapter 1 Introduction

The significance of a reliable groundwater resource assessment is of growing importance as the use of groundwater increases and water resources are stretched to accommodate the growing population. An essential component of a groundwater resource assessment is the quantification of surface water – groundwater interaction. Surface water – groundwater interaction is however a complex component of the hydrological system and this complexity translates into complications in the quantification. A new approach to the quantification of surface water – groundwater interaction is investigated in the hopes of creating a pathway to improving the understanding of this interaction.

1.1. The problem of quantification

Surface water – groundwater interactions take place via different mechanisms on varying scales and are influenced by numerous processes. The complexity of these interactions makes the quantification of the actual volume moving between the two water resources problematic. There are numerous methods available for the quantification of the amount of groundwater contributing to a rivers baseflow, lakes or wetlands as well as methods for quantifying the loss of water from a losing stream. However, surface water – groundwater interaction is still poorly understood and difficult to quantify due to the inherent heterogeneity of aquifers, variable influencing factors, different time scales of surface water and groundwater, and the fact that groundwater is a hidden resource that cannot be directly measured in most cases (Sophocleous (2002); Eijkelenburg (2004); Kirk (2006); Kalbus, *et al.* (2006); Hughes, *et al.* (2007); Levy and Xu (2012)).

South Africa has a history of preferential use of surface water to supply the country's water needs, which is evident in the number of dams which cover the countries river systems and associated infrastructure including large scale transfer schemes. This preference is also evident in the methods available and used to quantify the countries water resources, where groundwater has been sorely neglected. Hydrological methods, such as the Pitman model, have been the most popular methods utilised in South Africa. However, as available surface water resources are pushed to their limits with more dams and water transfer schemes constructed, groundwater usage has become increasingly prevalent and so have research efforts to quantify this resource.

1.2. A new approach

In light of the persisting lack of understanding of surface water – groundwater interactions, the importance of the groundwater contribution to streamflow and the increasing use of groundwater, a new approach to the quantification of this is proposed. Although multiple methods exist for the quantification of the groundwater contribution to streamflow, the addition of the proposed method will be advantageous. The method would be beneficial in terms of using a different dataset comprising water quality data and as part of a multi-method approach which has been suggested by numerous authors (Oxtobee and Novakowski (2002); Environment Agency (2005b); Rosenberry and LaBaugh (2008); Allen *et al.* (2010); Levy and Xu (2011); Sophocleous (2002); Kalbus, *et al.* (2006)).

The method of quantifying the groundwater contribution to streamflow currently used in the latest Groundwater Resource Assessment (GRA2) of South Africa is based on a water balance approach alone, while the proposed new method combines the water balance with solute mass balances. The incorporated solute mass balances serve to better constrain the water balance used to quantify the groundwater baseflow. However, the concept of using two sets of mass balance equations simultaneously is not a novel idea. The use of the basic principal and the Mixing Cell Model (MCM) are also fairly common, but the use of the MCM to quantify the groundwater component of streamflow is an innovative application. The suitability and precision of the MCM to the proposed use of quantifying groundwater – surface water interaction is investigated by applying the method to a number of test sites and comparing the results with traditionally used methods.

1.3. Thesis Structure

The thesis is structured as follows:

- ↳ Chapter 2 is a general overview of surface water – groundwater interaction to create a better foundation for evaluating the proposed method. The overview covers the basic principles, influencing factors and the various reasons for the complex nature of this interaction.
- ↳ Chapter 3 reviews a number of available methods of surface water – groundwater interaction investigation from an international and local perspective. Specific attention has been given to the methods of quantification presently used in the groundwater

resource assessment of South Africa. The historical applications of the Mixing Cell Model (MCM) are also covered and discussed within this chapter.

- ↳ Chapter 4 covers the methodologies applied in the study. The MCM is described in terms of the basic concept, the mathematical methodology, the software programme used and its slight adaption for the application to surface water – groundwater interaction. The methodology of the chemical hydrograph separation method used is additional given as well as a short description of the field work performed.
- ↳ Chapter 5 contains the three pilot study area investigations. The MCM is applied to datasets from the surface water – groundwater interaction test site developed by the University of the Free State and data collected along the middle Modder River during a fieldwork survey. The MCM is subsequently applied to a set of quaternary catchments in the Limpopo Province that have calibrated estimates of groundwater baseflow for the Sami and Hughes models. The MCM is lastly applied to the quaternary catchment D73F, located in the semi-arid Northern Cape, to assess the applicability of the algorithm-based MCM in a regionally-defined zero groundwater baseflow zone. Each pilot study comprises of a general overview of the area, conceptualisation for the MCM application, results, and discussion and comparison section.
- ↳ Chapter 6 is a general discussion of the MCM results including discrepancies found and model limitations imposed by the scope of the study.
- ↳ Chapter 7 covers the main conclusions made from the results of this project.
- ↳ Chapter 8 is a description of the consequential recommendations for both the application of the MCM and further investigation regarding the MCM that is required.
- ↳ Appendices A – E accompanying this study, include a step-by-step guide for a MCM application using the MCMsf programme, water quality data used in the MCM runs and the detailed errors associated with each model run.

Chapter 2

Basic principles of Surface water – Groundwater Interaction

A general background into what is surface water – groundwater interaction is given to create a better foundation for evaluating a method that aims to quantify this interaction. The overview includes the basic principles of surface water – groundwater interaction, influencing factors and the various reasons for the complex nature of this component of the hydrological cycle.

A river receives water from a number of sources, varying from direct rainfall to discharge from the adjacent aquifers. The three main sources are overland runoff, interflow and groundwater inflow (Figure 2-1). Overland runoff occurs mostly during storm conditions where precipitation infiltrating into the soil has resulted in the soil capacity being reached. Once the soils capacity has been reached any additional precipitation will flow over the land surface in response to the gradient of that land surface, usually flowing towards the low-lying river valley. The larger the gradient the more likely runoff will occur. On the other hand, water infiltrated into the soil layer will percolate through the unsaturated zone towards the saturated zone where the water becomes groundwater by definition. However, the water within the unsaturated zone may not reach the groundwater as lateral movement through the unsaturated zone can also occur. This lateral movement, known as interflow, can be in response to a number of factors including a steep gradient or the intersection of an impermeable layer. Interflow will discharge where the land surface is intersected allowing this water to reach a stream without ever entering the groundwater zone. Groundwater contributing to a stream is defined as water that has percolated into the subsurface, reached the saturated zone and then moved within this zone to a river where it is discharged directly. Water reaching the river by this mechanism is known as groundwater baseflow and tends to sustain streamflow during dry periods. Groundwater baseflow was traditionally defined as the total baseflow to a stream, but interflow has been found to also substantially contribute to the baseflow of a river. Thus, the baseflow of a stream is considered to comprise of both interflow and groundwater contributions. Figure 2-1 is a conceptual representation of the various flows into a river system including the discussed overland flow, interflow and groundwater baseflow.



Figure 2-1 The various water sources to a river system. Water can reach the river by means of overland flow, interflow and groundwater discharge (Taken from Schreiber-Abshire, *et al.*, 2005).

2.1. Basic types of interaction

In the past, surface water and groundwater were seen as separate water resources and dealt with individually. However, in more recent times the inter-connectedness of these two resources has become evident. Surface water – groundwater interaction is the general term used to describe this inter-connectedness. The actual movement of water comprising this interaction has many forms and is highly variable, but the two main differentiated types of surface water – groundwater interaction are effluent (gaining) streams and influent (losing) streams. A gaining stream is defined as a river that is fed directly by groundwater, forming part of the rivers baseflow (Figure 2-2a). A losing stream is defined as a river which is losing water to the underlying aquifer through the stream bed (Figure 2-2b).

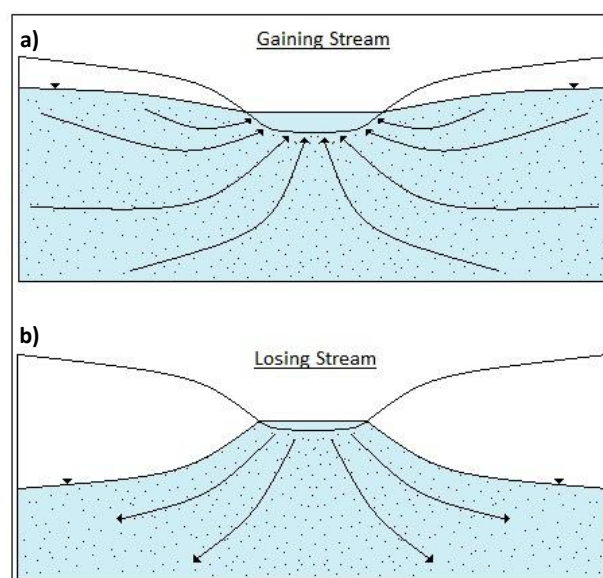


Figure 2-2 The main surface water – groundwater interaction types. a) A gaining stream receiving groundwater from the underlying and adjacent aquifer due to the water table being higher than the river stage. b) A losing stream discharging water into the underlying aquifer due to the river stage being higher than the water table (Modified from USGS (1998)).

Surface water – groundwater interaction along a non-theoretical river course is often not as simple and cannot be defined by one interaction type alone. A river course can change from a gaining stream to a losing stream or *vice versa*, numerous times. The surface water – groundwater interaction of a losing stream can be further divided into connected and disconnected streams. A connected losing stream is shown in Figure 2-2b, where the river is directly connected to the underlying aquifer and the water table. A disconnected losing stream does not have a direct connection to the underlying aquifer as the unsaturated zone separates the two (stream A in Figure 2-3). There is a localised upwelling in the water table below a losing stream. From Figure 2-3 it can be seen that a gaining and losing stream vary with their relative position to the water table. A gaining stream's river stage is below the water table, while a losing stream's river stage is above the water table.

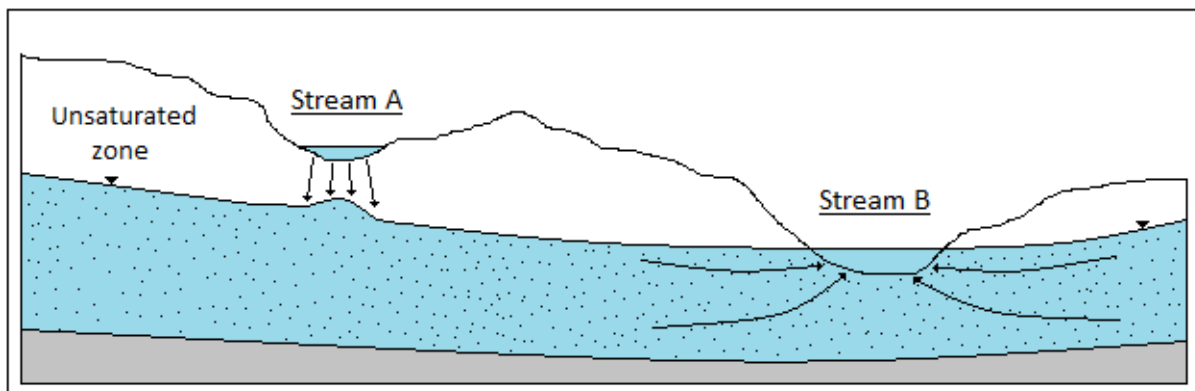


Figure 2-3 A landscape division of surface water – groundwater interaction types. Stream A is a disconnected losing stream with the river stage positioned above the regional groundwater table, while Stream B is a gaining stream with the river stage positioned below the regional groundwater table.

2.2. Surface water – groundwater interactions for different landscapes

2.2.1. Mountainous Upper course (Headwaters)

Surface water – groundwater interactions vary depending on which course of the river is investigated. The surface water interactions taking place along the upper course of a river (headwaters) will vary slightly to the interactions taking place along the lower course of a river. The upper course, usually located in a mountainous area, is characterised by highly variable precipitation and water movement over and through the steep slopes alongside the river. Along the steep slopes of the v-shaped river valley the flow of water to the stream can occur by three different mechanisms (Figure 2-4). These flow mechanisms include runoff, interflow and groundwater discharge to the stream. Interflow and runoff occur more rapidly here during precipitation events than in the lower courses of the river, due to the steep slope along the river. Groundwater is the main source of baseflow when there is no precipitation. Water

percolates through the unsaturated zone reaching the water table and moves in response to a hydraulic gradient towards the river (Figure 2-4a). When a precipitation event occurs, rainfall infiltrates the top soil layer and due to the steep slope flows within this layer towards the river in the form of interflow (Figure 2-4b). The additional water in the soil layer flowing towards the river will create a mound in the water table resulting in water being discharged along the river banks. After a period of rainfall, the soil will reach its field capacity and runoff will occur due to the steep slope as shown in Figure 2-4c (USGS, 1998).

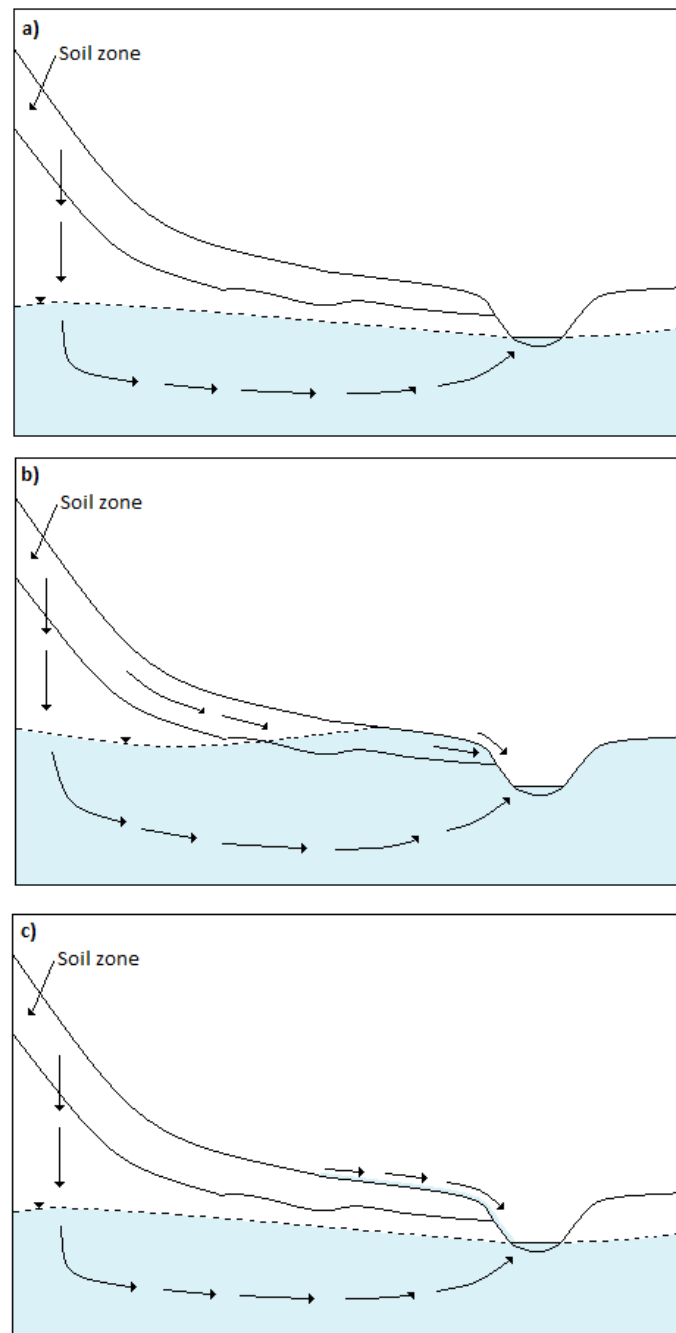


Figure 2-4 The different flow types contributing to the baseflow of an upper course river section. a) Groundwater contributing to the streams baseflow during low flow conditions. b) Interflow and groundwater contributing to baseflow during the beginning of a rainfall event. c) Runoff and groundwater contributing to stream baseflow after a period of rainfall (Modified from USGS (1998)).

2.2.2. Level, Lower course (Flood plains)

The mechanisms by which water contributes to a lower course river section will be different to those occurring in the upper course. Rivers have a much wider river valley in the lower courses and well-developed flood plains, resulting in a less steep gradient towards the river. An increasing extent and density of the riparian vegetation tends to characterise the middle and lower courses of a river when compared to the upper courses. Surface water – groundwater interaction in the lower course of a river is mainly affected by the interchange of local and regional groundwater flow systems, flooding and evapotranspiration. Groundwater from the regional flow system discharges directly to the river as well as various places across the flood plain (Figure 2-5). Wetlands or small lakes can be formed due to terraces present in the alluvial valley having their own local groundwater flow systems. These small local groundwater flow systems overlie a regional groundwater flow system which complicates the hydrology of the river. The contribution of two different groundwater sources to the river and floodplain is further complicated when recharge from flood waters are superimposed on these systems (USGS, 1998).

In most rivers' lower courses the water table is close to the land surface in the river valley (Figure 2-5). Vegetation along the river and in the floodplain is likely to have root systems which intersect the water table, resulting in the plants transpiring at their maximum potential rate using water directly from the groundwater system. The water taken up by these plants causes a drawdown in the water table resulting in the plants intercepting groundwater that would have contributed to the rivers baseflow. If the riparian vegetation is extensive, in the growing season, a large drawdown in the water table which could even cause infiltration of river water into the subsurface (USGS, 1998).

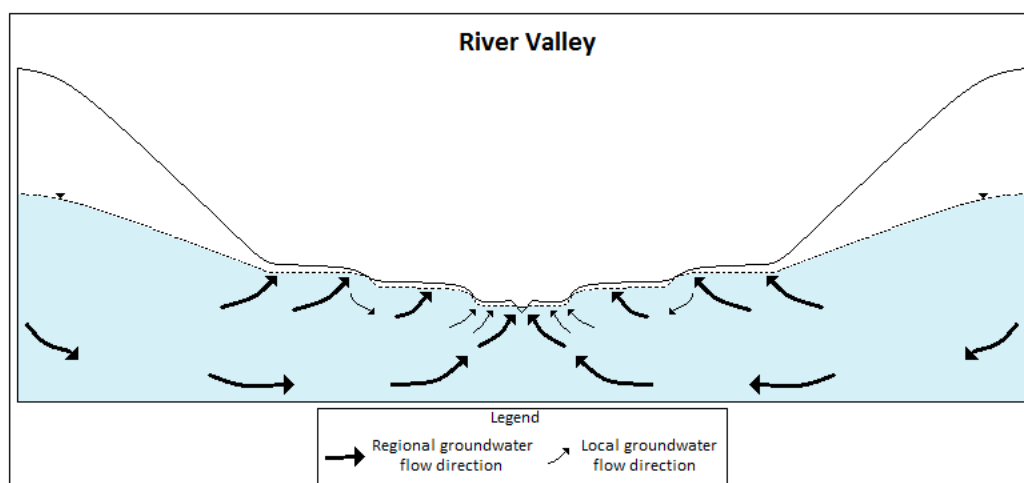


Figure 2-5 Regional and local groundwater flow systems in a lower course river and the interaction of these two systems taking place within the alluvial flood plain (Modified from USGS (1998)).

The alluvial channel deposits along the lower courses of a river are far more extensive than in the upper reaches. The alluvial channel aquifer comprises of unconsolidated sediments allowing for the rapid transport of water within this medium (Figure 2-6). Flow within these sediments can also be parallel to the river, leading to a downstream movement of groundwater instead of towards the river itself.

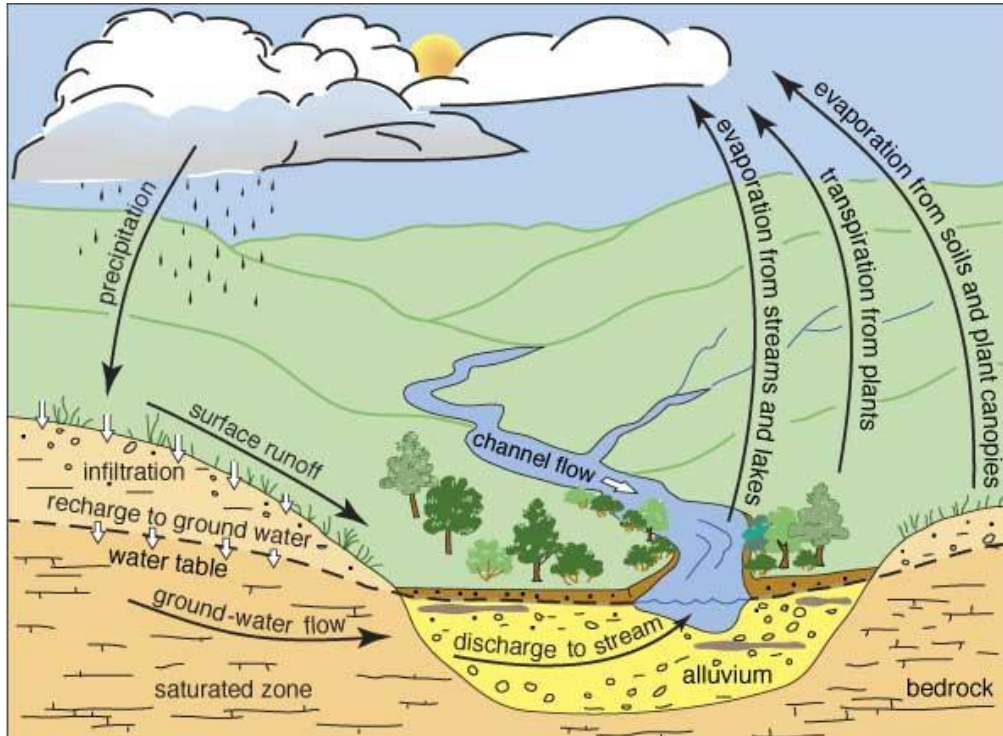


Figure 2-6 A conceptual representation of the alluvial aquifer along the lower courses of a river and the various water flow mechanisms taking place here. Two different systems result, namely the adjacent alluvial channel aquifer and the terrestrial bedrock aquifer (Taken from Suchy, *et al.*, 2005).

These alluvial channel deposits form a subsurface zone of sediment in which stream water readily exchanges. This zone of interaction is commonly referred to as the hyporheic zone. The hyporheic zone is the best known location of surface water – groundwater interaction. Water is exchanged between surface water and groundwater systems through this physical, chemical and biological filter (White, 1993; Hancock, 2002;). The type of surface water – groundwater interaction taking place along a river will affect the ecology of the hyporheic zone (Figure 2-7). In a gaining stream, where upwelling of groundwater through the hyporheic zone is contributing nutrient-rich water to the river, an increased productivity of the organisms is generated. In a losing stream, where downwelling of river water occurs, the sediments of the hyporheic zone are well-oxygenated, rich in labile carbon and host diverse faunal assemblages supporting the micro-scale ecosystem of the groundwater system (Hancock, *et al.*, 2005). The biota related to the upwelling of groundwater into the river has even been used to identify zones of focused

groundwater discharge (Alley, *et al.*, 2002). Localized flow systems are also supported by the hyporheic zone (Figure 2-8). These flow systems can be caused by local geomorphologic features such as stream bed topography, stream bed roughness, meandering or heterogeneities in sediment hydraulic conductivities. These flows systems allow the exchange of water across the interface between surface water and groundwater by the flow in and out of stream beds and banks forming the hyporheic zone (Alley, *et al.*, 2002).

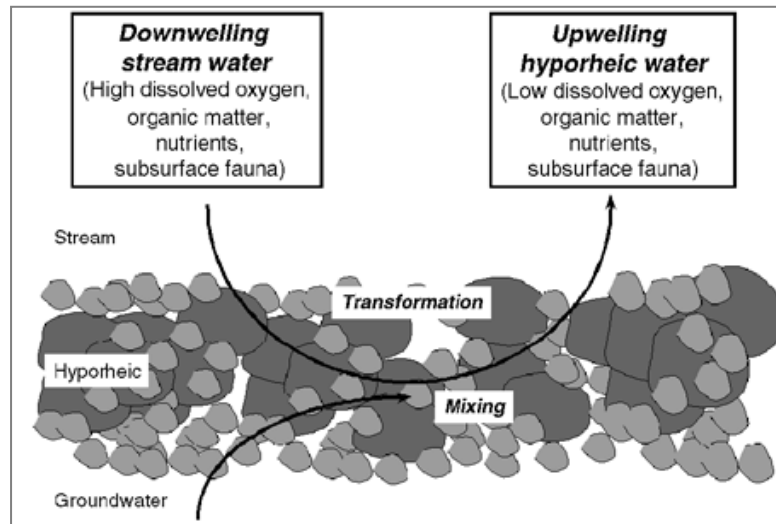


Figure 2-7 The effects occurring in hyporheic water and fauna in response to the direction of the surface water – groundwater interaction taking place (Taken from Hancock, *et al.* (2005)).

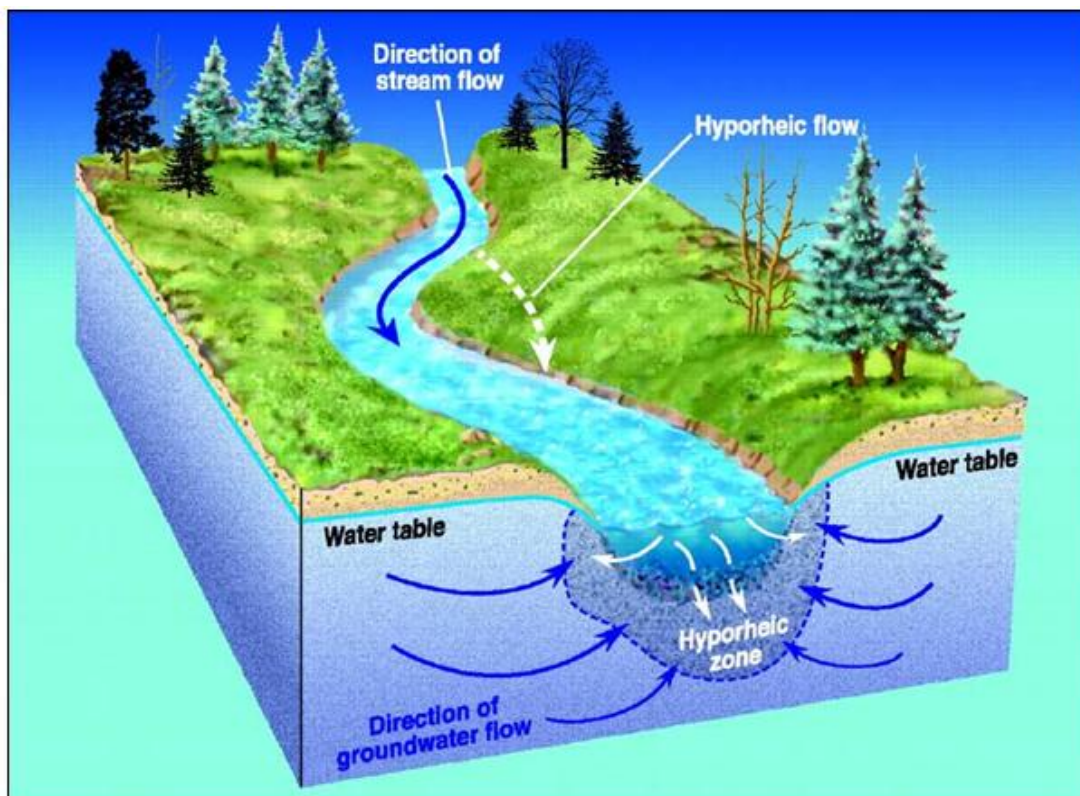


Figure 2-8 The hyporheic zone, beneath and adjacent to a lower course river, where surface water and groundwater mix and through which localised flow systems are formed due to geomorphological conditions such as meandering (Taken from Alley, *et al.*, 2002)

A further complication to surface water – groundwater interaction investigations is the different perspectives from each discipline, namely ecology, hydrology and hydrogeology. This difference in perspective is clear when addressing the hyporheic zone (Figure 2-9). From Figure 2-9 it can be seen that ecologists define the hyporheic zone as a fluctuating habitat, hydrologists define the hyporheic zone as an area where surface water interacts with the subsurface and hydrogeologists define the hyporheic zone as an area of mixing of surface water and groundwater and a zone through which surface water – groundwater interaction occurs (Withtüsler, 2006).

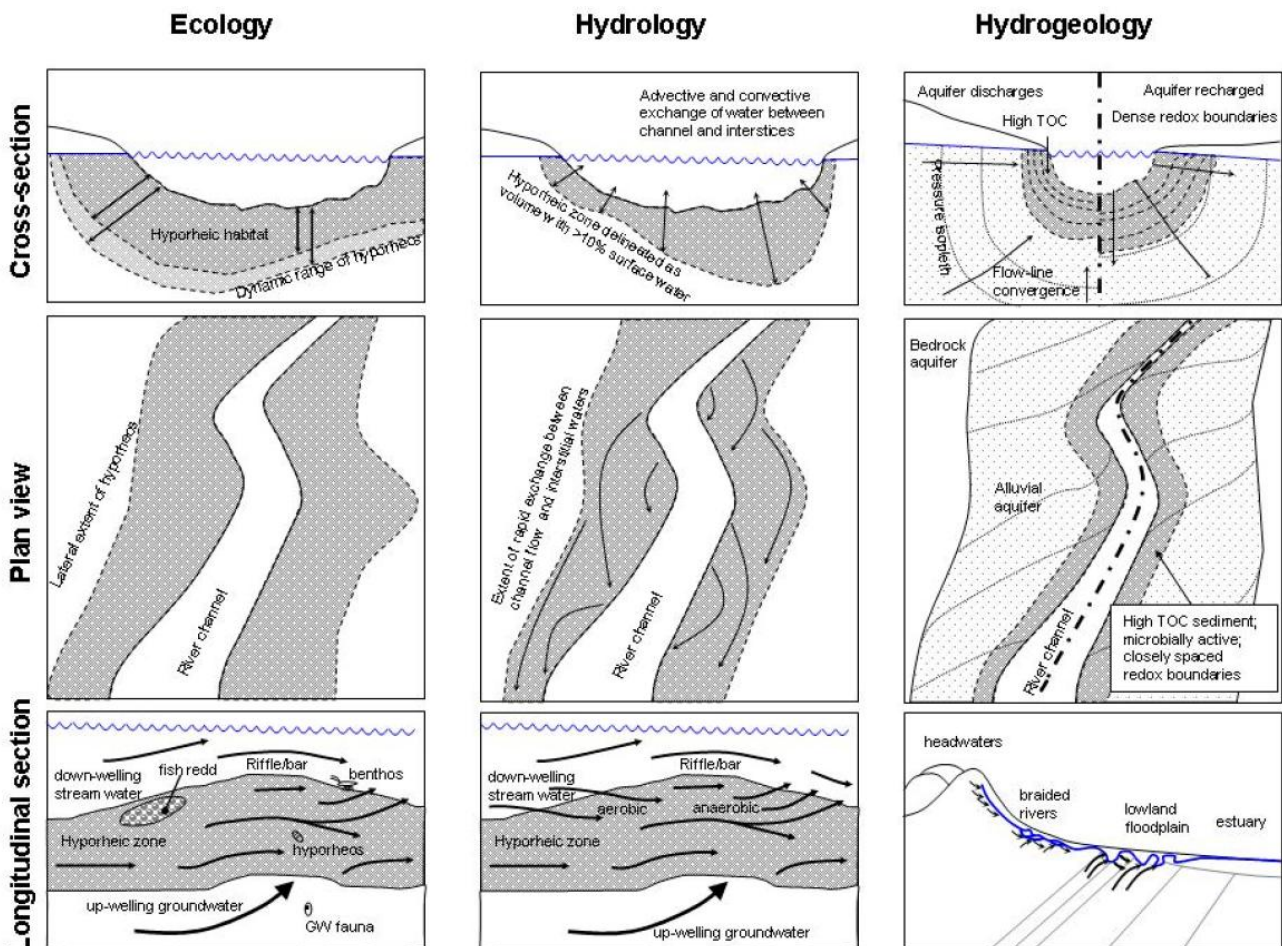


Figure 2-9 Conceptual models of the hyporheic zone from different research disciplines (Taken from Environment Agency, 2002).

The geomorphological differences between an upper course and lower course river section result in different flow mechanisms which have been seen to influence the surface water – groundwater interactions taking place there. Xu, *et al.* (2002) developed a hydrogeomorphological classification system to assist with the separation of the groundwater component of streamflow using hydrographs. The geomorphological types were based on the different courses of a river and shown in Figure 2-10. The upper course type (Type 1) is classed

as interflow dominated, while the lower course type (Type 3) is classed as a groundwater discharge zone. This geomorphologically-based classification of streams for the quantification of the groundwater contribution to streamflow confirms that this aspect plays an important role in the dynamics of the interaction taking place along the river.





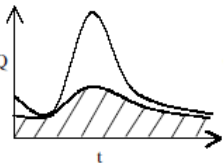
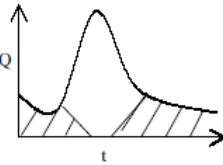
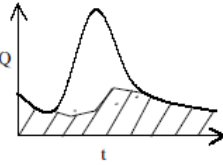
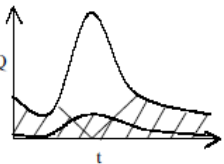
<i>Hydrogeomorphic Types</i>				
Geomorphologic typing	Type 1: upper catchment	Type 2: middle course	Type 3: lower catchment	Type 4: Special cases
Interaction scenarios	Interflow dominant in upper catchment	Intermittent in middle course	Groundwater discharge zone	Under some circumstances
Hydraulic connection				
Baseflow separation concept				

Figure 2-10 Hydrogeomorphologically defined types based on the different courses of a river and their various surface water – groundwater interaction characteristics (Modified from Xu, *et al.*, 2002)

2.3. Influencing factors

A number of factors influence surface water – groundwater interactions taking place in all landscapes, such as meteorological conditions, evapotranspiration, preferential pathways and abstraction. There are many more influencing factors such as the geometry of the streambed, clogging layers in the streambed, properties of the vadose zone, flow durations, presence of other water sources and many more that are not covered within the scope of this study (Witthüser, 2006).

2.3.1. Meteorological

Changing meteorological conditions and differences in topography affect surface water – groundwater interaction. Infiltrating precipitation tends to create localised mounds in the water table adjacent to surface water bodies and at low-lying points in the landscape where the unsaturated zone is thinner (Figure 2-11). These mounds in the water table caused by focused recharge can reverse the gradient between the surface water and groundwater levels. The reverse in gradient could result in increased groundwater discharge to surface water bodies, or it can cause losing streams to become gaining streams. A mound in the water table adjacent to

the river is specifically called bank storage. Bank storage is formed in response to a rapid rise in the river stage usually caused by storm precipitation or water released from an upstream reservoir. However, the water lost to this adjacent river bed will slowly return to the river as long as the river stage does not surpass the river banks (USGS, 1998).

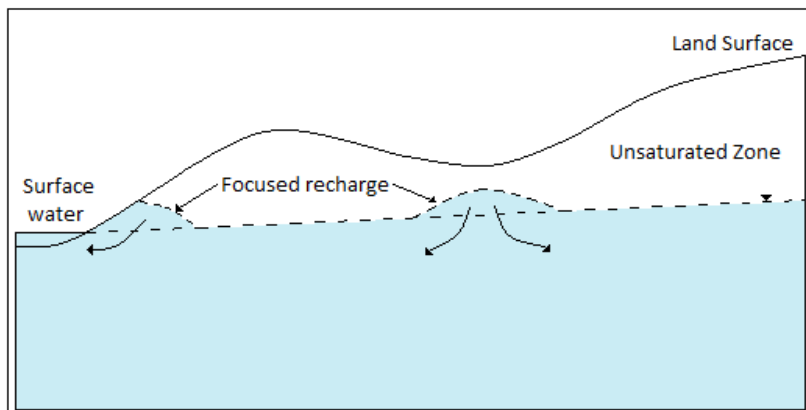


Figure 2-11 Focused recharge in response to changing meteorological conditions, resulting in increased groundwater inflow to a river (Modified from USGS (1998)).

2.3.2. Evapotranspiration/Riparian Vegetation

Evapotranspiration also has the ability to change the direction of the interaction flux. Riparian vegetation adjacent to a river can lower the water table because plants roots penetrating the saturated zone can directly transpire groundwater. The drawdown in the water table can reverse the gradient alongside the river, causing an initially defined gaining river to become a losing river (Figure 2-12). This reverse in gradient can greatly reduce the amount of groundwater contributing to the streams baseflow. However, the draw down in the water table due to riparian vegetation evapotranspiration is highly variable and closely related to the growth seasons of the riparian vegetation (USGS, 1998).

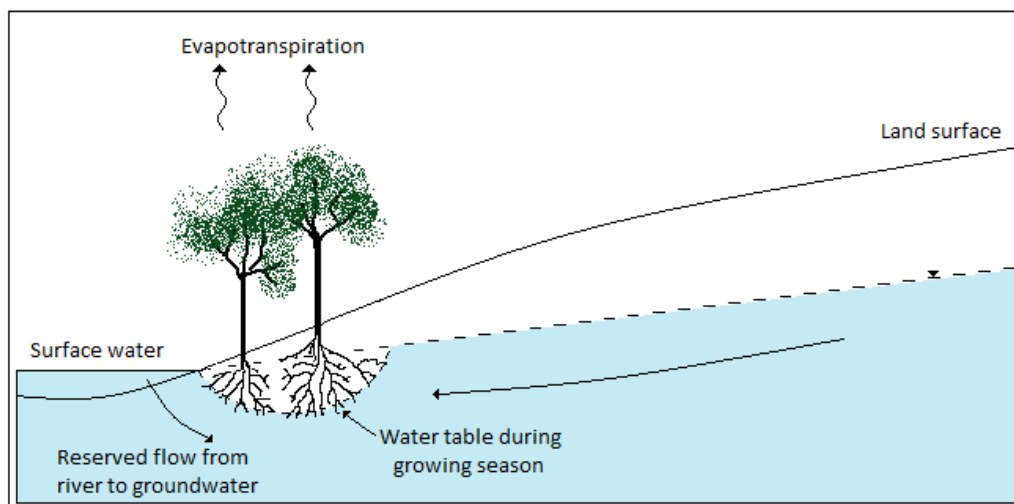


Figure 2-12 Evapotranspiration from riparian vegetation can cause a drawdown in the water table alongside the river, reserving the gradient and the interaction flux.

The riparian vegetation alongside a river can also serve as a pathway for preferential recharge (Department of Water Affairs and Forestry, 2005). The riparian vegetation increases the amount of water infiltrating into the subsurface by a number of mechanisms such as reduced runoff due to interception of precipitation and preferential pathways created by root systems and animal burrows. The increased infiltration leads to an increase in the amount of water available to percolate down to the saturated zone and eventually the amount of water recharging the groundwater. Nelle (2004) refers to the riparian zone as the riparian sponge because it has the ability to absorb, store and then slowly release stored water over an extended period. The riparian vegetation thus also plays an important role in the interaction of water within the river valley where the vegetation can be pumping water out of the system via transpiration or it can be increasing the amount of water entering the system.

2.3.3. Preferential pathways

Secondary structures such as fractures, faults and joints create pathways within primary aquifers for preferential flow to take place therein. Approximately 98% of the aquifers in South Africa are classified as secondary aquifers (IWR, 2011; Parsons, 2004). These secondary aquifers comprise of mostly fractured-rock which supply groundwater through the openings created by the fractures within the hard rock. This forms two different flow systems, namely a slower diffusion of groundwater through the rock matrix and a faster flow through the fractures in the aquifer. The openings occur in a highly irregular fashion which complicates the prediction of aquifer properties and the simulation of the aquifer (Talma and Weaver, 2003). The heterogeneity of the fractured-rock aquifers tends to limit the use of some methods traditionally used for characterising porous-media aquifer systems (Cook, 2003). Determining the volume of groundwater discharge to a river is more complicated in a fractured-rock aquifer due to the groundwater inflows from irregularly spaced fractures (Cook, 2003; Levy and Xu, 2011). Cook (2003) recommends quantifying this volume by either measuring the discharge of streams that drain the fractured rock catchments or by measuring concentrations of various solutes within the stream and applying solute mass balance methods.

2.3.4. Groundwater abstraction

Groundwater abstraction from a shallow aquifer that has a direct connection to a surface water body can greatly influence the surface water – groundwater interactions taking place. The number of abstraction points over an area will determine the scale of the impact, ranging from a local impact for a small numbers of wells to a regional impact for a large number of wells. The withdrawal of water from a shallow aquifer near surface water bodies can impact on the available surface water resources by capturing groundwater flow that would have discharged to the surface water body otherwise or by inducing flow from the surface water body into the subsurface (Figure 2-13). A groundwater system under pre-development conditions or no abstraction conditions is in a steady state where the amount of recharge entering the system is equal to the amount of groundwater discharged (Figure 2-13a). Once abstraction from a constructed borehole has started the shallow aquifer groundwater flow system is altered (Figure 2-13b). When a well is pumped in close proximity to a river, it initially obtains water from the water stored in the aquifer and creates a cone of depression of the potentiometric head. The resulting gradients intercept some of the regional groundwater flow, which otherwise would have discharged into the river (Witthüser, 2006). This abstraction of groundwater that would have otherwise reached the river is commonly referred to a baseflow reduction. In Figure 2-13b two abstraction points are shown, one close to the river and one further away. Initially or for sustainable abstraction rates the river will remain a gaining stream although some groundwater is captured by abstraction. If groundwater is continually abstracted at non-sustainable abstraction rates the associated water table drawdown will be extensive (Figure 2-13c). The large drawdown in the water table can result in the failure of borehole water supply as seen in Figure 2-13c, as well as a reverse in the water table gradient causing a gaining stream to become a losing stream. Once the cone of depression reaches the stream, it induces flow from the stream into the aquifer which is commonly referred to as induced recharge or induced stream infiltration (Witthüser, 2006). The borehole close to the river has induced river flow into the subsurface and is thus directly abstracting from the surface water resource. In this manner groundwater abstraction from boreholes close to a surface water body have the potential to negatively impact the surface water resources (USGS, 1998).

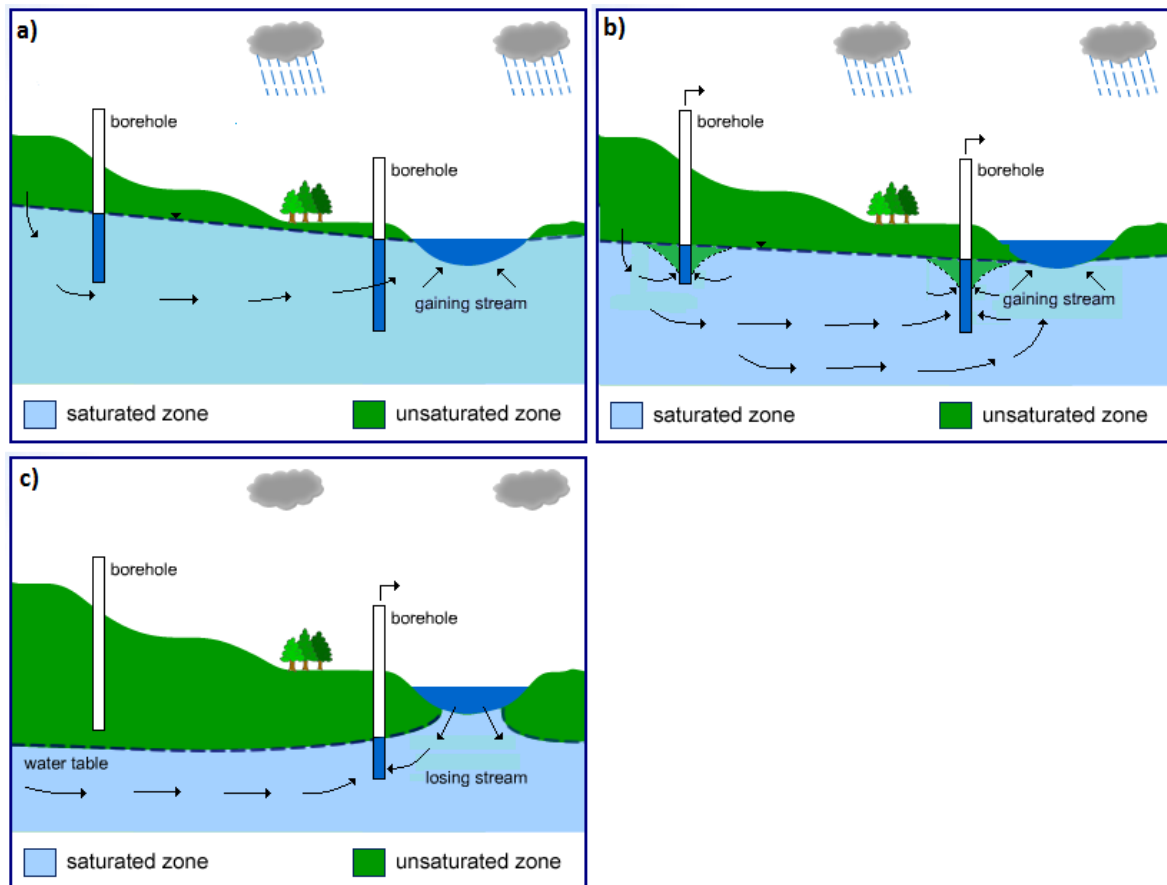


Figure 2-13 The process by which groundwater abstraction from boreholes located near a river can negatively impact surface water resource. a) Under pre-development conditions the groundwater flow system is at equilibrium. b) Sustainable groundwater abstraction will slightly impact the surface water resource, but the river will remain a gaining stream. c) Large groundwater abstractions can cause a drastic drop in the water table causing water supplies to fail and induce river flow into the subsurface (Modified from DWA, 2011).

Groundwater abstraction within a secondary, fractured-rock aquifer is more complex in that there are two flow systems that the borehole can intersect, the matrix or the fracture network. The intersection of a fracture by a borehole substantially increases the yield of that borehole as water moves faster along these openings than within the rock matrix. A groundwater abstraction point within a fractured rock aquifer located further than 100m away from a river could still have a dramatic effect on the river due to a direct link via fractures.

Chapter 3 Literature Review

Numerous methods for quantifying surface water – groundwater interactions exist, ranging from the simple and site-specific to complex and extensive. A number of these available methods are described and discussed on both an international and national perspective. The methods currently used in the latest South African Groundwater Resource Assessment are reviewed and described in detail. The application history of the Mixing Cell Model (MCM) is also investigated and discussed.

3.1. International Approaches and Methods

3.1.1. Guidelines

The Australian government initiated the Water for the Future – Water smart Australian programme in order to aid in integrated water management and the quantification of double accounting in water resource assessments. The project's objectives were to develop a practical and moderately priced methodology for assessing the different connections between groundwater and river systems. The project compared estimates of surface water – groundwater interaction using flow differences, hydraulic gradient analysis, hydrograph baseflow separation and geochemical comparisons in ten representative catchments. A method of quantifying the surface water – groundwater interaction was subsequently recommended in the 2012 final report based on a predefined level of importance of a water resource. For low importance groundwater and surface water systems, a groundwater balance method is recommended. For catchments with moderate importance groundwater and surface water resources, baseflow separations using the Tracer method and Lyne and Hollick Filter method are recommended. For high importance water resource systems, baseflow separations using the Tracer and Lyne and Hollick Filter methods complimented with run of river sampling methods would be the minimum recommendation. However, it is important to note that the higher the accuracy of a surface water – groundwater interaction assessment is, the higher the cost will be. The indicative costs per catchment for a poor to moderate, moderate to high, moderate to high (instrumentation), high to excellent and high to excellent (instrumentation) are \$10 000, \$20 000, \$85 000, \$150 000, and \$500 000, respectively. It was concluded that the chemical hydrograph separation method (Tracer method) is sensitive to the groundwater and surface water end members applied but the method has the best potential for providing reasonable

catchment scale estimates of groundwater inflow to a river over time (Australian Government, 2012a and 2012b).

The National Water Initiative (NWI) is the Council of Australian Government's principal water policy agreement. One of the main objectives of the NWI is the conjunctive management of surface water and groundwater resources. In light of this, the Groundwater Project of the eWater Cooperative Research Centre is developing modeling tools which will incorporate a surface water – groundwater interaction capability for the new RiverManager© and WaterCast© products (Australian Government, 2004). Rassam and Werner (2008), in a comprehensive review of surface water – groundwater interaction modeling approaches and their applicability to Australia, found that groundwater – surface water interactions are poorly handled in existing surface water and groundwater models. In river models the interaction volume is simply modeled as a loss term where as in groundwater models the river is simply modeled as a boundary condition. In more sophisticated models, able to account for the interaction more explicitly, more data and a higher degree of modeling expertise is usually required. Rassam and Werner (2008) thus suggest that surface water – groundwater interaction processes that are most relevant to the Australian landscape should be identified to facilitate the selection of a modeling tool which will incorporate an appropriate balance between surface water and groundwater processes. It follows that this balance can only be achieved through the use of custom-built, special-purpose models developed to answer particular management questions. Jolly, *et al.* (2008) summarise the research done by the eWater Cooperative Centre and describe three simplified modelling approaches that are currently in development, namely a reach scale model, 'Groundwater-Surface Water Link', which operates as a groundwater link to river models and accounts for interactions at the river-reach scale; a sub-reach scale model, 'Floodplain Processes', which dynamically models bank storage, evapotranspiration, and floodplain inundation enabling a more refined modelling of surface water – groundwater interactions, and can be linked to ecological response models; and a catchment scale model that estimates the surface and sub-surface flow components to streams (Jolly *et al.*, 2008).

The Environment Agency is an executive, non-departmental public body which aims to protect and improve the environment in England and Wales (House of Commons, 2006). The Agency has a legislative duty to manage the sustainable development of groundwater resources. Conceptual and numerical model development is the main objective of the Agency in order to efficiently meet their regulatory responsibilities. The Agency currently invests £3 million per

year on groundwater resource assessments and modelling (Environment Agency, n.d. (a)). The need for a regional groundwater conceptual or numerical model has been identified for selected areas in England and Wales, mainly in major aquifers. Groundwater resource assessment and modelling is an iterative process beginning with the development of a conceptual model which is used as a basis for testing ideas and to identify data and knowledge gaps. The conceptual model is then refined when new data or understanding of the area improves. If there is sufficient data and a need, the groundwater modelling process can be taken further by developing a numerical model which is a computer-based representation of the conceptual model. The numerical model is then used to make predictions which aid in making decisions regarding the management of groundwater resources (Environment Agency, n.d. (a)). The Lowland Catchment Research (LOCAR) and Catchment Hydrology and Sustainable Management (CHASM) programmes have resulted in considerable field-based activity investigating surface water – groundwater interactions, forming sixteen field test sites (Environment Agency, 2005a). Resources Assessment Methodology (RAM) and Impact of Groundwater Abstractions on River Flows (IGARF) are two of the tools utilized by the Environment Agency to support their management and protection of groundwater. RAM sets the resource availability status for river reaches and associated groundwater. IGARF evaluates the effects of groundwater abstraction on surface water flows (Environment Agency, n.d. (b)).

Rosenberry and LaBaugh (2008) compiled a comprehensive overview of available techniques and methods to describe and quantify surface water – groundwater interaction as part of a U.S. Geological Survey and U.S. Department of the Interior project. The report's objectives were to create an awareness of the scope of the methods available as well as to serve as a guide to surface water – groundwater interaction studies for water-resource investigators. The report covers scale appropriate methods and an in-depth description of most methods. LaBaugh and Rosenberry (2008) suggest watershed-scale modelling, groundwater flow modelling, flow-net analysis or dye and geochemical tracer tests for catchment scale studies, defined as larger than a kilometre or more in length or width. The measurement of streamflow at two places over an intermediate scale (ten to hundreds of meters) which enables the calculation of gains and losses in that river reach is recommended for the identification of interaction zones. Tools such as seepage meters, mini-piezometers and buried temperature probes are more appropriate and recommended by LaBaugh and Rosenberry (2008) for local, small scale studies.

3.1.2. Hydrograph Separation Techniques

The unit hydrograph separation method distinguishes between streamflow originating from surface runoff and groundwater. The method is popular as it only requires readily available streamflow data (Australian Government, 2012b). The widespread method of estimating fluxes to and from groundwater aquifers using streamflow data traditionally starts with using the measured rainfall at the surface and then estimating infiltration, redistribution, evapotranspiration, percolation of residual water through the unsaturated zone and discharge of groundwater to streams, respectively (Wittenberg and Sivapalan, 1999). Wittenberg and Sivapalan (1999) refer to this approach as reductionist or “bottom-up” approach in a report for the Centre for Water Research, University of Western Australia. However, these approaches are not suitable for arid or semi-arid conditions where only a small fraction of precipitation reaches the groundwater because the relative errors in the measurement of precipitation can exceed both groundwater recharge and discharge. Wittenberg and Sivapalan (1999) thus suggest a holistic or “top-down” approach which is based on the analysis of measured streamflow. Observed total streamflow is separated into quick flow and baseflow by following previous applications of this approach (Chapman, 1997; Chapman and Maxwell, 1996; Fröhlich et al., 1994; Nathan and McMahon, 1990), with the exception that a nonlinear reservoir algorithm is used. The results from the application of this method compare reasonably well to response functions estimated by other authors based on theoretical, bottom-up approaches and lysimeter measurements (Wittenberg and Sivapalan, 1999).

A hydrograph separation technique described by Moore (1992) was applied to Boulder Creek, USA using extensive groundwater elevation and streamflow data to determine the groundwater discharge component during storm conditions (Hannula, *et al.*, 2002). The estimates of groundwater discharge produced by Hannula, *et al.*, (2002) were found to be reasonable based on the facts that the estimates did not exceed the total flow in the stream, the estimates followed both storm and seasonal trends and the parameters entered into the calculations were physically based (Hannula *et al.*, 2002).

Moore (1991) describes a simple method for hydrograph analysis that is based on relationships of storage depletion to aquifer properties and flow rates during water-level and streamflow recessions. The method was developed to be used in fractured-rock environments. The method was applied in the headwaters of the Melton Branch basin, USA where traditional methods assuming a constant transmissivity did not produce reasonable estimates of groundwater

baseflow. Analysis of the streamflow hydrograph and water level hydrographs during the non-growing season of the area indicates that storm runoff constitutes most of the stream flow after the end of overland runoff, but that discharge from groundwater dominates streamflow again after eight days of recession (Moore, 1991).

There are however contrasting opinions regarding hydrograph separation techniques. Halford and Mayer (2000) argue, from an analysis of 13 sites in the USA, that these methods can be unreliable if used alone, while Arnold and Allen (1999) claim to have had good results for applications on six USA streams, where a correlation between the separation technique estimates with catchment mass balance estimates were found.

3.1.3. Environmental Tracer Methods

Environmental tracer methods have been used to quantify the groundwater discharge to rivers for the past few decades as they offer advantages over physically-based methods, in that they can potentially provide more accurate information on the spatial distribution of groundwater inflows with less costly resources. Cook, *et al.* (2003) make use of ^{222}Rn , CFC-11, CFC-12, major ions and temperature measurements of river water and springs to quantify rates of groundwater discharge to a tropical lowland river in Northern Australia. The method makes use of a numerical model which simulates concentrations of a number of different tracers allowing most parameters to be constrained. The method was found to produce more accurate estimates of groundwater inflow to the river than the simple mass balance method conventionally used. The method concludes that CFC-11 and CFC-12 are suitable to infer rates of groundwater inflow to streams, where ^{222}Rn and major ion tracers are traditionally used.

3.1.4. Isotopes

In the report *Progress in isotope tracer hydrology in Canada*, Gibson, *et al.* (2005) argue that Canadian researchers have played an important role in the development and refinement of isotope hydrology techniques. Fritz, *et al.* (1976) defines the pre-event and event water components of watershed runoff in one of the earliest applications of stable isotopes, with multiple subsequent applications in various physiographic regions of Canada. Cey, *et al.* (1998) quantify the groundwater discharge to a small perennial stream in southern Ontario by performing chemograph separations using $\delta^{18}\text{O}$ and electrical conductivity on two large rainfall events with different antecedent moisture conditions in the catchment. Both events indicated that pre-event water was dominated by groundwater, with a 64-80% contribution towards discharge added by pre-event water. The study also investigates three other techniques to

estimate the contribution of groundwater to the stream, namely streamflow measurements using the velocity-area technique, mini-piezometers measuring hydrometric measurements and seepage meters directly measuring the water flux into or out of the stream. Cey, *et al.* (1998) conclude that large-scale measurements provided a better estimate of groundwater discharge than point-scale measurements, due to the heterogeneous nature of the site. Techniques which can incorporate spatial averaging on a relatively small scale are recommended for proposed new approaches.

3.1.5. Site Specific Scale

The Natural Science and Engineering Research Council of Canada funded a study to facilitate an improvement in the understanding of the surface water – groundwater interactions taking place between a fractured-rock aquifer and a bedrock stream. Oxtobee and Novakowski (2002) made use of air-photo interpretation, electrical conductivity, temperature and isotopic surveys, mixing calculations and point measurements from mini-piezometers, seepage meters and weirs to identify and quantify the interaction between the creek and local aquifer. Groundwater and surface water could easily be distinguished within the study area on the basis of differences in electrical conductivity, temperature and isotopic signatures. Oxtobee and Novakowski (2002) conclude that groundwater discharge in fractured bedrock stream environments mainly occur as discrete point sources related to open fractures which differs from the diffuse, continuous seepage observed in alluvial aquifer environments. Techniques which conventionally are applied to studies in porous media, namely electrical conductivity, temperature and hydraulic head surveys, were found to produce reasonable estimates of groundwater discharge to a stream in a fractured bedrock situation.

In order to better characterise the hyporheic zone, the measurement of groundwater flow on a small scale is vital. High-resolution methods for the estimation of surface water – groundwater interactions are described and tested in a report presented by the Environment Agency. Borehole-based, buried flow meters, direct measurement of the flux at the surface water—groundwater interface, geophysical and thermal techniques are investigated. The report concludes that none of the devices are ideal for all situations and thus a combination of the methods would provide the best results (Environment Agency, 2005b).

The Lambourn River in the United Kingdom is used as a case study for a detailed surface water – groundwater interaction investigation. Allen *et al.* (2010) states a variety of techniques are available to identify and quantify surface water – groundwater interaction processes at a site

scale, i.e. hydrochemistry (Tetzlaff and Soulsby 2008; Mencio and Mas-Pla 2008), fluorescence properties of organic matter (Lapworth et al. 2009), physical parameters (Keery et al. 2007; Schmidt et al. 2007; McGlynn et al. 1999) and process-oriented modelling approaches (Krause et al. 2007). However, each method has its own advantages and limitations which complicate the selection of only one particular method for a specific-site investigation. The conjunctive use of more than one method would increase the overall confidence and understanding of the complex hydrological processes taking place at this scale. An extensive network of boreholes, piezometers and water quality sampling sites were utilized in order to apply a combination of geological, hydraulic and hydrochemical approaches to investigating the surface water – groundwater interactions. These multiple methods have facilitated the development of a comprehensive conceptual model of the study area which according to the authors is clear in certain respects but more ambiguous in others (Allen *et al.*, 2010). This ambiguity in spite of extensive data, illustrates some of the problems faced when considering surface water – groundwater interactions. The study has shown that even a seemingly simple surface water—groundwater system can be hydrologically complex at a local scale. Due to chemically similar groundwater in different components of the system and the heterogeneity of the alluvial aquifer, the hydraulic relationship between the river, the alluvial aquifer and underlying aquifer are still only partially understood in spite of the extensive available physical and geological data. Allen *et al.* (2010) mention recent studies which have emphasized the complexity of surface water—groundwater exchange processes (Krause et al., 2007; Grapes et al., 2005; Griffiths et al., 2006). The realisation of this complexity has implications in how these exchanges are investigated and managed. Methodologies need to be developed which can encompass detailed local scale knowledge into decisions applied at the larger catchment scale and monitoring and sampling extents would need to be carefully considered to ensure an appropriate density.

Rosenberry, LaBaugh and Hunt (2008) describe three of the more commonly used methods applied at the local scale for the investigation of surface water – groundwater interaction, as part of a project funded by the U.S. Department of the Interior and the U.S. Geological Survey. The methods include water-level measurement and flow-net analysis, hydraulic potentiometer (mini-piezometer) and seepage meter methods. The water-level measurement and flow-net analysis method involves the measurement of water levels in a network of wells in combination with measurement of the river stage to calculate gradients and then the flux. The Hydraulic Potentiometer method makes use of multiple mini-piezometers to measure gradients. The Seepage Meters method makes use of seepage meters

to directly measure flow across the sediment-water interface at the bottom of the surface-water body. Rosenberry *et al.* (2008) conclude that all three of the methods have different advantages and disadvantages, making the selection of a method for a local study area dependent on the characteristics of that specific site (Rosenberry and LaBaugh, 2008).

3.1.6. Analytical Methods

Craig and Read (2010) state it is generally understood that any exact solution to a differential equation that can be expressed in terms of polynomial, logarithmic, exponential, and/or trigonometric functions is an analytical solution. The most basic analytical solution for determining the groundwater contribution to streamflow is based on Darcy's Law, where the flux is a function of the difference between the river stage and the aquifer head which can be expressed as:

$$q = k\Delta h \quad (3.1)$$

and,

$$\Delta = h_a - h_r$$

where,

h_a is the aquifer head,

h_r is the river head,

q is the flow between the river and the aquifer (positive for gaining streams and negative for losing streams), and

k is a constant representing the streambed leakage coefficient or a conductance term.

In Equation 3.1 the flux (q) per unit area is directly proportional to the head gradient between the surface water and groundwater, forming a linear function. Figure 3-1a graphically represents the linear relationship between the flux and change in head. Figure 3-1b indicates the effect when the influent flow occurs at a slower rate than the effluent rate of flow and Figure 3-1c indicates the reverse situation of a faster influent rate. A non-linear version of the Darcy principal was proposed by Rushton and Tomlinson (1979), in Sophocleous (2002) which compensates for the effects of streambed resistance by considering upper limits for fluxes (Figure 3-1d). Rushton and Tomlinson (1979) also proposed an equation using both linear and exponential functions for non-linear cases without an upper limit (Figure 3-1e).

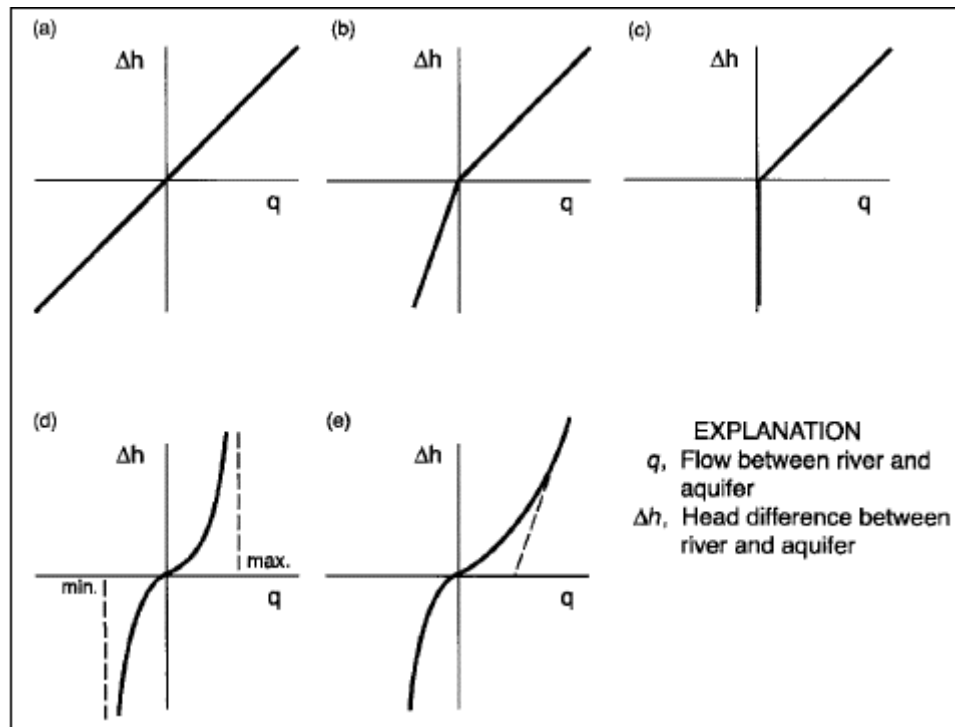


Figure 3-1 The relationship between the stream and aquifer as a function of the various head-difference scenarios where a), b) and c) depict linear relationships while d) and e) depict non-linear relationships (Taken from Sophocleous, 2002).

A number of analytical solutions exist for the description of the influence of groundwater abstraction on surface water – groundwater interactions. Analytical solutions include work by Theis (1941); Glover and Balmer (1954); Hantush (1965); Jenkins (1968); Wallace, *et al.* (1990); Grigoryev (1957), cited in Butler, Zlotnik and Tsou (2001); Bochever (1966), cited in Butler *et al.* (2001); Glover (1974); Stang (1980); Hunt (1999); Wilson (1993); Zlotnik *et al.*, (1999); Damara (2001); Bakker and Anderson (2003); Chen and Yin (2004); and Di Matteo and Dragoni (2005). The methodology, assumptions and limitations of these analytical solutions are covered in detail in Witthüser (2006), Dennis and Witthüser (2007) and Moseki (2013).

Witthüser (2006) created a comparison table for the above analytical solutions based on assumptions, limitations and complexity of each solution. The Di Matteo and Dragoni (2005) analytical solution was found to be the least suitable method and not recommended for use in abstraction licensing. Based on South African processes, Witthüser (2006) recommended the Chen and Yin (2004); Stang (1980) and Butler *et al.* 2001) solutions for use and further investigation.

3.1.7. Integrated Surface water—Groundwater Models

A review of surface water – groundwater interaction modelling in arid/semi-arid floodplains was performed as part of the Australian Hydrological Modelling Initiative, Groundwater – Surface water Interaction Tool Project. Jolly and Rassam (2009) state significant advances have been made over the last fifteen years in the modelling of surface water – groundwater interactions, progressing from relatively simple 1D and 2D analytical and empirical approaches to highly refined, 3D spatially-distributed integrated models. However, these large progressive steps in modelling power have been accompanied with the scaling down of routine hydrological and hydrogeological monitoring networks (Silberstein, 2003). The advances within the modelling discipline will not be limited by computing power or solution methods, but rather by the availability of suitable data to parameterize, calibrate and validate the models. It is thus seen that simple analytical options are still a useful option in data scarce situations. Jolly and Rassam conclude that the use of sophisticated numerical models should be limited to data-rich situations, where calibration and validation may be performed (Jolly and Rassam, 2009).

The currently available fully integrated surface water – groundwater flow and transport models are a reflection of the increasing complexity of the existing hydrological models. Partington, *et al.* (2011) state accurate quantification of streamflow generation mechanisms are still not possible within these advanced models, in that the groundwater component of baseflow at a particular point along the stream cannot be specified. Partington, *et al.* (2011) developed a Hydraulic Mixing Cell (HMC) method, as part of the Linkage Scheme supported by the Australian Research Council. The HMC model only uses the flow solution from fully integrated surface water – groundwater flow models to determine the groundwater component of streamflow. As the model only requires hydraulic information the need for the simulation of tracer transport is eliminated, which could be advantageous if tracer concentration data was not available. The trend seen in methods applied for quantifying the groundwater component of streamflow is integration into groundwater flow numerical models. These methods as discussed by Harington *et al.* (1988) and Partington *et al.* (2010) are adequate methods for determining the groundwater component of streamflow, but the data required to set up the models are often not widely available, especially in a South African context.

3.1.8. SWAT

The Soil and Water Assessment Tool (SWAT) is a conceptual, continuous time model developed to assess water supplies and non-point pollution sources on watersheds and large river basins using readily available data (Arnold, *et al.*, 1998). Daily precipitation is the main input to the model and the groundwater flow is computed in the model by creating a shallow aquifer storage. The return flow from the shallow aquifer is then calculated using an empirical equation described by Arnold *et al.* (1993) and a relationship for the water table height in response to recharge. The model was however originally developed to predict the impacts of agriculture management on erosion and sedimentation rates. The model has sub-basin components which are categorised into hydrology, weather, sedimentation, soil temperature, crop growth, nutrients, pesticides and agricultural management (Arnold *et al.*, 1998). Arnold, *et al.*, (1999) conducted a comparison study of the estimates of regional recharge and discharge produced by the SWAT model and a combination of other hydrograph techniques, namely a digital recursive filter used to separate baseflow from total streamflow and a modified hydrograph recession curve displacement method to estimate recharge to the shallow groundwater system. The baseflow estimates from both methods were compared to measured baseflow for three watersheds in Illinois and results from another separation technique. The comparison between the models and to the measured baseflow values indicated that both methods followed the same regional trends (Arnold *et al.*, 1999). However, the model seems to estimate total baseflow to the river instead of the groundwater component of baseflow.

3.1.9. GSFLOW

The U.S. Geological Survey's Groundwater and Surface water Flow model (GSFLOW) is an integrated hydrological model developed to simulate coupled groundwater and surface water resources. The model is based on the integration of the USGS Precipitation-Runoff Modelling System (PRMS) and the USGS groundwater flow model MODFLOW. The coupled approach towards integrated hydrologic modelling used in GSFLOW, partitions the surface and subsurface systems into separate regions and the governing equations which describe flow in each of these regions are then integrated or coupled using iterative solution methods (Markstrom, *et al.*, 2008). GSFLOW provides a robust modelling system for simulating flow through the entire hydrological cycle and can be used to evaluate the effects of land-use changes, climate variability, groundwater abstraction on surface and subsurface flow and many more. A numerical algorithm is made use of to simulate the most important processes affecting surface water and groundwater flow systems. The interaction between surface water and groundwater

can be simulated in catchments ranging from a few square kilometres to several thousand square kilometres and allow for simulation periods that range from months to decades. However, the model has a large number of inherent assumptions. There are assumptions and limitations associated with each of the modules and packages contained within GSFLOW. GSFLOW is a non-linear model and is thus limited by the possibility of non-convergence among any or all coupled dependent variables, or due to inappropriate input data or parameters. The GSFLOW model has a number of limitations in terms of the discretization of time and space. One limitation is that the model has a computational time step of one day which requires all flow and storage data in a mean daily format. The small time step might also lead to errors as flow near the land surface tends to occur faster than flows in the subsurface. The size of finite difference cells are constrained by the relative width of cells compared to the width of the river. A large cell relative to a stream that flows over a cell will result in model errors and misrepresentation of surface water – groundwater interaction, or where a cell width is equal to or less than the width of a river the model may not converge (Markstrom *et al.*, 2008). There are also additional assumptions associated with the canopy zone, land-surface precipitation and temperature, the soil zone, streams, lakes, groundwater and unsaturated zone functions (Markstrom *et al.*, 2008).

3.1.10. MODFLOW

The U.S. Geological Survey's three-dimensional, groundwater flow model (MODFLOW) was released as a versatile simulator of groundwater flow within an aquifer almost thirty years ago and the programme is still in widespread use (Swain, 1994). MODFLOW solves the three – dimensional groundwater flow equation by means of finite difference approximations (McDonald and Harbaugh, 1988). The aquifer is divided into cells which have dimensions x , y and z and the aquifer properties within are assumed uniform as this is required for the finite difference equations. The head is assigned and calculated at the centre of each cell by iterating the finite difference equations for all nodes until the maximum head change in any cell within the previous and current iteration is less than the user-specified value and then the process is repeated. However, the appropriate time step for this iteration will vary for surface water and groundwater because the response time of surface water systems is usually faster than groundwater systems.

The modular design of the MODFLOW model allows for the addition of new packages to both expand the capacity of the model and improve the accuracy. A number of additional packages

have been developed, ranging from the simulation of the effect of artificial recharge to the interaction between surface water and groundwater.

The RIVER package and the STREAM package

The RIVER and STREAM packages allow the user 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 within the programme MODFLOW (Pattle Delamore Partners Ltd. (2000), cited by Moseki, 2013).

The flow between the stream and the aquifer (Q_{riv}) is calculated as (Moseki, 2013):

$$Q_{riv} = C_{riv}(H_{riv} - h) \quad \text{for } h > R_{bot} \quad (3.2)$$

where,

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

The direction of flow to and from the river varies depending on the head defined in the aquifer (Moseki, 2013):

$$Q_{riv} = C_{riv}(H_{riv} - R_{bot}) \quad \text{for } h \leq R_{bot} \quad (3.3)$$

and, the hydraulic conductance can be calculated using Darcy's law as (Moseki, 2013):

$$C_{riv} = \frac{k' L_c w}{b'} \quad (3.4)$$

where,

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

Moseki (2013) highlights that most of the parameters required in Equation 3.4 are often not known which results in C_{riv} being adjusted during the model calibration, which is not the optimum process to follow. The MODFLOW RIVER package is further limited in terms of modelling vertical seepage through the unsaturated zone as the package makes use of Darcy's equation and not the Richard equation for unsaturated flow (Dennis and Witthueser (2007), in Moseki (2013).

MODBRANCH

The BRANCH package, referred to as MODBRANCH when incorporated into MODFLOW, simulates unsteady, non-uniform flow in open channels using an implicit, weighted four-point finite difference approximation for the dynamic wave equations. The advantage of MODBRANCH is that the model allows for the simulation of unsteady flow in a network (dendritic or looped) of single open-channel reaches or branches (Moseki, 2013).

Flow within the River is solved using the one-dimensional continuity equation, which is a steady-state water balance assuming the inflow rate equals the outflow rate with no change in storage (Dennis and Witthuiser (2007), in Moseki, 2013). Surface water – groundwater interactions are modeled as one-dimensional vertical leakage through a clogging layer in the streambed. The model's additional data requirements include the channel geometry, initial flow conditions at all cross-sections and boundary conditions at channel edges (Moseki, 2013).

Swain (1994) describes three add-on packages designed to simulate the interaction of surface water – groundwater, namely the channel stage River package, the flow-routing Stream package and the unsteady open-channel flow model BRANCH using the MODBRANCH coupling programme. The River package assumes a constant river stage and computes the surface water – groundwater interaction as leakage across a confining riverbed, but the flow in the river is not simulated and thus acts as an infinite source or sink. A programme developed by Swain (1994) allows direct-flow connections to be simulated between MODFLOW and the three surface water – groundwater interaction packages. The programme facilitates the modelling of different sections of a river using the various interaction packages simultaneously within MODFLOW.

DAFLOW-MODFLOW

The Diffusion Analogy Surface Water Flow model (DAFLOW), employing a one dimensional diffusive wave approximation for in-channel flow, allows for flow routing and contains an iterative time-stepping approach for coupling the surface and subsurface interactions (Moseki, 2013). The model allows for the surface water – groundwater interaction to be quantified while considering streambed resistance.

The leakage (Q_i) from a section of river into a single, specified aquifer cell is calculated for each time step (Moseki, 2013):

$$Q_i = \frac{k'lw(H_a - Y - B_e)}{b'} \quad (3.5)$$

where,

- k' is the hydraulic conductivity of the streambed,
- l is length of the sub-reach in hydraulic connection with the aquifer cell,
- w is the average width of the stream in the cell,
- H_a is the head of the aquifer in the cell,
- Y is the average depth of the stream in the sub-reach,
- B_e is the average elevation of the streambed, and
- b' is the thickness of the streambed.

Jobson and Harbaugh (1999) couple the surface water flow model DAFLOW to the modular groundwater flow model MODFLOW to improve the models ability to simulate surface water – groundwater interaction by allowing two different scales of time-steps to be used. DAFLOW has been structured to include subroutines which allow multiple time steps to be run iteratively within one MODFLOW time step. The subroutines within DAFLOW can compensate for the difference in response times between surface water and groundwater. Jobson and Harbaugh (1999) mention, but do not compare DAFLOW to the other available surface water – groundwater interaction packages available for MODFLOW. Jobson and Harbaugh (1999) conclude that DAFLOW provides a highly stable solution scheme which is easy to run and requires a minimum of field data and calibration. Jobson and Harbaugh (1999) recommend the use of the model for upland streams.

However, Dennis and Witthüser (2007) found the accuracy of this model varies depending on the slope of the modelled area, with the accuracy increasing with an increase in slope. The model is also sensitive to the defined temporal discretization in relation to the streambed slope (Dennis and Witthüser, 2007).

Stream-Routing (SFR1) package

The increased awareness of the interconnection between surface water and groundwater lead to the development of a new Stream-Routing (SFR1) package for the groundwater flow model, MODFLOW (Prudic, *et al.*, 2004). The SFR1 package replaces the older Stream (STR1) package written for earlier versions of MODFLOW. The new SFR1 package has several improvements from the previous STR1 package with the main difference between the two packages being that the river depth is computed at the midpoint of the river reach in SFR1 rather than at the beginning of the reach as in STR1. The main limitation of the new SFR1 package is that it is not

suiting for modelling the transient exchange of water between the river water and groundwater on a short-term time scale of days or minutes. This is due to the assumption that streamflow and associated dissolved solutes are routed between the stream reaches only on the basis of continuity. The package also makes the assumption that solutes are completely mixed within each stream reach which limits the applicability for large rivers and simulating short-term effects on surface water – groundwater interaction (Prudic *et al.*, 2004).

The SFR1 approach uses one channel cross-section for a stream segment, divided into three parts based on eight paired horizontal and vertical locations (Figure 3-2). Eight horizontal distances, relative to the left edge of the cross section, and the corresponding eight vertical altitudes, relative to the specified top of streambed, are used for computing stream depth, top width, and the wetted perimeter (Markstrom *et al.* (2008), in Moseki, 2013).

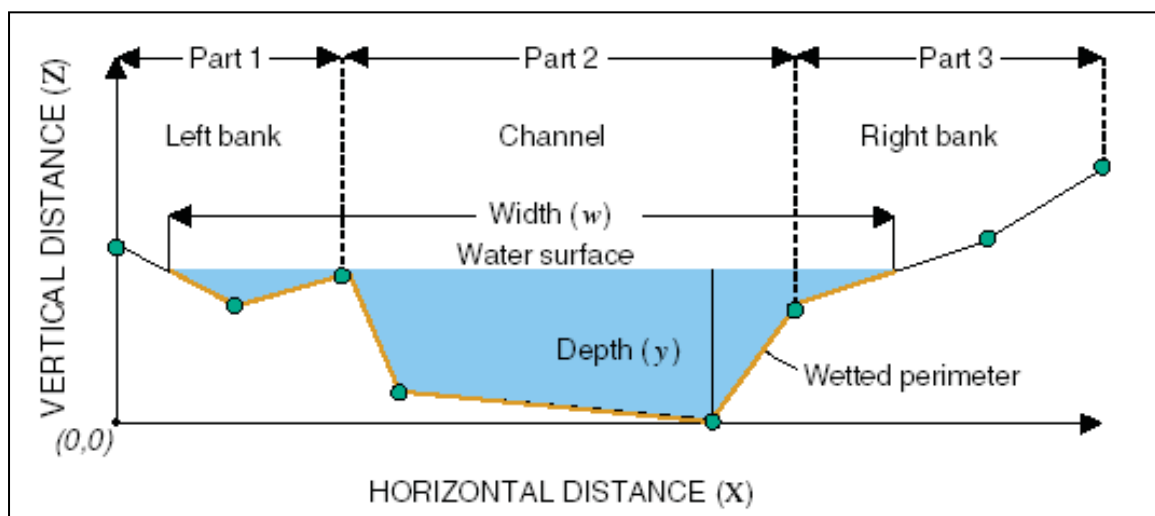


Figure 3-2 An 8-point cross section for calculation of depth, width and wetted perimeter for a stream segment (Taken from Markstrom *et al.*, (2008).

Surface water – groundwater interaction simulated as seepage by the SFR1 package is limited by streambed conductance and the head difference between the stream and aquifer. The package is able to account for seepage when the stream is hydraulically separated from the water table by simulating seepage based on the head difference between the stream and the bottom of the streambed (Niswonger and Prudic (2005), cited in Moseki, 2013).

Stream-Routing (SFR2) package

Hydraulically disconnected streams found in semi-arid regions are becoming increasingly prevalent as groundwater abstractions lower groundwater levels in valley aquifers beneath rivers (Niswonger and Prudic, 2005). In order to investigate the connection of surface water and

groundwater through an unsaturated zone another streamflow-routing package (SFR2) for MODFLOW was developed which has the capability to simulate unsaturated flow beneath rivers. The capability of the SFR2 package to model unsaturated flow is seen as an improvement towards modeling perched rivers more realistically. A kinematic wave approximation is used to simulate unsaturated flow beneath rivers, that assumes that the downward flow beneath the river is purely due to the force of gravity (Niswonger and Prudic, 2005). The SFR2 package still has all of the SFR1 package capabilities, a simulation for a time delay in recharge and maintains the applicability of MODFLOW-2000 to catchment-scale situations. The package has a number of associated assumptions which can limit the applicability of the model, but in two test simulations the magnitude and downward progression of the wetting front were in agreement with results from the U.S. Geological Survey's Variably Saturated Two-Dimensional Flow Transport (VS2DT) model (Niswonger and Prudic, 2005).

Vertical seepage through a homogeneous unsaturated zone is approximated by means of kinematic waves and a simplified Richards' equation (Moseki, 2013):

$$\frac{\partial \theta}{\partial t} = \frac{\partial q}{\partial z} = \frac{\partial}{\partial z} \left[D(\theta) \frac{\partial \theta}{\partial z} - K(\theta) \right] \quad (3.6)$$

where,

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

The gravity-dominated flux in the unsaturated zone is equal to the unsaturated vertical hydraulic conductivity, limited to a maximum of the saturated hydraulic conductivity. Seepage across the streambed is thus restricted by the underlying vertical hydraulic conductivity in the unsaturated zone. SFR2 data requirements include the infiltration rate and wetted area of the stream. The volume of water that seeps from a stream is calculated by multiplying the infiltration rate and the wetted area of the stream. The relation between the river stage and volume of water discharged is calculated using Manning's equation. Figure 3-3 indicates how the unsaturated zone under a stream is represented within a single MODFLOW cell. The unsaturated zone is discretized into several compartments defined by eight points in the example given by Niswonger and Prudic (2005) (Figure 3-3).

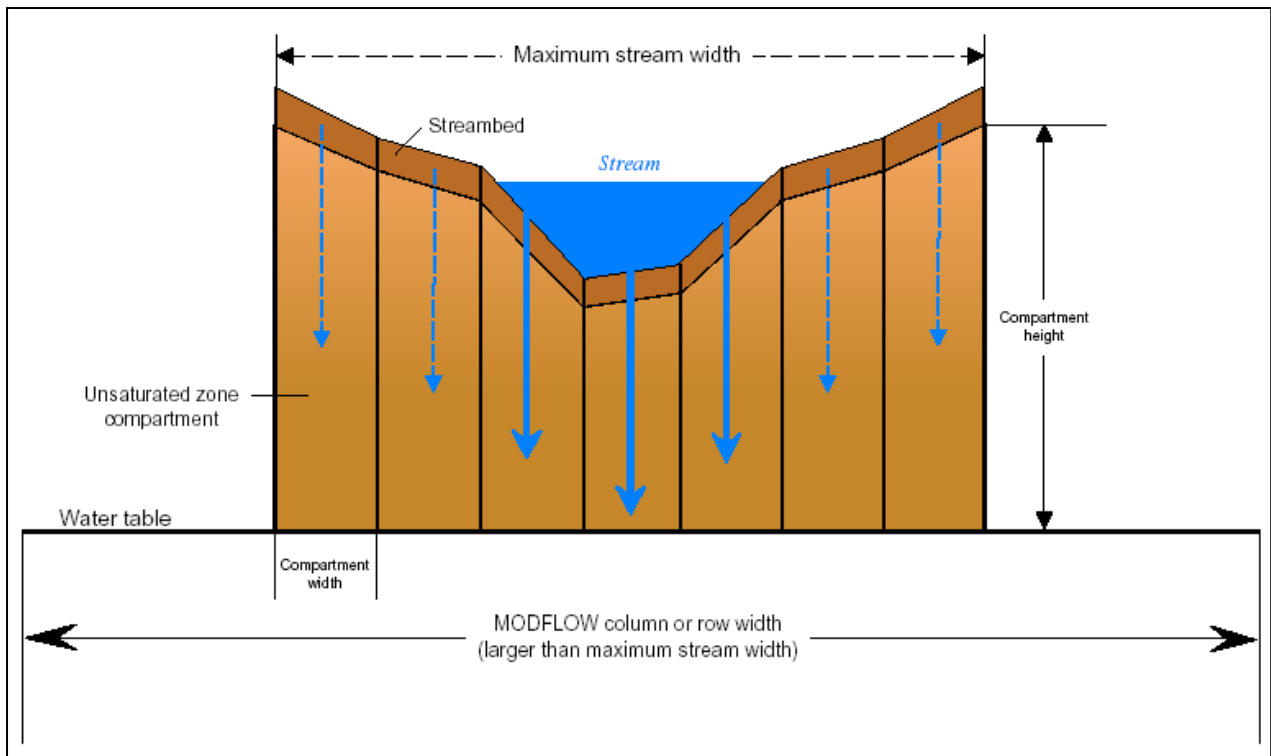


Figure 3-3 The discretization of the unsaturated zone under a stream, of a variable cross-section, within a single MODFLOW cell. Dashed lines represent the maximum stream depth and the associated infiltration at the maximum depth (Taken from Niswonger and Prudic (2005), in Moseki, 2013).

3.1.11. MIKE SHE and SHETRAN

MIKE SHE is an integrated hydrological modelling system for simulating surface water and groundwater flow, while SHETRAN provides three-dimensional coupling of water flow and contaminant transport (Moseki, 2013). The model is based on the assumption of an unconfined aquifer underlain by an aquiclude. The catchment is discretized by a horizontal orthogonal grid and vertical columns of horizontal layers representing the various hydrological compartments (Moseki, 2013).

Finite-difference solutions of mass, energy and momentum, as well as empirical solutions are used to model hydrological processes in MIKE SHE and SHETRAN. The surface water – groundwater interaction is modeled as a function of the head difference between the river and the aquifer, taking into account clogging layers in the streambed and disconnected streams (Moseki, 2013).

The SHE and SHETRAN model have been developed from a hydrological point of view to model hydrological processes which has resulted in an over-simplistic approach to surface water – groundwater interaction. The aquifer is defined as an unconfined aquifer represented as a single layer, which results in the model not being able to model three-dimensional groundwater flow (Witthüser, 2006). Witthüser (2006) does however recommend the use of this model to

simulate water quality issues because the model allows three-dimensional coupling of surface water flows to the single layer aquifer and includes sediment, contamination transport and surface water – groundwater interaction modelling.

3.1.12. FEFLOW and MIKE 11

The Finite Element Subsurface Flow and Transport Simulation System (FEFLOW) is a numerical model using finite element differential equations to model fluid flow and transport of dissolved constituents and/or heat transport processes in the subsurface. The model covers a broad range of functionality for porous-media flow and transport simulation and is accessible via a comprehensive user-interface (DHI-WASY, 2012). FEFLOW is coupled via the IFM-Tool (FEFLOW Interface Manager) with the MIKE11 surface water software to model surface water – groundwater interaction.

The MIKE11 software simulates 1D unsteady flow in surface water networks and quasi-2D flow on floodplains with a finite-difference scheme. Structures in the river such as weirs, bridges, pumps and user-defined structures can be incorporated into MIKE11. FEFLOW and MIKE11 are not coupled in an iterative manner. This non-iterative coupling is achieved by exporting discharges calculated by FEFLOW at coupled boundary nodes (3rd type or Cauchy boundary conditions) to MIKE11 H-points (calculating points of a MIKE11 network), as an additional baseflow boundary condition. The MIKE11 programme will calculate its time step as often as needed to reach the actual time level of FEFLOW. Once MIKE11 has calculated its time steps to coincide with the FEFLOW time steps, the water levels of the H-points calculated in MIKE11 are exported to the FEFLOW coupling boundary nodes, and FEFLOW calculates its next time step (Moseki, 2013).

The FEFLOW programme with the incorporated MIKE11 software was the most comprehensive modeling approach to surface water – groundwater interaction at the time of the study performed by Dennis and Witthüser (2007). Dennis and Witthüser (2007) however, continue to state the data and skill requirements to ensure the use of the model at its full capacity are great and thus might limit the use of the model.

3.2. South African Approaches and Methods

3.2.1. Guidelines

The Department of Water Affairs and Forestry developed the Guidelines for Groundwater Resources Management in Water Management Areas of South Africa in 2004. The Guidelines were aimed at the integration of coordinated groundwater management into Integrated Water Resources Management (IWRM) at different levels within Catchment Management Agencies (CMA). The guideline contains a step by step plan for groundwater resource assessments in South Africa, including initial/conceptual planning, water balance calculation, strategic environmental assessment, characterisation of the aquifer and detailed planning and reconnaissance. The guideline elaborates on the Water Balance Calculation in that a conceptual model of groundwater and surface water resources should be developed on which a water balance equation is based. It is important to gain a good understanding of the groundwater discharge from the system in order to ensure a balance between inflows and outflows. Groundwater discharge may present itself in the following forms: abstraction from boreholes, baseflow to rivers, baseflow to springs, baseflow to wetlands, discharge to the sea, transpiration from vegetation and evaporation from shallow groundwater. Methods which are available to directly measure or infer groundwater discharge volumes to rivers are summarised in the report. The use of streamflow hydrographs can be used to separate streamflow into its components based on the assumption that these different components will appear at different time intervals. The graphical method of hydrograph separation separates quick flow from slow flow purely based on graphical properties, but this method tends to inadequately describe stream chemistry during storm runoff events. Another method available is the isotope/chemical hydrograph separation where stable environmental isotopes are used in conjunction with the stream hydrograph to estimate the groundwater baseflow. The report continues that the hydrograph separation technique used will depend on the available data. The guideline suggests Bokuniewicz and Zeithin's (1980) method of directly measuring the groundwater inflow to surface water by means of a simple drum and plastic bag and Paulsen *et al.* (1997) ultrasonic groundwater seepage meter method which takes continuous measurements of groundwater seepage into the surface water body, as small scale methods for estimating the groundwater – surface water exchange. The report also documents best practices and step by step methodologies which can be used to implement the protection strategies as stated in the National Water Act 1998. The National Water Act 1998 gave rise to Resource Directed Measures

which includes resource classification, determination of resource management classes, reserve determination and setting of resource quality objectives. The guideline describes a comprehensive step by step implementation of the Resource Directed Measures. Furthermore, suggested methods of determining the groundwater-fed baseflow are given under the quantification of the Reserve. The guideline suggests the Smatkin or Herold hydrograph separation methods, Darcy's law, chemical investigations or numerical modelling for the quantification of groundwater –surface water interaction. However, the DWAF policy with regards to this step in the Resource Directed Measures is still in the formulation stage (DWAF, 2004).

Moseki (2013) states that there is no “one fits all” method to assess surface water – groundwater interactions and a number of factors along with the conceptual understanding of the area and the applicability of the data, play an important role. Moseki (2013) recommends the use of a framework to choose a particular method for a surface water—groundwater investigation at a particular location. A framework created by Moseki (2013) to inform decision processes in the selection of appropriate methodologies for the evaluation of surface water – groundwater interaction in the context of South Africa's environmental conditions is shown in Figure 3-4. Gaining a conceptual understanding of the area under investigation is recommended by Moseki (2013) as the first step in the process of evaluating the surface water – groundwater interaction, followed by whether exchange is possible, the degree of interconnectivity and subsequently the type of interaction occurring. Once this has been determined then an appropriate method of quantification can be selected. Moseki (2013) strongly recommends the use of more than one method to increase the confidence levels in the results and decrease the associated uncertainties.

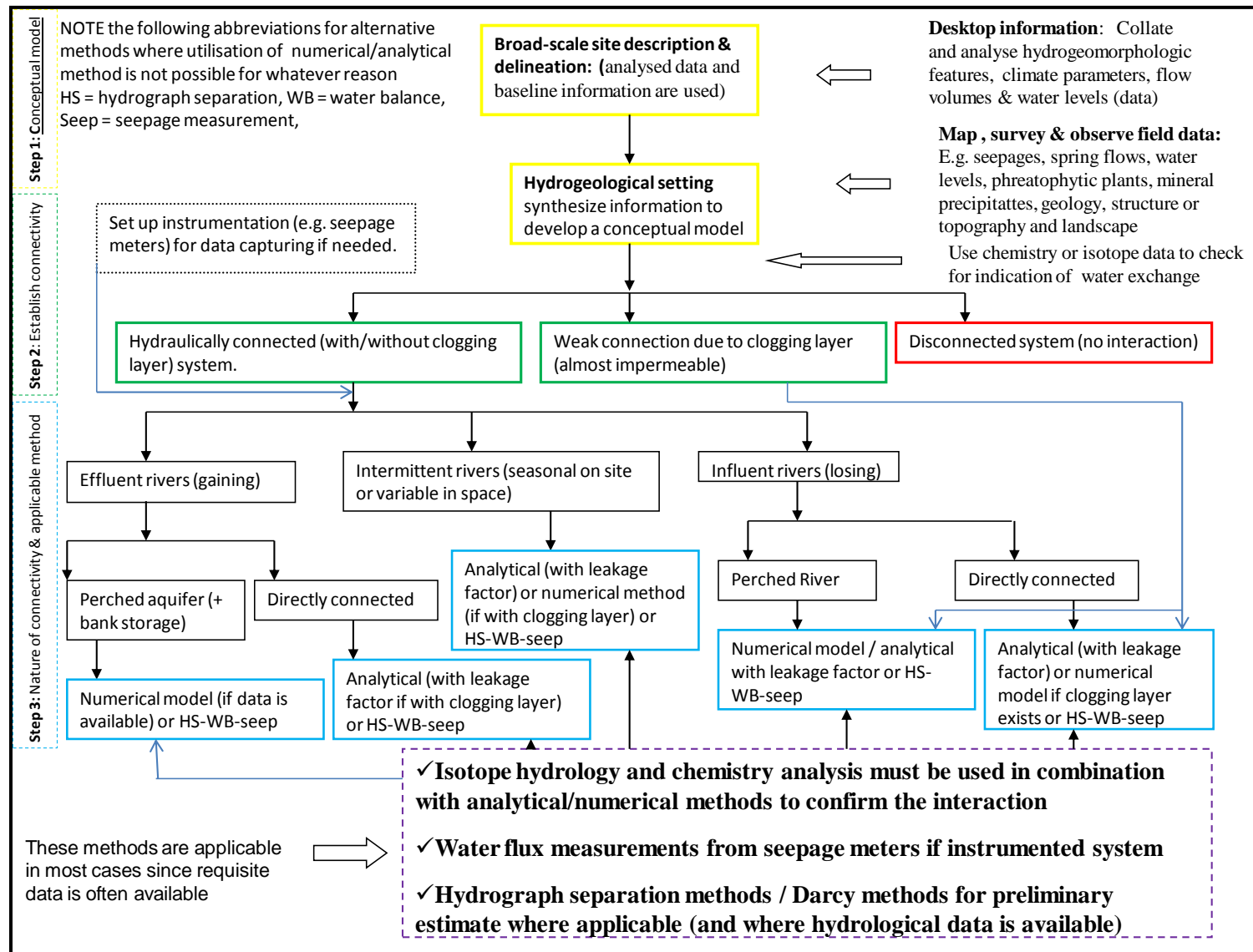


Figure 3-4 A framework for the selection of an appropriate method for the assessment of surface water-groundwater interactions (Taken from Moseki, 2013).

3.2.2. Hydrograph Separation techniques

Hydrogeomorphological approach

Xu, *et al.* (2002) presented a geomorphologic framework on which the quantification of groundwater baseflow from a streamflow hydrograph can be discussed. The geomorphologic framework was developed in order to supplement the hydrograph separation techniques used in South Africa to quantify groundwater – surface water interaction. Xu *et al.* (2002) list the various methods of hydrograph separation utilised in South Africa, namely the RCD, Concentration ratio, Herold, SARES and Smakhtin methods. According to Xu *et al.* (2002) the RCD method is not frequently applied, the Concentration method is favoured for interflow investigations, the SARES method is favoured for ecological reserve investigations and the Herold and Smakhtin methods are the most acceptable and popular methods for the quantification of groundwater – surface water exchange. Halford and Mayer (2000) however found that hydrograph separation techniques are insufficient tools when used unaccompanied by additional methods to determine the interaction between groundwater and surface water. Hydrogeological investigations are traditionally qualitative and aimed at understanding the groundwater flow occurrences, where numerical solutions have been favoured for quantitative investigations (Xu *et al.*, 2002). Numerical simulation techniques tend to be costly and require additional calibration data. Xu *et al.* (2002) proposes an alternative approach where geomorphic characteristics of rivers are used to create hydrogeologic rules aimed to increase the consistency of the separation of streamflow by hydrograph techniques. Rivers are geomorphologically classified into upper catchment areas, middle river courses and lower river courses. The rivers are then further classified on the hydrogeomorphological type, namely constantly losing or gaining streams, intermittent streams, gaining streams with or without storage and interflow-dominant streams. Four different relationships between rivers and groundwater are defined based on geomorphologic typing, interaction scenarios, hydraulic connection and baseflow separation. Xu *et al.* (2002) propose an algorithm for the estimation of the monthly groundwater discharge which incorporates qualitative knowledge. The algorithm estimates the groundwater contribution to baseflow through a summation of the decay of the previous groundwater contribution and a rainfall-induced flow increment, where each different relationship between groundwater and surface water will result in different parameter values. The proposed approach was applied to the Sabie River, South Africa and Xu *et al.* (2002) report reasonable estimates which are comparable to estimates presented by Vegter (1995). Xu *et al.* (2002) conclude that the proposed approach can add meaning to simple hydrograph separation

techniques, but should be applied with caution as it is based on an hydrogeomorphological understanding and is subjective in nature (Xu *et al.*, 2002).

Recession Curve Displacement (RCD) Method

The RCD (Recession Curve Displacement) method developed by Rorabaugh (1964) is a recession analysis hydrograph separation technique which is based on the upward displacement of the recession curve during a rainfall event. The total recharge to the groundwater system during the rainfall event is shown to be approximately double the total potential discharge to the stream at a critical time (T_c) after the hydrographic peak, by means of an algorithm. The total volume of groundwater recharge due to the rainfall event (R) can be estimated from the stream hydrograph by (Brodie and Hostetlet, 2005):

$$R = \frac{2(Q_2 - Q_1)K}{2.3026} \quad (3.7)$$

where,

Q_1 is the baseflow at critical time (T_c) extrapolated from the pre-event recession curve,
 Q_2 is the baseflow at critical time (T_c) extrapolated from the post-event recession curve, and
 K is the recession index.

The recession index (K) is estimated from the stream hydrograph record, and then used to determine the critical time (T_c) from the relationship $T_c = 0.2144K$. Figure 3-5 graphically describes the various parameters in Equation 3.7. When the recharge to groundwater is known, Equation 3.7 can be rearranged to solve for the groundwater baseflow from the pre-event recession curve.

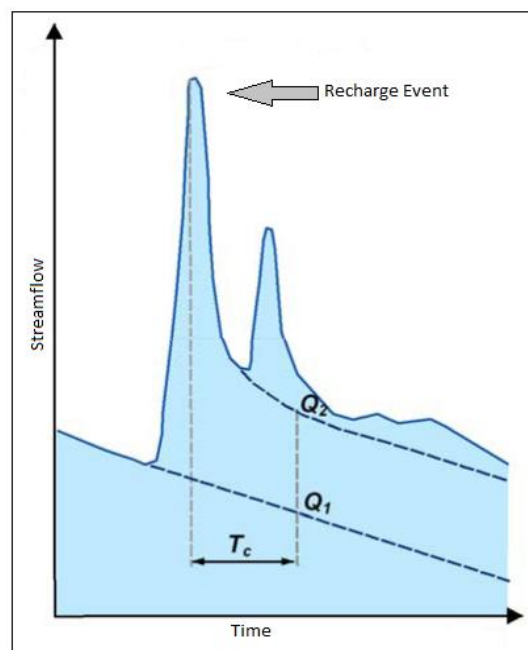


Figure 3-5 Graphical representation of the RCD method parameters T_c , Q_2 and Q_3 for a stream hydrograph recharge event (Modified from Rutledge and Daniel (1994)).

Concentration Ratio Method

The Concentration Ratio method referred to by Xu, *et al.* (2002) is a chemical hydrograph separation method. The method is also referred to as the Tracer method in Australian Government (2012a) and as a chemical method in DWAF (2006c). End-members of streamflow, runoff and groundwater are chemically defined with recommended tracers including chloride (Cl), silica (Si), hydrogen-2 isotope (^2H) and oxygen-18 isotope (^{18}O). Thus, at any point along the river, the proportion of river flow supplied by groundwater discharge is calculated based on a chemical mass balance equation (DWAF, 2006c):

$$Q_g = Q_T \left(\frac{C_r - C_d}{C_g - C_d} \right) \quad (3.8)$$

where,

Q_g is the baseflow volume,

Q_T is the total measured streamflow, and

C_d , C_g and C_r are the concentrations of surface runoff, groundwater and streamflow.

Herold Method

The Herold hydrograph separation method was developed by Herold (1980). The method is used in the Water Resources 1990 project to separate monthly flows into surface and groundwater components. The method is based on the following equation (DWAF, 2006a):

$$Q_g = Q_{gi-1} \cdot Decay + Q_{i-1} \cdot PG \quad (3.9)$$

where,

Q_{gi} is the groundwater contribution of the current month,

Q_{gi-1} is the groundwater contribution of the preceding month,

Q_{i-1} is the total streamflow of the preceding month,

$Decay$ is a groundwater factor ($0 < Decay < 1$), and

PG is a groundwater growth factor (%).

The calculated groundwater baseflow (Q_{gi}) is thus the combined effect of previous groundwater after decay ($Q_{gi-1} \cdot Decay$) and rainfall induced recharge ($Q_{i-1} \cdot PG$). The Herold method has four parameters which it requires for computation, namely $DECAY$, PG , $GGMAX$ and $QGMAX$. $GGMAX$ is varied on a monthly time step in response to $DECAY$ (Groundwater decay factor) and PG (Groundwater growth percentage). $DECAY$, PG and $QGMAX$ are selected on the basis of a realistic division of groundwater and surface water (Levy and Xu, 2011). The calibration of these parameters is facilitated by a graphical output of total and groundwater hydrographs, but the selection remains completely subjective as what is the definition of a “realistic” division.

SARES method

SARES is a computer programme forming part of the Decision Support System (DSS) software developed by Hughes and Münster (1999). The programme was developed as a tool to rapidly access an initial low-confidence estimate of the quantity component of the Reserve for rivers at the outlet of any quaternary catchment. The final result of the programme is a time series of monthly flow volumes recommended for the quantity component of the Ecological Reserve (Hughes and Münster, 1999).

In order to estimate the Ecological Reserve the in-stream flow requirement needs to be determined, which consists of four components in the SARES programme, namely low and high flow maintenance quantities and the high and low flow drought quantities. The low flow maintenance quantity can be assumed to the groundwater contribution to stream flow (Hughes and Münster, 1999).

The maintenance low flow requirements are determined based on the study area's Ecological Management Class (EMC) and the defined CVB index, which is a combination of the variability index and the Baseflow Index (BFI). The equation used to determine the maintenance low flow requirement is (Hughes and Münster, 1999):

$$MLIFR = LP4 + \frac{(LP1 \cdot LP2)}{(CVB^{LP3})^{(1-LP1)}} \quad (3.10)$$

where,

MLIFR is the maintenance low flow total as % natural MAR, and
LP1-4 are parameters associated with the EMC set (Table 3-1).

Table 3-1 Parameters LP1—LP4 for each Ecological Management Class option (Taken from Hughes and Münster, 1999).

Parameter	Ecological Management Class						
	A	A/B	B	B/C	C	C/D	D
LP1	0.9	0.905	0.91	0.915	0.92	0.925	0.93
LP2	79	61	46	37	28	24	20
LP3	6	5.9	5.8	5.6	5.4	5.25	5.1
LP4	8	6	4	2	0	-2.0	-4.0

Smakhtin Method

The Smakhtin method is a recursive digital filter method for monthly baseflow separation developed by Smakhtin (2001). The method is based on daily separation technique developed by Nathan and McMahon (1990). A filter parameter (a) is used to separate quick flow from baseflow. Quick flow is assumed to be made up of interflow and storm runoff, and the remaining flow is assumed to be baseflow. The monthly Smakhtin groundwater baseflow equation is (Smakhtin, 2001):

$$QB_m = Q_m - q_m \quad (3.11)$$

and,
$$q_m = aq_{m-1} + 0.5(1 + a)(Q_m - Q_{m-1}) \quad (3.12)$$

where,

q_m and q_{m-1} are the current and previous monthly flow attributed to high-flow events,
 Q_m and Q_{m-1} are the current and previous months total monthly flow,
 QB_m is the part of the total monthly flow which could be attributed to baseflow, and
 a is the filter parameter.

The determined baseflow is further constrained to ensure that the volume does not become negative or exceed the original total monthly streamflow in any month.

3.2.3. Statistical Analysis

River heterogeneity signatures, derived from a combination of geomorphological province, eco-region and an index of river flow variability (i.e. the Hydrological Index (HI)), is used to prioritise conservation of South African rivers (Le Maitre and Colvin, 2008). The Hydrological Index is a general index of flow variability calculated from two standard flow statistics, namely the Coefficient of Variation Index (CVI) and the Baseflow Index (BFI). A similar principal is applied by Le Maitre and Colvin (2008) to assess the effectiveness of river flow statistics in characterising the contribution of groundwater to a river flow system. Flow statistics, extracted from the SPATSIM modelling system, were used to estimate the contribution of groundwater to streamflow and subsequently compared in terms of principal aquifer types in South Africa. The analysis found that the river flow statistics commonly used in river investigations (CVI, BFI and HI) and flow concentration statistics, in relation to the percentage zero flows in a catchment, are complex and variable on a national scale. When the relationships were investigated on a smaller scale, catchments within the Crocodile-Marico Water Management Area, they were still found to be complex. Nel, *et al.* (2004) suggest that there might not be a statistic which can be used as an indication of the groundwater contribution to baseflow as the Hydrological Index (HI) is used for river conservation planning. However, Le Maitre and Colvin (2008) found that zero flows

might be a useful indicator of groundwater baseflow, especially if combined with groundwater flow concentrations. Zero flow statistics and groundwater flow concentrations are recommended for further testing (Le Maitre and Colvin, 2008).

3.2.4. Hydrological Modelling

Hydrological models are used as simple, conceptual representations of the hydrological cycle or parts thereof to better facilitate understanding the processes occurring therein and consequently predicting what may happen in the future. South Africa is a semi-arid to arid area which has a number of different climate, rainfall and vegetation zones. There have been a number of successful developments and applications of hydrological models in the country despite the limitations of variable climatic conditions and data scarcity (Hughes, 2008). Hughes (2008) reviews hydrological models which have been specifically developed for the arid southern region of Africa. The hydrological model which has been the most extensively applied within southern Africa is the Pitman monthly time-step model. The model has however undergone numerous revisions since its development in the 1970's (Hughes, 2008). The Pitman model is an explicit soil moisture accounting model which represents interception, soil moisture and groundwater storages with model functions to allow inflow and outflow from these components (Hughes, 2008). Most model versions use a semi-distributed system where each sub-area has its own hydrometeorological inputs and parameter set, including components to simulate abstractions from distributed farm dams and direct flow from the river or major dam at the outlet of each sub-area. The modified Pitman model developed by Hughes (2004) incorporates a groundwater component to estimate groundwater baseflow. The Hughes version of the Pitman model has 24 model parameters, but there are guidelines available for parameter estimation, provided by the WR90 study (Midgley and others, 1994). Hughes (2008) includes that the model does however perform better in humid and temperate areas than in the more arid regions which is a consequence of the poorly defined real spatial variations in rainfall input, limitations in the temporal distribution of rainfall within a single month and the relatively simplistic approach to simulating runoff generation.

The Agricultural Catchments Research Unit (ACRU) is a multi-purpose model originating from an evapotranspiration study conducted by the University of Natal (Schulze, 1989). The model integrates water budgeting and runoff components of the terrestrial hydrological system with risk analysis. The model can be applied in crop yield modelling, design hydrology, reservoir yield simulation, irrigation water demand and supply, planning of optimum water resources and

regional water resource assessment among other applications (Schulza, 1989). One of main outputs of the model is a water balance, but the model is not directly geared towards the quantification of the exchange between groundwater and surface water. Hughes (2008) highlights that the model has a large number of parameters that require quantification and has been applied in mostly temperate and humid areas in South African to assess the impacts of various land use modifications.

The Variable Time Interval (VTI) model, developed at the Institute for Water Research (IWR), Rhodes University, South Africa has been applied to a large number of basins under the Southern African FRIEND programme (Hughes, 2008; Hughes, 1997). The VTI model is basically a daily time step model which can use smaller modelling intervals during periods of higher activity. The model requires short interval rainfall data for increased modelling interval times and the main moisture accounting routines are complex resulting in a large number of parameters (Hughes, 2008). Hughes (2008) concludes that the successful use of the VTI model would require a detailed understanding of the models structure, a good conceptual understanding of the main runoff generation mechanisms in the catchment and good quality climate data.

The Spatial and Time Series Information Modelling (SPATSIM) system was developed by the Institute of Water Research to replace the outdated integrated modelling environment package (HYMAS) used by the SA FRIEND project. The SPATSIM system has since been adopted as the main modelling environment to be used for the update of the South African water resources information system, WR90.

Hughes (2008) concludes his review of available hydrological models by highlighting that the success of any modelling study depends on the quality and appropriateness of the model as well as the quality of the input data and the level of experience of the user of the model. The better use of available data and the development of comprehensive guidelines for these models will serve to greatly improve model estimates, while only limited changes to existing models would be required (Hughes, 2008).

3.2.5. South African Groundwater Resource Assessment Phase II (GRA2)

The Groundwater Resource Assessment of South Africa – Phase Two (GRA2) initiated by the Department of Water Affairs in 2003 was aimed at building on the short-comings of Phase One (GRA1) and more accurately quantifying the groundwater resources in South Africa on a national scale. The project produced a methodology for the quantification of the country's groundwater resources, which includes algorithms for the estimation of storage, recharge, baseflow and the impact on the reserve as well as present groundwater use. Several datasets were produced as the methodology was applied to the production of a set of maps which can be used on various levels of planning and management (DWAF, n.d.). The assessment methodology is presented in a series of reports: Quantification, Planning Potential Map, Recharge and Groundwater/Surface Water interaction, Aquifer Classification and Groundwater Use. The assessment methodology for the quantification of groundwater – surface water interaction is given in DWAF (2006a) and referred to herein as the Sami model. This assessment methodology will still be discussed in detail. The WRSM2000 programme currently used by surface water planners in the Department of Water Affairs includes this groundwater – surface water interaction methodology. However, DWA (2009) highlights that this methodology has been reviewed by hydrogeologists and its applicability has been brought into question (Dennis, 2005; Sami and Witthüser, 2006). An additional method which is used to estimate the groundwater contribution to baseflow is the Hughes model, a modified version of the Pitman model developed by Hughes (2004) and based in the SPATSIM system. Both the Sami and Hughes model methodology will be described in detail.

3.3. The Sami Model

3.3.1. Overview of Sami model initiative

Double accounting, groundwater underflow and the simplistic linear approach incorporated by the widely-used MODFLOW modelling program were highlighted in DWAF (2006a) as problems to consider when developing a SW-GW interaction methodology. The methodology used by MODFLOW to determine surface water – groundwater interaction is considered simplistic due to the assumption therein that the relationship between head difference and water exchange is linear. DWAF (2006a) states surface water – groundwater interaction is not linear because hydraulic resistance will result in non-linearity as streamflow increases. The method used by MODFLOW is further critiqued on the grounds that it cannot process changes in streamflow over time and erroneously implies that the rate of flow from the river into the groundwater

system will equal the rate of flow to that river from the groundwater system. DWAF (2006a) suggests that these problems could be overcome by using a non-linear equation to simulate the interaction as this would be a more realistic approach.

The main aim of the Sami model was the development of a methodology which could determine the impacts of abstraction on baseflow without the necessity of modelling. The simulation of interactions under abstraction conditions is important as the related decline in groundwater levels can capture ambient groundwater that would have otherwise discharged as baseflow and in extreme cases induce streamflow into the groundwater system. Groundwater abstraction upsets the natural, steady-state condition of the water table by increasing recharge or decreasing discharge until a new equilibrium is reached. However, DWAF (2006) adds that until the new equilibrium is reached, where pumping is balanced by baseflow depletion, the abstraction results in aquifer storage depletion. Groundwater abstraction calculations should thus include both aquifer storage depletion and baseflow depletion components. The transition from aquifer storage depletion to streamflow depletion is a slow process and depends on the rate at which discharge can be captured (aquifer diffusivity), the location of pumping wells and time. DWAF (2006) concludes that determining the magnitude of potential groundwater abstraction should be aimed at developing relationships between abstraction and baseflow depletion, instead of simply on projected drawdown.

3.3.2. Methodology summary

The Sami model is based on an eight-stepped methodology, including determining the amount of groundwater discharging to the surface water body in question. The eight steps are hydrograph separation, estimation of recharge, groundwater storage increments from recharge, evapotranspiration from shallow groundwater, groundwater outflow, groundwater baseflow and transmission losses, interflow, and groundwater abstraction.

Hydrograph Separation

Method:

The Herold hydrograph separation method is performed, using monthly streamflow data, in order to separate the total baseflow (groundwater and interflow) contributing to the river from the total runoff.

The Herold hydrograph separation equation:

$$Q_{gi} = (Q_{gi-1} \cdot Decay) + (Q_{i-1} \cdot PG) \quad (3.13)$$

where,

- Q_{gi} is assumed to be the groundwater contribution,
- Q_{i-1} is the total streamflow of the preceding month,
- $Decay$ is a groundwater factor ($0 < Decay < 1$), and
- PG is a groundwater growth factor (%).

From Herold's equation it can be seen that the groundwater baseflow (Q_{gi}) is defined by previous groundwater after decay ($Q_{gi-1} \cdot Decay$) and rainfall induced recharge ($Q_{i-1} \cdot PG$), which includes interflow.

Inputs:

- Monthly streamflow data (WR90 data, observed gauging weir data or stochastic hydrographs)
- Previous months groundwater contribution (Q_{gi-1})
- Groundwater $Decay$ factor
- Groundwater growth factor (PG)

Assumptions:

The general assumptions of a hydrograph-separation include:

- Hydraulic characteristics of the contributing aquifer can be estimated from stream-discharge records,
- Periods of exclusively groundwater discharge can be reliably identified, and
- Stream-discharge peaks approximate the magnitude and timing of recharge events

The assumptions in the Herold method hydrograph separation methodology are:

- Streamflow below a certain, pre-defined parameter ($GGMAX$) is groundwater flow,
- This upper limit of groundwater ($GGMAZ$) can be correctly varied month to month based on the surface water runoff from the preceding month and calibrated parameters defining groundwater decay ($DECAY$) and groundwater growth (PG),
- Parameters $DECAY$, PG and $QGMAX$ can be appropriately selected on the grounds of a "realistic" division between surface water and groundwater, and
- The selection of $DECAY$, PG and $QGMAX$ is facilitated by a graphical output of total and groundwater hydrographs.

Estimation of Recharge

Method:

It is required to first estimate the recharge by calculating subsurface storage by reverse engineering of the Pitman model, in order to subdivide baseflow into groundwater baseflow and interflow.

- a) Calculating soil moisture storage (S):

The Pitman Runoff-soil moisture relationship equation:

$$Q = FT \left(\frac{S - S_L}{S_T - S_L} \right)^{POW} \quad (3.14)$$

where,

- S is the actual soil moisture storage (mm),
- S_L is the minimum soil moisture storage below which no runoff occurs,
- S_T is the maximum soil moisture storage,
- POW is the power function of the runoff-soil moisture curve, and
- FT is a parameter of the maximum baseflow depth at S_T .

The Sami model then reverse engineers this relationship to calculate the soil moisture storage (S), by using parameters S_L , S_T , FT and POW from WR90 and the total baseflow volume from the hydrograph separation for Q .

- b) Calculating monthly recharge (Re):

Once soil moisture is calculated, or input from WRSM2000 obtained, potential monthly recharge is calculated using the Hughes Recharge-soil moisture relationship:

$$Re = GW \left(\frac{S - S_L}{S_T - S_L} \right)^{GPOW} \quad (3.15)$$

where,

- Re is the potential recharge,
- S is the actual soil moisture (mm),
- S_L is now the soil moisture, below which there is no recharge (mm),
- GW is the maximum amount of recharge at maximum soil moisture (S_T) in mm, and
- $GPOW$ is the power function of the storage-recharge relationship.

DWAF (2006) state parameters GW and $GPOW$ can either be calibrated to achieve a fit with long term mean annual measurements obtained from other methods, or initial values could be chosen equal to the FT and POW parameters of the Pitman Runoff-soil moisture relationship.

Inputs:

- Baseflow value from hydrograph separation (Q), or time series of the Pitman S variable
- Parameters SL , ST , FT and POW
- Parameters SL , GW and $GPOW$

Assumptions:

Runoff-soil moisture relationship:

- The assumption is made that the Pitman runoff-soil moisture relationship estimates runoff comprised of interflow and groundwater baseflow, which under another assumption that the Herold method separates baseflow and interflow from total flow, allows baseflow from the Herold method to be used in the Pitman runoff-soil moisture relationship.
- The assumption is made that the same maximum soil moisture value (ST) can be used in both the Pitman runoff-soil moisture relationship and the Hughes recharge-soil moisture relationship.
- The parameter POW is assumed to represent the relationship between total basin moisture and the spatial distribution of this moisture.

Recharge-soil moisture relationship:

- The surface characteristics can be represented by a single storage, given that direct recharge can occur where there are bare rock areas.
- The depth of recharge can be estimated as a non-linear relationship with the ratio of current storage to the maximum storage.

Groundwater Storage Increments from recharge

Method:

Direct recharge from soil moisture is incremented to the groundwater aquifer storage. However, if the aquifer has reached its calculated capacity, the excess becomes interflow. As a result, aquifer recharge may be somewhat less than the calculated potential recharge. Actual recharge is calculated by subtracting the excess (interflow) from the potential recharge.

Thus, if the sum of groundwater storage and incremented recharge is greater than the aquifer capacity (CAP), then the potential recharge will equal the actual recharge and if less than the aquifer capacity (CAP), then the potential recharge minus the excess recharge ($EXCESS_1$) will be the actual recharge.

It should be noted that under abstraction conditions, the aquifer capacity could be increased by reducing groundwater storage which would lead to less excess recharge. The capacity of the aquifer is calculated as:

$$CAP = b \cdot S \quad (3.16)$$

where,

CAP is the capacity of the aquifer,
 b is the aquifer thickness estimated from the recommended drilling depth, and
 S is Storativity.

Inputs:

- Aquifer thickness and storativity (aquifer capacity)
- Groundwater storage

Assumptions:

- Aquifer capacity can be estimated by the product of aquifer thickness and aquifer storage.
- Once the aquifer capacity has been reached, any additional recharge would become interflow.
- Aquifer thickness can be estimated by the recommended drilling depth below groundwater level.

Evapotranspiration from shallow groundwater

Method:

Evapotranspiration from groundwater is calculated using:

$$ET = ((MAE \cdot MDIST \cdot CROP) - RAIN) \cdot \left(AREA \cdot \left(\frac{STORE - SWL}{CAP - SWL} \right) \right) \quad (3.17)$$

where,

MAE is the mean annual evaporation,
 MDIST is the monthly distribution of evaporation (%),
 CROP is the monthly A-pan crop factor for an appropriate cover,
 RAIN is the monthly rainfall variable,
 AREA is the area where Evapotranspiration from groundwater can take place,
 STORE is the variable of groundwater storage,
 CAP is the capacity of the aquifer, and
 SWL is the static water level.

The static water level is calculated as:

$$SWL = (CAP - \Delta\bar{h}) \cdot S \quad (3.18)$$

where,

$\Delta\bar{h}$ is the degree of annual groundwater level fluctuation,

CAP is the aquifer capacity, and

S is storativity.

Rainfall is subtracted from monthly evapotranspiration ($MAE \cdot MDIST \cdot CROP$), to obtain the evapotranspiration demand from groundwater. The evapotranspiration demand from groundwater is then multiplied by an aquifer storage factor ($AREA \cdot \left(\frac{STORE - SWL}{CAP - SWL}\right)$), to allow evapotranspiration to be decreased as groundwater storage is depleted. The calculated evapotranspiration is then decremented from groundwater storage.

Inputs:

- Parameters MAE , $MDIST$, $CROP$, $RAIN$, $AREA$, $STORE$, SWL and CAP
- Aquifer thickness and Storativity (CAP)
- Degree of annual groundwater level fluctuation ($\Delta\bar{h}$)

Assumptions:

- Evapotranspiration from groundwater only takes place when the evapotranspiration demand is not met by the amount of rainfall that month.
- Monthly evapotranspiration can be estimated by the product of mean annual evapotranspiration, monthly distribution of evapotranspiration and crop factor.

Groundwater Outflow

Method:

A Darcian approach is used to calculate groundwater outflow (underflow). DWAF (2006) state “groundwater outflow is calculated using the Darcian approach of the product of parameters transmissivity and hydraulic gradient oriented out of the catchment”. The maximum hydraulic gradient, defined by a parameter $HGRAD$ (channel gradient), is decremented as groundwater storage approaches the static water level, by multiplying the gradient with an aquifer storage factor:

$$i = HGRAD \left(\frac{STORE - SWL}{CAP - SWL} \right) \quad (3.19)$$

where,

$HGRAD$ is the maximum hydraulic gradient, and

i is the hydraulic gradient.

The calculated groundwater outflow is then decremented from groundwater storage.

Inputs:

- Transmissivity
- Parameters *HGRAD*, *STORE*, *SWL* and *CAP*

Assumptions:

Darcy's assumptions:

- The groundwater discharge is directly proportional to the transmissivity, hydraulic gradient.
- The flow dimensions are assumed to be one-dimensional, as this form of Darcy's law describes one-dimensional, pipe flow.
- The groundwater flow is slow and the Reynolds number is less than 10, where resistive forces of viscosity are dominant and laminar flow occurs.

Groundwater baseflow and transmission losses

Method:

Groundwater storage has been decremented by both the calculated evapotranspiration and outflow. Groundwater baseflow is now calculated as a function of the head difference between the new decremented groundwater level and the surface water level.

The groundwater head is calculated as the difference between storage and the static water level ($STORE - SWL$), while the surface water head is calculated as the monthly runoff volume divided by the catchment area ($Runoff/AREA$).

Effluent conditions are simulated when the groundwater head is greater than the surface water head (groundwater lost to the river as baseflow). Influent conditions are simulated when the groundwater head is less than the surface water head (transmission losses into the groundwater system). The model allows for the calculation of both groundwater and surface water head differences on a monthly time step, thus overcoming the short coming highlighted in MODFLOW of the unrealistic assumption of constant head conditions in the river.

Once the direction of the interactive water flow is determined, the groundwater baseflow or transmission losses are calculated using a non-linear equation to account for the effects of hydraulic resistance:

$$GW_{baseflow} = (1 - e^{HEAD \cdot (-0.05)}) \cdot BFMAX \quad (3.20)$$

where,

$GW_{baseflow}$ is the groundwater baseflow,

$HEAD$ is the difference between the calculated groundwater and surface water heads, and

$BFMAX$ is parameter of the maximum rate of groundwater baseflow.

Parameters $BFMAX$ and SWL can be calibrated on the principal that groundwater baseflow approximately equals, but doesn't exceed the total streamflow at the lowest flow period on a hydrograph. If there is no interaction between groundwater and surface water, then groundwater baseflow is set to zero and if the calculated groundwater baseflow exceeds the total baseflow, the groundwater baseflow is defaulted to the total baseflow volume.

Inputs:

- Groundwater storage less evapotranspiration and outflow ($STORE$)
- Static water level (SWL)
- Total monthly runoff
- Catchment area
- Maximum rate of groundwater baseflow ($BFMAX$)

Assumptions:

- The decremented groundwater storage parameter ($STORE$), resulting from the series of steps in the methodology is representative of the groundwater storage of the catchment
- The surface water head is sufficiently estimated by the monthly runoff volume divided over the catchment area.
- The assumptions from each calculation which is incorporated into the estimation of groundwater baseflow.

Interflow

Method:

DWAF (2006) state the interflow of a catchment (under virgin conditions) is calculated as the difference between the total baseflow and the calculated groundwater baseflow. However, abstraction decreases groundwater storage which increases aquifer capacity, implying that

interflow is expected to be less under abstraction conditions because of the potential recharge volume can become the actual recharge volume due to the increased capacity of the aquifer.

The depletion of interflow is calculated by:

$$Interflow = Q_g - GW_{baseflow} - EXCESS_1 + EXCESS_2 \quad (3.21)$$

where,

Q_g is the total baseflow,

$GW_{baseflow}$ is the groundwater baseflow,

$EXCESS_1$ is the recharge in excess of aquifer capacity under virgin conditions, and

$EXCESS_2$ is the recharge in excess of aquifer capacity under abstraction conditions.

Inputs:

- Total baseflow (Q_g)
- Groundwater baseflow ($GW_{baseflow}$)
- Aquifer capacity at virgin and modified conditions
- Potential recharge

Assumptions:

- Interflow is the difference between total baseflow and groundwater baseflow.
- Groundwater abstraction decreases groundwater storage which is assumed to increase aquifer capacity.

Groundwater abstraction

Method:

DWAF (2006) assume that groundwater abstraction depletes groundwater storage and groundwater baseflow in a non-linear manner. This non-linear relationship is dependent on the transmissivity and storativity of the aquifer, the distance from the stream and the time since abstraction started. The streamflow depletion solution is an analytical solution of the Glover and Balmer (1954) stream-depletion method (DWAF, 2009).

The Sami model groundwater abstraction equation:

$$\%GW = \frac{100}{(1+e^{(k_3+(k_2 \cdot t')})} \quad (3.22)$$

and,

$$t' = \frac{4Tt}{x^2S} \quad (3.23)$$

where,

$\%GW$ is the percentage of groundwater abstraction derived from groundwater storage,

t' is a dimensionless time parameter calculated as,

T is the transmissivity,

S is the storativity,

x is the distance to the stream,

t is the time since pumping began, and

k_3 and k_2 are parameters used to ensure that the percentage of abstraction from groundwater storage ($\%GW$) is 100% when pumping is commenced.

Inputs:

- Transmissivity (T), storativity (S), time since pumping started (t), and the distance to the stream from the abstraction point (x)
- Calibrated parameters k_3 and k_2

Assumptions:

- Groundwater abstraction is assumed to deplete groundwater storage and groundwater baseflow in a non-linear fashion depending on transmissivity, storativity, the distance to the stream and the time since pumping started.

Assumptions associated with the Glover and Balmer (1954) streamflow depletion method, which was based on the earlier work of Theis (1941) (Contor, 2011):

- The river is infinitely long.
- The aquifer is semi-infinite; the only boundary to the aquifer is the connected river.
- The results aggregate the effect upon the entire length of the river.
- The river is straight.
- The river fully penetrates and is in full communication with the aquifer.
- The aquifer is homogeneous and uniform.
- Saturated thickness of the aquifer is constant over time.

3.3.3. Review

The Herold method of hydrograph separation used in the Sami model is widely accepted and used in South Africa due to the fact that the only data requirement is readily available streamflow data. The method is an improvement with respect to earlier hydrograph separation techniques, but remains subjective according to Xu, *et al.* (2002). The general technique was criticised by Halford and Mayer (2000) on the grounds that the major assumptions of the method are commonly violated making hydrograph separation techniques poor tools for estimating groundwater discharge or recharge. Levy and Xu (2011) note that the hydrograph separation methods are indeed informative, but when applied to a single downstream hydrograph, were not able to account for spatial heterogeneity. On the other hand, Arnold and Allen (1999) reported to have found a good correlation between a separation technique and catchment mass balances for six USA streams and Wittenberg and Sivapalan (1999) successfully analysed streamflow to determine all the main components of groundwater balances, for a catchment in the humid part of Western Australia (Xu and Beekman, 2003). Levy and Xu (2011) refer to Parsons and Wentzel (2007) describing the Herold hydrograph separation method as estimating the groundwater contribution to streamflow by assuming that during each month the groundwater contribution will not drop below a certain amount. This statement holds a certain degree of uncertainty in the fact that the “groundwater contribution” could be interpreted as either total baseflow to the stream (groundwater + interflow) or as the actual groundwater contribution to baseflow (only groundwater). Subsurface baseflow was originally considered to be comprised of only groundwater discharge, but it is now understood to be comprised of both groundwater discharge and interflow. The question is whether Herold refers to the former or latter understanding of groundwater baseflow.

The Sami model does allow for an alternative route to be taken, in that instead of estimating the soil moisture value from the Pitman runoff-soil moisture relationship, that one could directly utilise time-series Pitman S-values from a WRSM2000 model run. This alternative method would compensate for the possible misconception in the hydrograph separation step. The time-series Pitman S-values from WRSM2000 (using default values) would be acceptable for a desktop or initial estimation of recharge for the quaternary.

The Sami model assumption that aquifer capacity can be estimated from the product of aquifer thickness and aquifer storage (Storativity) is reasonable. The determination of the aquifer thickness and storativity for the calculation however has uncertainties. The aquifer thickness is

estimated from the *Recommended Drilling Depth Below Groundwater level* from the Map of National Groundwater Resources Map of South Africa. Vegter (1995) considered aquifer storage as groundwater stored in the upper weathered and fractured zone. Vegter's *Optimum Drilling Depth* was determined by the addition of the depth to the fresh bedrock, estimated by statistical analysis of the data within the National Groundwater Database (NGDB) at the time, and the mean depth to groundwater level (DWAF, 2003). The use of this *Optimum Drilling Depth* is not the optimum situation as the data used to determine these depths is outdated and includes the assumption that the groundwater capacity is determined from the weathered subsurface (main water bearing depth) and a further distance of half the weathered depth (Figure 3-6). The aquifer capacity is based on a fixed groundwater level, but as groundwater is recharged, the water level will rise. In reality the depth for interflow and groundwater baseflow is variable. In times of high rainfall the groundwater level can reach the surface, thus excess recharge would result in direct surface runoff. Figure 3-7 indicates that if the aquifer capacity is based on a static water level, when the water table does rise, groundwater would then be accounted for as interflow. This highlights that there is ambiguity when determining the upper limit of the aquifer, namely is it defined by the static water level or by the ground surface. The methodology does not state where the storativity values are obtained from, but are probably obtained from the WR90 database. The methodology also does not explain how the groundwater storage is estimated. The method and inherent uncertainties associated with determining this initial groundwater storage will affect the final calculation of groundwater baseflow because this initial value will be the base from which the other water balance components are subtracted from to determine the final decremented groundwater storage.

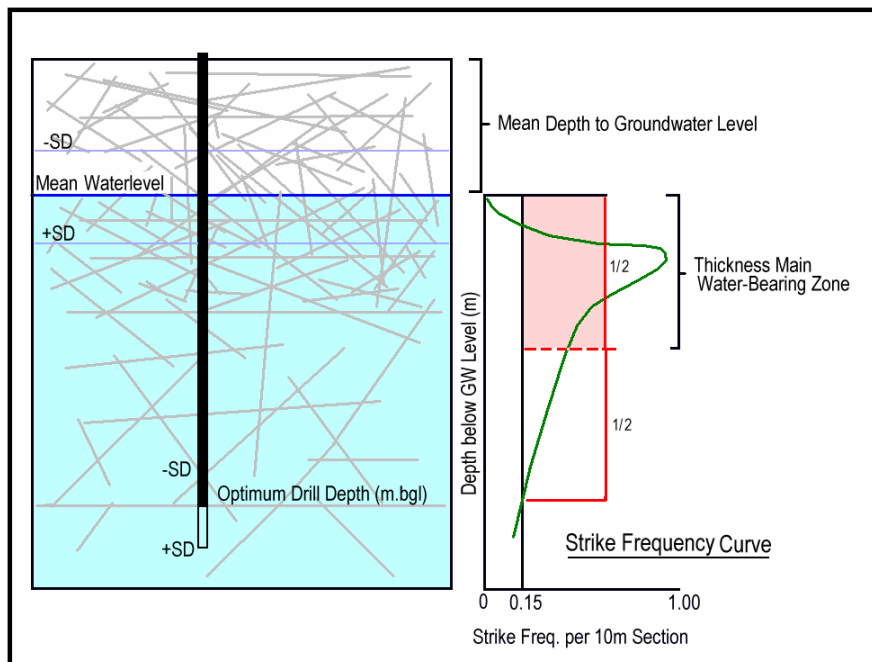


Figure 3-6 A schematic representation of the thickness of Vegter's main water-bearing zone and optimum drilling depth (Taken from DWAF, 2003)

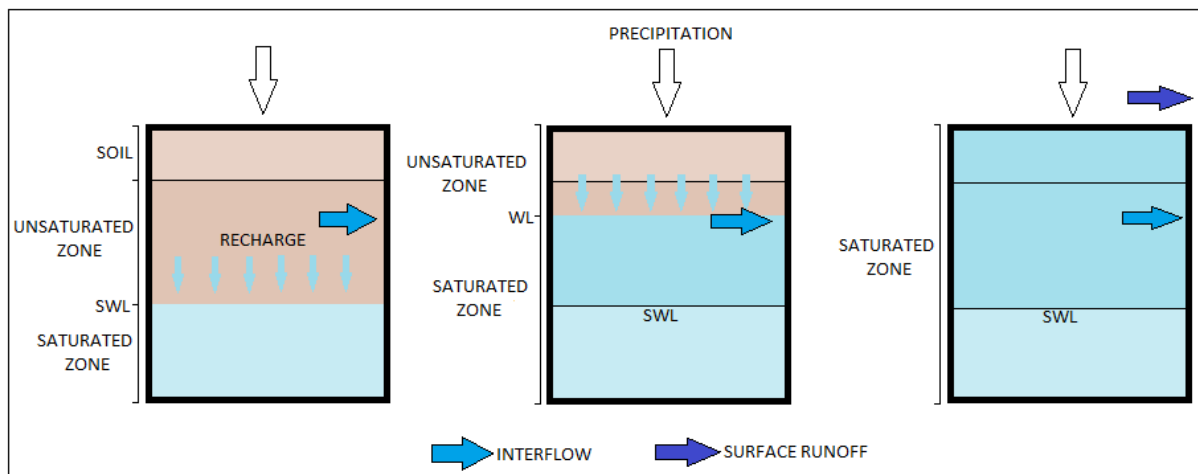


Figure 3-7 A graphical representation of the subsurface, indicating the variability of the groundwater level in relation to recharge.

The proposed equation for calculating evapotranspiration from groundwater is acceptable when one considers the difficulty in estimating evapotranspiration from groundwater. The assumption that evapotranspiration will only occur from groundwater if the evapotranspiration demand is not met from rainfall is incorrect, as this depends on the depth of the water level. There is an *evapotranspiration extinction depth*, the maximum depth at which water can move upwards under the forces of evapotranspiration (Figure 3-8). This depth is determined by the type of plants in the area. It can thus be seen that even if the evapotranspiration demand is not met by the rainfall, if the water level is below the *evapotranspiration extinction depth* than no evapotranspiration from the groundwater will occur.

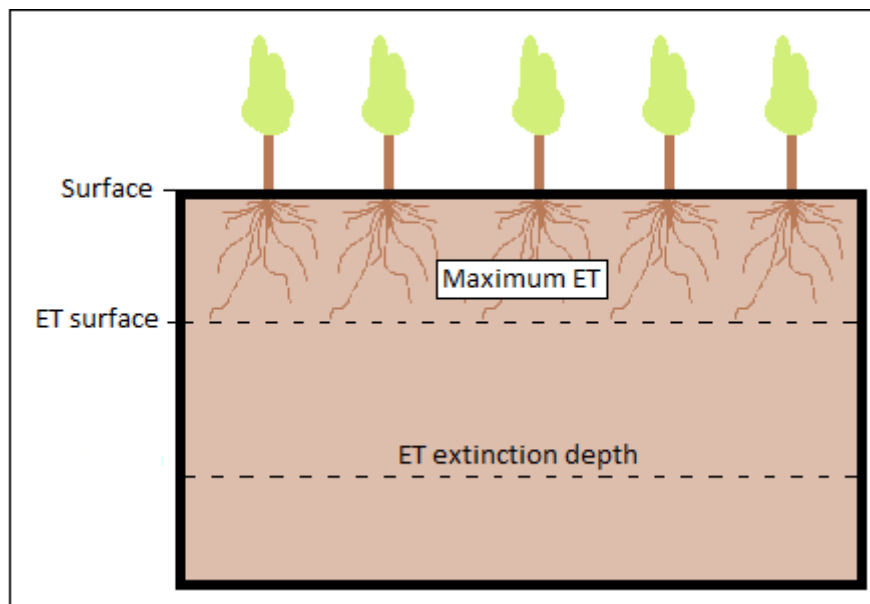


Figure 3-8 A schematic representation of the evapotranspiration extinction depth, the depth below which no evapotranspiration can occur.

The use of the Darcy equation in the Sami model to determine the groundwater outflow could lead to an over-simplification as the rate of groundwater outflow is directly proportional to the transmissivity and the hydraulic gradient. Phillips and Ingersoll (1998) report on the estimation of lateral subsurface outflow estimation by Lines (1979), where the Darcian approach was used. The equation used in this study is:

$$Q = TiL \quad (3.24)$$

where,

Q is the groundwater flow (m^3/d),

T is the transmissivity (m^3/d),

$HGRAD$ is the maximum hydraulic gradient, and

L is the length of the section perpendicular to the direction of flow.

However, in the Sami model methodology groundwater outflow is stated to be the product of parameters transmissivity and hydraulic gradient only, and does not mention the parameter L , the length of the section perpendicular to the direction of flow. The two components of this equation with the most uncertainty are transmissivity and the horizontal hydraulic gradient according to Phillips and Ingersoll (1998). Transmissivity values are often estimated at specific sites and then extrapolated over a larger area and the horizontal hydraulic gradient can vary along the length of the cross-section under investigation (Phillips and Ingersoll, 1998). DWAF (2006a) does not report on where the transmissivity values were obtained from.

The Sami model assumption that interflow is equal to the difference between the total baseflow and groundwater baseflow volumes is correct, if other factors such as bank storage are ignored. The assumption that groundwater abstraction decreases groundwater storage and thus

increases aquifer capacity is valid, but the loss of interflow and increased recharge would only be seen if the aquifer regularly reached its capacity. The extension of the Sami model to include the use of time-series of Pitman S variable to calculate recharge allows the user to remove the subjective nature of the hydrograph separation. The two routes for calculating recharge also differ in the manner in which interflow is calculated. The hydrograph separation technique calculates interflow as the residual total baseflow after groundwater baseflow has been deducted. The time-series of Pitman S-values technique calculates the groundwater baseflow and interflow without the catchment hydrograph separation volume. The calculation of interflow under virgin and modified conditions is covered in the Sami model methodology, but the independent calculation of interflow if the time-series Pitman S-values were used is not covered. Equation 3.21 could be used in the Pitman S method, but DWAF (2006a) does not state where this total baseflow value would be obtained from.

The groundwater abstraction equation in the Sami model is an analytical solution of the Glover and Balmer (1954) streamflow depletion method. The Glover and Balmer method is useful, but has a number of restrictive assumptions. The Sami model equation only indicates relative amounts of abstraction influence on the two available storages (groundwater storage and groundwater baseflow). This relative determination implies that in order to determine the quantitative amount of groundwater that can be abstracted would require the setting of an acceptable influence on the storages. The simplified water balance approach using averaged catchment parameters, results in the model not being able to accurately quantify the surface – groundwater interactions for single abstraction points.

Several shortcomings were found by Witthüser (2006) when reviewing the Sami model. Each quaternary catchment in the Sami model is discretized into two compartments allowing only vertical flow between the two layers instead of additional horizontal lagging of water movement as seen in other models such as the SHE model (Witthüser, 2006). Additionally, the global nature of the parameters is a limitation leading to Witthüser (2006) not recommending the use of the Sami model. Moseki (2013) also reviewed the Sami model and did not recommend the model for use in South Africa, as the model had not been validated for fractured-rock aquifers. Moseki (2013) also stated the model as a work in progress because it has the potential to make a valuable contribution, but the initial success of the pilot test in the Schoonspruit Catchment (Mare et al. (2007), cited by Moseki, 2013) has not been replicated in subsequent tests using different datasets and/or study areas.

In conclusion, the Sami model is able to simulate the interactions between surface and groundwater, and the effects of abstraction in a realistic manner for the quaternary catchment scale. However, Levy and Xu (2011) report on Seward *et al.* (2006) and Xu *et al.* (2002) suggesting that the estimation of groundwater discharge rates on a regional scale might not be an acceptable approach towards implementing the National Water Act (1998). The groundwater baseflow determination does indeed meet the initial criteria set by DWAF (2006) of overcoming the problem of constant surface water head, inherent in the MODFLOW program, by allowing the groundwater and surface water heads to differ from month to month, and taking into account the hydraulic resistance of the river bed by incorporating the natural log function into the estimation of groundwater baseflow. The groundwater storage used in the calculation of groundwater baseflow is influenced by the uncertainties associated with either the hydrograph separation and Pitman runoff-soil moisture relationship value or the times series Pitman S-values from WRSM2000 and the calculated initial groundwater storage, recharge, evapotranspiration, and groundwater outflow volumes. Hughes, Kapangaziwiri and Barker (2010) suggest that the overall uncertainty associated with the Sami model is more likely due to the estimation of parameters using the scarce data available in South Africa rather than the model structure. However, the model structure is not adequate for detailed groundwater investigations.

3.4. The Hughes Model

3.4.1. Overview

The procedure for determining the groundwater component of the ecological reserve is not as well established as the river component, in South Africa (Hughes, 2004). The main reason for this is the lack of quantitative information regarding the contribution of groundwater to surface water. Hughes (2004) state that the problem with quantifying the groundwater contribution to streamflow is in finding a method which is able to estimate recharge and groundwater discharge from available data, while allowing the integration with a surface water estimation approach which would be acceptable for both hydrologists and geohydrologists in South Africa. The widely-accepted Pitman model was determined by Hughes (2004) as a reasonable starting point, as the model is extensively utilized for the simulation of stream runoff. Hughes (2004) set to solve the problem of quantifying the groundwater contribution to streamflow by incorporating a recharge and groundwater discharge component into the existing Pitman model. The various

components of the Pitman model as well as the components incorporated by Hughes (2004) are discussed in detail.

3.4.2. Pitman Model

The Pitman model has two main inputs, monthly precipitation expressed as a mean annual percentage and monthly potential evapotranspiration. Additional compulsory data includes basin area, a time series of basin average rainfall, seasonal distributions of evaporation, irrigation water demand, other water demands and monthly parameter distribution factors. Optional input data consists of time series basin average potential evaporation, upstream inflow and transfer inflow. The original Pitman model flow diagram and structure is illustrated in Figure 3-9. Following this flow diagram the main algorithms or functions used to simulate the flow are described (Hughes, 2004).

Rainfall distributed function (RDF)

The input precipitation data is distributed by the rainfall-distributed function over time, using the RDF parameter in SPATSIM, which uses a cumulative mass curving over four iterations. The lower the RDF parameter the more evenly distributed the rainfall will be represented in the model. See Table 3-2 for a full list of parameters used in the Pitman model (Hughes, 2004).

Interception function

The Interception function is based on the parameter PI , which has the ability to vary seasonally and allows for two different vegetation types to be defined. The depth or amount of rainfall intercepted is calculated based on an empirical relationship between the set parameter PI and the monthly rainfall depth or amount. The amount of water subtracted from the total rainfall to account for interception (interception storage) is used to satisfy the potential evaporation rate. The rain water which is not intercepted then forms the input into the surface water runoff function (Hughes, 2004).

Surface Runoff function

The Pitman model allows for a parameter AI to be set, which represents the portion of the basin that is impervious, allowing for surface runoff to be directly generated from this portion of the basin (Hughes, 2004). The infiltration of precipitation on the remaining pervious section of the basin is calculated by means of the Surface Runoff function. The Surface Runoff function was originally a symmetrical triangular distribution of basin absorption rates based on a minimum ($ZMIN$) and a maximum ($ZMAX$) absorption rate. The amount of rainfall contributing to surface

runoff is represented by the area under the triangle between $ZMIN$ and the rainfall rate. In the SPATSIM version of the Pitman model, a parameter $ZAVE$ has been incorporated to permit an asymmetrical triangle, allowing for a distribution of runoff (Hughes, 2004).

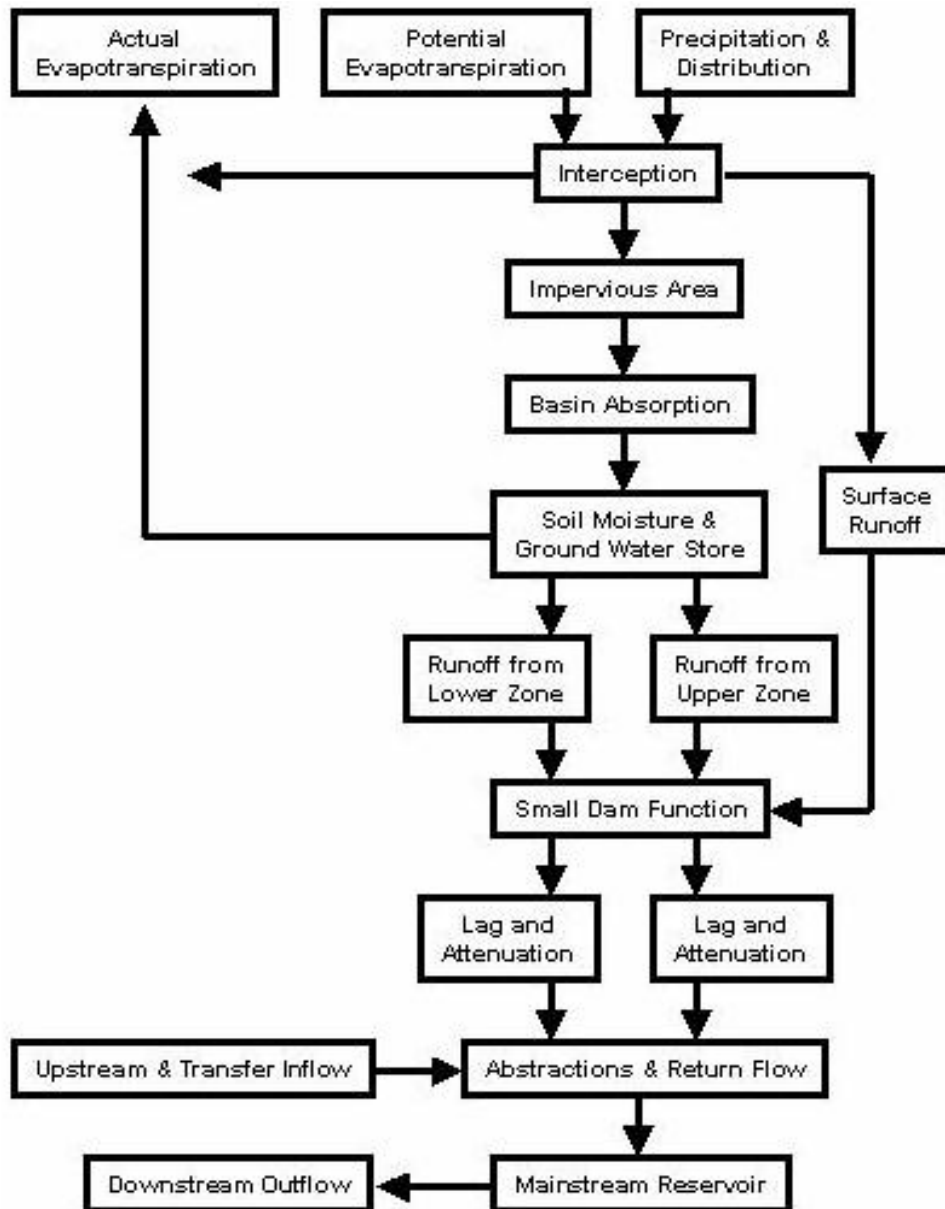


Figure 3-9 The basic structure and flow distribution of the Pitman model (Taken from Hughes, 2004).

Soil Moisture Storage and Runoff function

The precipitation that has not been intercepted or contributed to surface runoff will increment the soil moisture storage (S). If the soil moisture storage exceeds the maximum soil moisture storage capacity (ST) the surplus becomes runoff from the Upper zone (Figure 3-9).

Runoff from soil moisture storage is controlled through a non-linear relationship between runoff and storage through a parameter *POW*, controlling the power function of the relationship (Hughes, 2004; DWAF, 2006):

$$Q = FT \left(\frac{S-S_L}{S_T-S_L} \right)^{POW} \quad (3.25)$$

where,

FT is the runoff rate at full storage (mm/month) i.e. *ST* is exceeded,

S is the soil moisture storage (mm),

ST is the maximum soil moisture storage capacity,

SL is the minimum soil moisture storage below which no soil moisture runoff occurs, and

POW is the relationship between the runoff and soil moisture storage.

The runoff from groundwater is determined by a parameter *GW*, which is the maximum groundwater runoff, but Hughes (1997) highlights that there is no theoretical background to set this parameter value.

Evaporation from Soil Moisture Storage function

The Evaporation function is based on the parameter *R* ($0 < R < 1$) and the potential evaporation volume. A low *R* value indicates an effective evaporation loss which will continue to take place even at low soil moisture storage levels, where a high *R* value will indicate an evaporative loss which will cease at a higher soil moisture storage level. The value of *R* can be related to vegetation types, in that a low *R* value would indicate deeper-rooted vegetation. The model also allows for different rates of evaporation to be set by means of the parameter *FF*, which scales the potential evaporation for certain areas in order to consider different vegetation types within the basin.

Runoff Delays and Lags function

The runoff from the Upper zone (*ST* is exceeded), and the runoff from the Lower zone (groundwater) are lagged at different rates, controlled by parameters *TL* and *GL* respectively. The runoff considered to be groundwater by the model is lagged longer than the remaining runoff from soil moisture storage using the Muskingum equation (Hughes, 2004).

Artificial Modification functions

There are a number of additional components of the model which allow artificial modifications to the hydrological system to be simulated. These include a small dam routine, direct abstraction from river water and return flow functions.

Table 3-2 A list and description of all parameters in the original Pitman model (Taken from Hughes, 2004).

Table 1 Pitman model parameters.		
Parameter	Units	Description
RDF		Rainfall distribution factor. Controls the distribution of total monthly rainfall over four model iterations
AI	Fract	Impervious fraction of sub-basin
PI1 and PI2	mm	Interception storage for two vegetation types
AFOR %	%	area of sub-basin under vegetation type 2
FF		Ratio of potential evaporation rate for Veg2 relative to Veg1
PEVAP	mm	Annual basin potential evaporation
ZMIN	mm month-1	Minimum basin absorption rate
ZAVE	mm month-1	Mean basin absorption rate
ZMAX	mm month-1	Maximum basin absorption rate
ST	mm	Maximum moisture storage capacity
SL	mm	Minimum moisture storage below which no runoff occurs
POW		Power of the moisture storage-runoff equation
FT	mm month-1	Runoff from moisture storage at full capacity (ST)
GW	mm month-1	Maximum runoff from groundwater
R		Evaporation-moisture storage relationship parameter
TL and GL	months	Lag of runoff (surface and groundwater respectively)
AIRR	km ²	Irrigation area
IWR	Fract	Irrigation water return flow fraction
EFFECT	Fract	Effective rainfall fraction
RUSE	m ³ × 10 ⁶ year-1	Non-irrigation demand from the river
MDAM	m ³ × 10 ⁶	Small dam storage capacity
DAREA %	%	sub-basin above dams
A and B		Parameters in non-linear dam area-volume relationship
IRRIG	km ²	Irrigation area from small dams

3.4.3. Hughes Components

The Pitman model has been re-coded by both the original author and others, resulting in a number of subsequent versions and additional components. However, the basic form of the Pitman model has been preserved. Hughes (2004) has incorporated two new functions into the Pitman model in order to more efficiently quantify the interaction taking place between groundwater and surface water. These two new components consist of a Recharge function and a Groundwater Discharge function.

Recharge function

The new recharge function is based on the assumption that recharge will occur where there are rock outcrops and that the surface characteristics of the area can be represented by a single storage parameter. By defining a parameter below which no recharge will occur (when soil moisture capacity has been reached), the depth of recharge can be estimated by a non-linear relationship between the current storage and the maximum storage (Hughes, 2004). Hughes (2004) adapts the original Pitman soil moisture - runoff relationship (Equation 3.25) to now estimate recharge by redefining the SL , GW and POW parameters. SL is redefined as the soil moisture threshold below which no recharge occurs, GW is redefined as the maximum amount of recharge, and POW is redefined as $GPOW$ and now represents the relationship between recharge and current storage (S). Figure 3-10 is a graph of the two different power parameters, POW and $GPOW$ indicating their relative differences (Hughes, 2004).

The newly defined recharge – soil moisture function:

$$RE = GW \left(\frac{S - S_L}{S_T - S_L} \right)^{GPOW} \quad (3.26)$$

where,

RE is the estimated recharge depth or amount (mm),
 GW is the maximum rate of recharge (mm/month), and
 $GPOW$ is the new relationship between recharge and soil moisture storage.

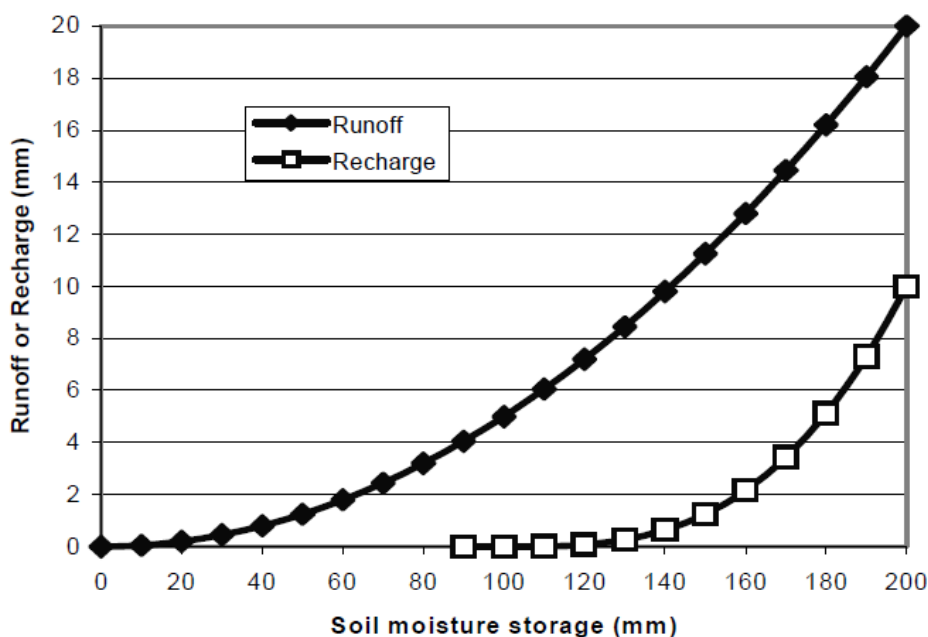


Figure 3-10 A graphical representation of the original soil moisture – runoff relationship (POW) and the redefined recharge – moisture relationship ($GPOW$) (Taken from Hughes, 2004).

Groundwater Discharge function

The Groundwater Discharge function aims to reduce the complexity of the spatial geometry of the basin to apply simple groundwater discharge principals. Hughes (2004) states the first step is to represent the basin as a square and the rivers as parallel lines separated by drainage slopes (Figure 3-11). The discharge is considered one-dimensional to further simplify the system. The number, length and width of the separating drainage slopes as well as the effective drainage density can thus be calculated from the modified basin area. However, the drainage density is a model parameter which can be deduced from maps and an understanding of the basin and channels (Hughes, 2004).

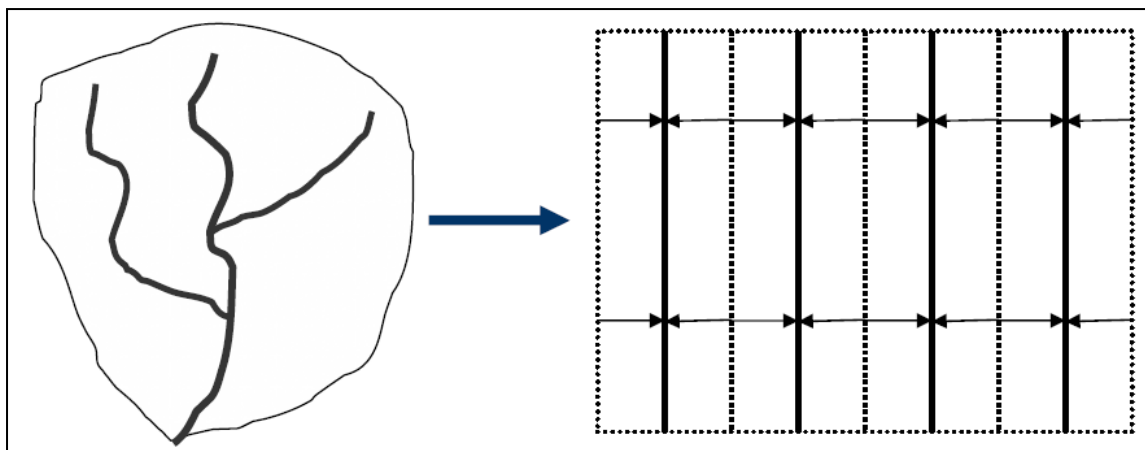


Figure 3-11 Conceptual simplification of a drainage basin as a square and rivers as parallel lines separated by drainage slopes. Solid lines are rivers, dotted lines are drainage divides and arrows indicate the direction of drainage on the eight drainage slopes (Taken from Hughes, 2004).

The number of channel lines is calculated first (Hughes, 2004):

$$Total\ channel\ length = Drainage\ Density \times Area$$

From Figure 3-11, it follows that:

$$Total\ channel\ length = No.\ drainage\ slopes \times Area$$

and,

$$No.\ drainage\ slopes = Integer\ even\ value\ of \left[Drainage\ density \times \frac{\sqrt{(Area)}}{2} \right]$$

where,

$$Drainage\ slope\ width = \frac{\sqrt{(Area)}}{No.\ of\ drainage\ slopes}$$

The volume of the “wedge” of groundwater stored under the drainage slope, as shown in Figure 3-12 for a single drainage slope, can be calculated as follows if the lower boundary is the river at the bottom of the slope (Hughes, 2004):

$$\text{Wedge volume} = (\text{Drainage width})^2 \times \text{Gradient} \times \frac{\text{Drainage length}}{2}$$

and,

$$\text{Volume of water in wedge} = \text{wedge volume} \times \text{Storativity}$$

Finally, outflows from this wedge representing the groundwater discharge to a river within a single slope element, can be calculated as (Hughes, 2004):

$$\text{Discharge} = \text{Transmissivity} \times \text{Gradient} \times \text{Time step} \times \text{Channel length}$$

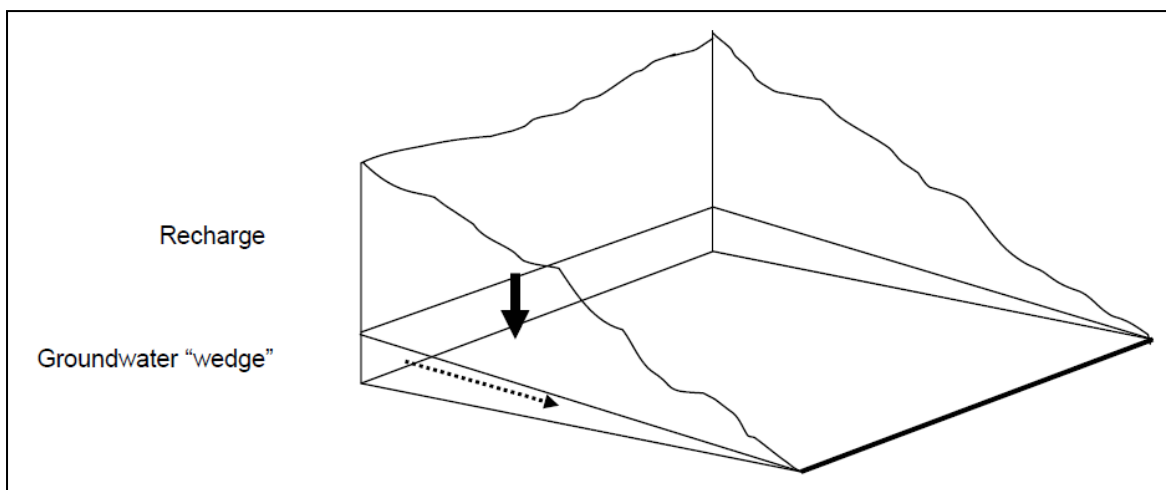


Figure 3-12 A single drainage slope element with the corresponding “wedge” representing the groundwater body that is above the conceptual river (Taken from Hughes, 2004).

Hughes (2004) performed several checks to ensure that the water balance was still achieved after the modifications to the Pitman model were done. The modified Pitman model is referred to as the Hughes model. The Hughes model was applied to two river basins in South Africa to test the modifications. The river basin examples showed that the modifications were a small improvement on the results from the original model. Hughes (2004) includes that from a perspective of representing processes involved in the runoff generation, the modifications and the modified parameter set are considered an improvement, but there is insufficient field data to confirm the runoff generation processes.

3.4.4. Review

The representation of the investigated basin as a square and the corresponding geometric representation of groundwater flow towards a river is an extremely simplistic view and ignores the complexities associated with groundwater flow. Hughes (2004) argues that the representation is sufficient as the calculations used are simple geometric equations. Following the new groundwater discharge function there is no longer a need to lag the groundwater component or Lower zone runoff, which renders the parameter GL as well as the lag routine unnecessary.

It is noted by Hughes (2004) that the model formulation would not be appropriate where groundwater flow did not follow the surface water flow. The fact that groundwater flow does not follow the surface water derived quaternary catchments in certain areas has been a contentious issue regarding groundwater resource assessments, but due to the fact that groundwater water cannot be directly observed there is not enough information to identify these locations. There are a number of situations listed by Hughes (2004) where the current modified version of the Pitman model application is not recommended. These situations include groundwater abstractions, evaporative losses of groundwater discharge from riparian areas, groundwater discharge to aquifer compartments in adjacent sub-basins and where the groundwater level is below the river level.

Hughes (2004) conclude that it is not suggested that the new components incorporated into the Pitman model are a completely realistic representation of groundwater flow to a river, but they are still effective and the parameters should be quantifiable from currently available data. The problem with these simplifications is, at what point the increasing degree of simplification will begin to misrepresent the groundwater flow processes occurring within a basin.

3.5. Discussion of Surface water – Groundwater Interaction methods

There is a trend in international methods of quantifying surface water – groundwater interaction towards numerical modelling solutions, as seen in the United States of America (USA), United Kingdom (UK) and Australia. However, a great expense accompanies the use of these numerical solutions as seen in the extensive monitoring infrastructure in the UK (Environment Agency, 2005a). There are a number of site specific study which have been done (Oxtobee and Novakowski (2002); Environment Agency (2005b); Allen *et al.* (2010); Tetzlaff and Soulsby (2008); Mencio and Mas-Pla (2008); Lapworth *et al.* (2009)), but Allen *et al.* (2010) conclude that the groundwater – surface water interactions can still only be partially understood even after extensive study at a site scale due to the inherent heterogeneity of the groundwater system. The advantages and disadvantages of each method are different and it has been suggested by numerous authors that a combination of methods would result in the best estimate of groundwater – surface water interaction (Environment Agency (2005b); Allen *et al.* (2010); Oxtobee and Novakowski (2002); Rosenberry and LaBaugh (2008)).

Hydrograph separations are extensively used all over the world to determine surface water – groundwater interactions because stream flow records are widely available. However, the use of a hydrograph separation technique alone has been highlighted as a poor determination method for the groundwater component of baseflow (Halford and Mayer (2000); Wittenberg and Sivapalan (1999); Xu *et al.* (2002)). A number of improvement techniques have been suggested for the hydrograph separation techniques, along with the incorporation of chemical data (Australian Government (2012a)).

In South Africa there is a clear trend towards the use of hydrological models (Pitman model, Hughes model, Sami model) for the estimation of groundwater – surface water interaction at a catchment scale, while hydrograph separation techniques are seen to be favoured for local scale investigations. There are guidelines for selecting a method that is more suitable for specific investigation or area, but comprehensive governmental guidelines specifically aimed at the quantification of the surface water – groundwater interaction are lacking.

3.6. Review of the Mixing Cell Model

3.6.1. Historical applications

The earliest mention of the mixing cell concept in relation to a hydrological system appears to be when Wentworth (1948) suggested the use of an array of cells with perfect mixing to explain a transition zone at a moving interface between fresh and salt water in a coastal aquifer. The use of mixing cell models and discrete reservoirs to model hydrologic systems has continued for decades, seen in the use by Craig (1957), Dooge (1959), Eriksson (1971), Simpson (1988), Harrington *et al.* (1998) and Partington, *et al.* (2010). The mixing cell model has also been extensively used within the chemical engineering field to investigate the movement of and chemical changes within and among reactor vessels. The names given to the mixing cell model within the chemical engineering field are varied, where Levenspiel (1972) refers to tank-in-series models, Deans and Lapidus (1960) refer to finite-stage models and Himmelblau and Bischoff (1968) refer to population-balance models. These models differ slightly in algorithms used and assumptions, but all make use of a cell of some sort as the basic sub-division of the system. The chemical engineering models differ to the hydrologic models in that they are applied on different scales (Campana, 1975).

Limitations on some of the earlier mixing cell models were highlighted by Campana (1975). Mixing cells models developed by chemical engineers allow for complex network configurations, but real-world hydrologic systems would not conform to the models geometric configurations due to scale differences. Earlier models also limited the user to a specific number of input and outputs that could be utilized per cell.

The Discrete State Compartment (DSC) model, developed by Simpson in 1972, overcame these limitations in that it consists of a set of interconnected cells of any desired size through which the transport of an incompressible fluid and dissolved matter is represented by a sequence of finite states and in theory these states could assume an infinite number of values. The model obtains solutions by iterating a recursive equation derived from the continuity equation and fluid and tracer transport can be modelled simultaneously (Campana, 1975).

The basic equation for each cell in an assigned network of a DSC model is (Simpson, 1973):

$$S(N + 1) = S(N) + (BRV(N + 1) \cdot BRC(N + 1)) - (BDV(N + 1) \cdot BDC(N + 1)) \pm R(N + 1) \quad (3.27)$$

where

$S(N+1)$ is the cell state or amount of substance in cell at iteration $N+1$,
 $S(N)$ is the cell state or amount of substance in cell at iteration N ,
 N is the iteration number,
 $BRV(N+1)$ is the boundary recharge volume at iteration $N+1$,
 $BRC(N+1)$ is the boundary recharge concentration at iteration $N+1$,
 $BDV(N+1)$ is the boundary discharge volume at iteration $N+1$,
 $BDC(N+1)$ is the boundary discharge concentration at iteration $N+1$, and
 $R(N+1)$ is the source/sink term for iteration $N+1$.

Equation 3.27 is a discrete form of the continuity equation and states that the amount of a substance in a cell at iteration step $N+1$ will equal the amount of substance in the cell at iteration step N , plus the amount that entered the cell at iteration $N+1$, minus the amount that leaves the cell, plus or minus any amount that was added from the external environment or subtracted from the cell to the external environment.

Two different algorithms were described by Simpson (1973), namely the Simple Mixing Cell (SMC) and the Modified Mixing Cell (MMC). The SMC is equivalent to the conventional mixing cell, while the MMC is somewhere between the perfect mixing of the SMC and pure piston flow (displacement only, no mixing). The SMC is also described as the “in-mix-out” algorithm in that the inflow tracer mixes with the cell contents and is then discharged, whereas the MMC is described as the “in-out-mix” algorithm in that the cell discharges before the inflow tracer mixes with the cell contents (Campana, 1975). The finite-state mixing cell model was unique due to the fact that it modelled mass transport in an aquifer system without the usual requirement of a dispersion coefficient and it is not a black box model (Campana, 1975).

Woolhiser, *et al.*, (1982) developed a method to quantify the inflows from several sources to a stream reach. The method requires that the chemical characteristics of the inflows are known and further assumes that water moving through a unique environment will have a characteristic chemical composition. Thus, the water in the stream would consist of a mixture of water from the different sources, with each representing a unique environment. Pinder and Jones (1969) estimated the proportion of stream discharge from groundwater and surface water using the differences in Total Dissolved Salts (TDS) content between the two. This method is limited to two sources and is thus of limited use. Visocky (1970) and Hall (1970) also made use of the differences in chemical composition of different sources.

Mathematically the Woolhiser, *et al.* (1982) method consists of a water mass balance equation and a mass balance equation for each of the selected ionic species, which are each equated to an error term. The unknown inflow rates are then estimated by minimizing the square percentage errors in each of the mass balance equations using quadratic programming. The method has inherent assumptions, of which the most important is that each ionic species is conservative within the reach. Woolhiser, *et al.* (1982) found that the method is less sensitive to errors in the chemical analyses if the concentration of each ion is divided by the concentration of that ion in the mixture, which is equivalent to minimizing the sum of squared percentage errors. Errors in estimates of a particular inflow are related to the proportion of the total ionic load contributed by that inflow relative to the total ionic load contributed to the river reach. If a significant inflow is absent from the calculation, the related error is shifted to the inflows with the most similar chemical composition (Woolhiser, 1982).

Another mixing cell model was developed by Adar (1984) to estimate the recharge rates from various sources into an aquifer by means of chemical and isotopic data. The model would be of the greatest use in areas with complex hydrogeological structures for which there is limited hydrologic information. The model's approach is a combination of two of the previously discussed mixing cell models, namely the Simpson (1973) and Woolhiser, *et al.* (1982) models. The Adar model makes use of the idea by Woolhiser, *et al.* (1982) for estimating unknown flows and the interconnected mixing cell concept from Simpson (1973). The model divides the investigated aquifer into mixing cells and mass balance equations are written for each cell expressing the conservation of water, dissolved chemical constituents and stable environmental isotopes. The mixing cell model developed by Adar (1984) estimates recharge rates by simultaneously solving the mass balance equations using quadratic programming (Adar, 1984).

The use of calculated transient groundwater fluxes from MODFLOW as the input data to a Compartmental Mixing Cell (CMC) model to simulate the transport of hydrochemical and isotopic species in regional groundwater systems is described by Harrington *et al.* (1998). The main advantage of their integrated modelling approach is that quantitative estimates of aquifer processes can be obtained with greater confidence than if they were determined using only one of the approaches. Harrington *et al.* also highlight that the main disadvantage of the CMC is the inherent assumption of complete mixing. Attempts have been made to allow for varying degrees of mixing within cells (Allison and Hughes, 1975), but this creates a large amount of parameterisation (Harrington *et al.*, 1998).

Partington *et al.* (2010) highlight the fact that tools for quantifying the groundwater component of streamflow are not readily available in the latest generation of fully integrated spatially distributed models. Partington *et al.* (2010) thus developed a Hydraulic Mixing Cell method (HMC) for quantifying the groundwater component of streamflow in fully integrated spatially distributed models. The mixing cell is based on the Modified Mixing Cell (MMC) developed by Campana (1975), but it differs in that it requires only hydraulic data.

3.6.2. Discussion of MCM applications

The popular trend in quantifying the groundwater component of streamflow is towards numerical modelling and more recent applications of the MCM have been found to follow the same trend, with the integration of mixing cell modes into numerical groundwater flow models. Harington *et al.* (1988) and Partington *et al.* (2010) are examples of this and are both adequate methods for determining the groundwater component of streamflow. However, the data required to set up the models is substantial and often not available in certain areas of South Africa. Considering the fact that South Africa has large areas of data paucity, where it is not feasible to setup numerical models, the usage of the mixing cell model developed by Adar (1984) could be advantageous in that it only requires water quality and minimal flow data.

Hydrochemistry and environmental isotope data have been traditionally limited to a qualitative geohydrological tool, used to support or reject hypotheses on prevailing flow regimes. However, there is a great potential of this data in quantitative geohydrology. Adar (1984) states accordingly that valuable information may be lost by excluding hydrochemical and stable isotope data from a quantitative study. Harington, *et al.* (1998) state to increase the confidence in a model, the incorporation of additional information is required. Environmental tracers and isotopes have the potential to provide such additional data to facilitate such an increase in confidence. The Adar mixing cell model is simple enough to be applied to data scarce areas while the incorporation of environmental tracer data to a water balance can increase the confidence in the estimated fluxes.

Chapter 4 Methodologies

The mixing cell model, tracer method and data collection methodologies applied for the study are presented within this chapter. The methodology of the mixing cell model developed by Adar (1984) is discussed in terms of the basic principal of the method, how this principal is mathematical applied, the software available and how the MCM has been slightly adapted for the quantification of groundwater—surface interactions. The methodology of the additional chemical hydrograph separation method (Tracer method) is briefly covered. Field investigations and data collection methodologies are discussed lastly.

4.1. Mixing Cell Model

4.1.1. Basic Concept

The concept of a mixing cell is essentially based on the continuity equation. The one-dimensional continuity equation states the amount of inflow to a system will equate the amount of outflow with no change in storage, for the considered time step (Figure 4-1).

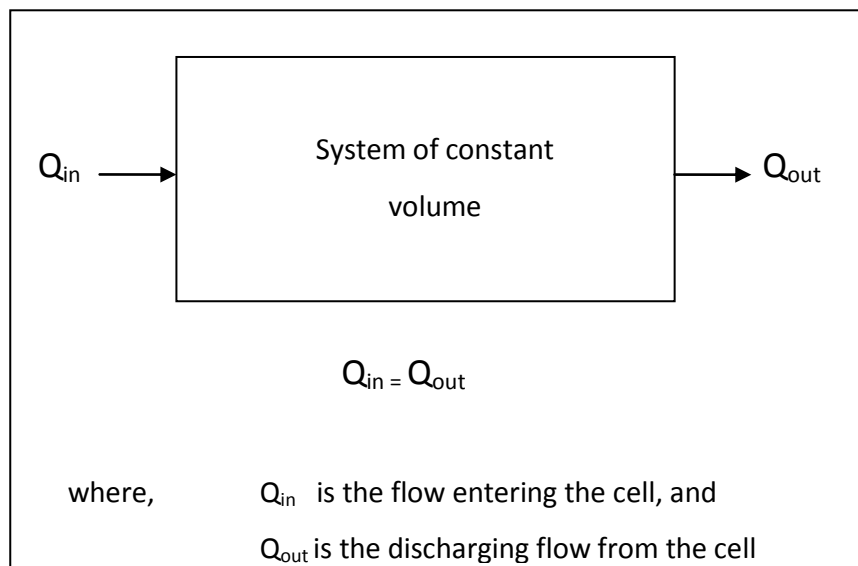


Figure 4-1 Basic principal of the mixing cell model

The mixing cell model builds on this foundation by sub-dividing a system into one or more mixing cells. A water balance equation is expressed for each cell to describe the movement into and out of the cells. The MCM requires that each of the inflows, present in the water balance equation, are chemically defined by a set of tracer concentrations. This water quality data is then used to describe a chemical mass balance equation for each cell. The chemical mass

balance equation serves to constrain the water balance equation in order to produce better estimates of the various unknown inflows to the system, than estimates made from the sole use of a water balance equation.

The main assumption of the mixing cell model is that any quantity or chemical property entering the cell is instantaneously dispersed throughout that cell. This implies that there is perfect mixing within the cell and the concentration of the entire cell and the discharging flow are the same. The “walls” of each mixing cell are assumed impermeable except for allocated connections to either an adjacent cell or the external environment (Campana, 1975).

4.1.2. Methodology

The mixing cell model developed by Adar (1984) was intended for the identification and quantification of multiple recharge sources, subsurface fluxes and physical aquifer parameters based on easily obtainable natural tracer concentration data.

The model relies on three types of conceptual models commonly applied in hydrology:

- 1) The evaluation of the motion of water and solutes with a multi-compartmental mixing cell model as suggested by Simpson (1988).
- 2) The solution of a set of water and dissolved constituents mass balance equations via a quadratic programming optimization scheme used by Woolhiser, *et al.* (1982) and Adar (1984).
- 3) A mathematical model combining an inverse process to estimate compartmental conductances and storage coefficients distributed in a multi-compartmental model for a non-steady flow as described by Adar and Sorek (1989) and Adar (1996).

The application of the MCM to estimate the groundwater component of streamflow only makes use of the first two conceptual models because a steady-state approach is used and determining the physical parameters of the system is not a priority. The mathematical principals used to express and solve the mass balance equations for each cell are discussed below. The mathematical model presented here is taken from the “Quantitative evaluation of flow systems, groundwater recharge, and transmissivities using environmental tracers” paper by Adar (1996) presented in the *Manual on Mathematical models in isotope hydrogeology* and therefore not separately referenced.

The assumptions associated with this model as stated by Adar (1996):

- 1) Tracers are considered conservative, all reactions including dissolution and precipitation are considered negligible and the spatial change in the concentration of the solute is solely the result of dilution.
- 2) The assumption inherent in all mixing cell models is that any quantity entering a cell is instantaneously dispersed throughout the cell, implying complete mixing.
- 3) Seasonal pulsation of fluxes for each cell can be represented by mean values covering a time interval in which the hydraulic head may be regarded as a constant.
- 4) Transport of dissolved constituents is dominated by advective forces, i.e. the compartmental Peclet number is infinite.
- 5) Concentrations of solutes, which are constant within each cell over a specific time step, are measured and known together with the concentration of the same solute or tracer in the inflow and outflow components.
- 6) Flows entering the aquifer system are known qualitatively, where most of the source-sink and discharge flow components are known quantitatively and qualitatively.

A set of mass balance equations for the water and solute fluxes over a given time period are written for each cell. The water balance for a fluid with constant density within the n -th compartment (cell) is expressed as:

$$Q_n - W_n + \sum_{i=1}^{I_n} q_{in} - \sum_{j=1}^{J_n} q_{nj} = S_n \cdot \frac{dh_n}{dt} \quad (4.1)$$

where,

- I_n is the number of sources which flow enters the n -th compartment,
- J_n is the number of leaving flows from the n -th compartment,
- q_{in} and q_{nj} is the the fluxes from the i -th source or compartment into the n -th one and from the n -th into the j -th cell, respectively
- Q_n and W_n is the the fluid sources and sinks, respectively
- S_n is the the storage capacity within cell n , and
- h_n is the the hydraulic head associated with that compartment.

If one identifies two points in time (t_1 and t_2) where the hydraulic head is the same, for example at the beginning and end of a season, the total magnitude of the derivative (dh_n/dt) has not changed over that time period. Hence, we obtain Equation 4.2:

$$\frac{1}{\tau} \int_{t_1}^{t_2} S_n \cdot \frac{dh_n}{dt} dt = 0 \quad (4.2)$$

By integrating the water balance (Equation 4.1) over the same time period, the following quasi steady-state equation is obtained:

$$\bar{Q}_n - \bar{W}_n + \sum_{i=1}^{I_n} \bar{q}_{in} - \sum_{j=1}^{J_n} \bar{q}_{nj} = 0 \quad (4.3)$$

All the parameters in Equation 4.3 have the same meaning as in the water balance (Equation 4.1), but now represent average values over the time period. A schematic diagram of a compartmental system is represented in Figure 4-2 to illustrate the flow parameters within the water balance equation (Equation 4.1).

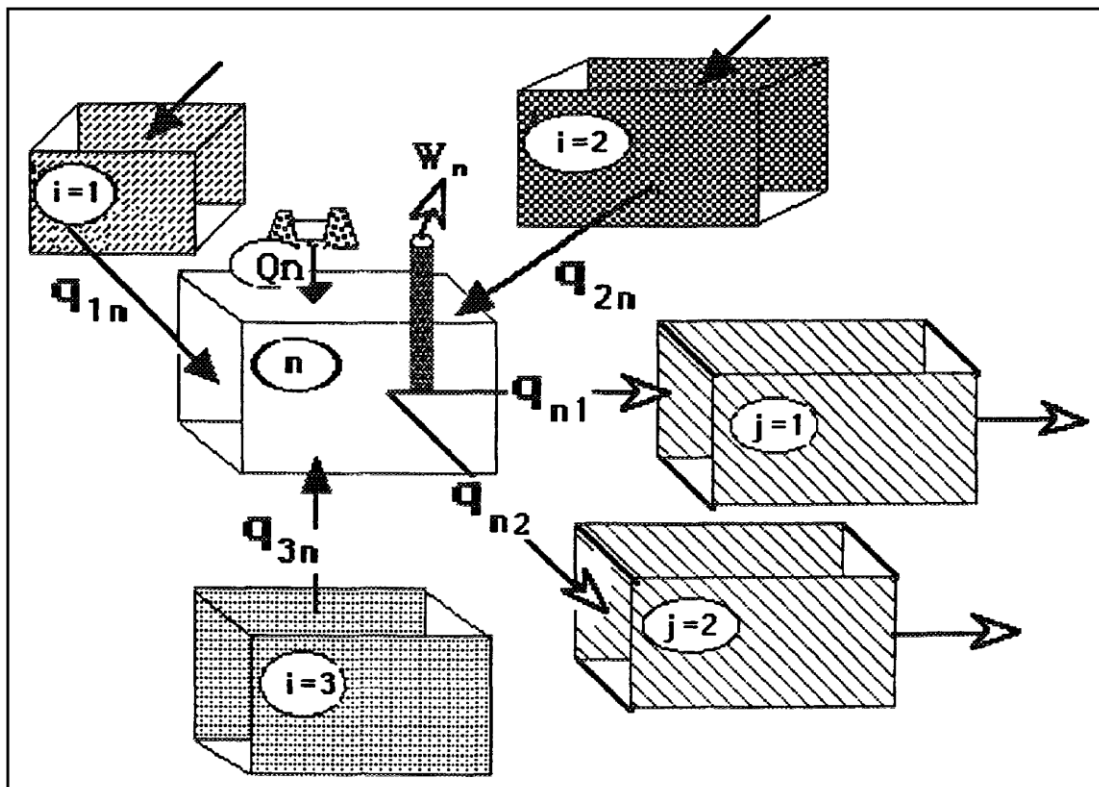


Figure 4-2 A schematic diagram explaining the parameters used in the mixing cell mathematical model. The parameter q indicates an unknown flux; i and j denote the direction of the flux, in and out respectively; Q_n denotes the known inflow to cell and W denotes the known out flow from the cell (Taken from Adar, 1993).

When the mixing cell concept is applied for quasi steady-state variations of concentrations and taking into account assumption 1 and Equation 4.3, the mass balance for a dissolved constituent k , in cell n is:

$$\tilde{C}_{nk} \bar{Q}_n - \bar{C}_{nk} [\bar{W}_n + \sum_{j=1}^{J_n} \bar{q}_{nj}] + \sum_{i=1}^{I_n} \bar{q}_{in} \bar{C}_{ink} = 0 \quad k = 1, 2, \dots, K \quad (4.4)$$

where,

\bar{C}_{ink} is the average concentration of solute k entering cell n with the incoming flux from cell i ,
 \bar{C}_{nk} is the average concentration of the k -th constituent within cell n , and
 \tilde{C}_{nk} is the average concentration of k associated with the source Q_n .

The mass balance equations, Equation 4.3 and Equation 4.4, should not be expected to close (without error) if real data is used, due to a number of possible sources of error. The water balance may not close due to the inadequacy of the assumptions that the annual change in storage is zero within the cell or errors in the measurement of Q_{nj} , C_{ink} and C_{nk} , or both. An error in identifying and measuring fluxes or rates of pumpage will also cause the water balance not to close. The chemical balance may not close due to one or more of the assumptions not holding for several of the solute species, or sampling and analytical errors. Mass balance errors may also be caused by incorrect quantifications of cell concentrations (Adar, 1984).

In order to account for the above mentioned inconsistencies, an error term is introduced to Equation 4.3 and Equation 4.4, to obtain Equation 4.5 and Equation 4.6, respectively:

$$\bar{Q}_n - \bar{W}_n + \sum_{i=1}^{I_n} \bar{q}_{in} - \sum_{j=1}^{J_n} \bar{q}_{nj} = e_n \quad (4.5)$$

$$\tilde{C}_{nk} \bar{Q}_n - \bar{C}_{nk} [\bar{W}_n + \sum_{j=1}^{J_n} \bar{q}_{nj}] + \sum_{i=1}^{I_n} \bar{q}_{in} \bar{C}_{ink} = e_{nk} \quad (4.6)$$

where,

e_n and e_{nk} are the deviations from the water and solute mass balances in cell n , respectively.

Equation 4.5 and Equation 4.6 are combined and expressed in matrix form:

$$\underline{C}_n \underline{q}_n + \underline{D}_n = \underline{E}_n \quad (4.7)$$

where \underline{C}_n is a matrix with known concentrations in cell n , of the form:

$$C_n = \begin{bmatrix} 1 & , 1 & , \dots & , -1 & , -1 & , \dots & , -1 \\ C_{1n1} & , C_{2n1} & , \dots & , C_{I_n n1} & , -C_{n1} & , \dots & , -C_{n1} \\ C_{1n2} & , C_{2n2} & , \dots & , C_{I_n n2} & , -C_{n2} & , \dots & , -C_{n2} \\ \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots \\ C_{1nK} & , C_{2nK} & , \dots & , C_{I_n nK} & , -C_{nK} & , \dots & , -C_{nK} \end{bmatrix} (K + 1)(I_n + J_n)$$

where the first row accounts for the water balance,
 the other K rows express the solute mass balances,
 K is the total number of tracers used in the analysis,
 $-$ denotes the coefficients which are associated with outgoing fluxes, and
 C_{ink} is the concentration of the k -th species flowing into cell n from cell i .

\underline{q}_n is vector matrix of the unknown fluxes through the boundaries of cell n , of the form:

$$q_n = [\bar{q}_{1n} \quad , \bar{q}_{2n} \quad , \dots \quad , \bar{q}_{I_n n} \quad , \bar{q}_{n1} \quad , \bar{q}_{n2} \quad , \dots \quad , \bar{q}_{nJ_n}] (I_n + J_n) \cdot 1$$

\underline{D}_n is a vector containing elements that are measured and known quantitatively in cell n , of the form:

$$D_n = [(\bar{Q}_n - \bar{W}_n), (\bar{C}_{n1}\bar{Q}_n - C_{n1}\bar{W}_n), (\bar{C}_{n2}\bar{Q}_n - C_{n2}\bar{W}_n), \dots, (\bar{C}_{nK}\bar{Q}_n - C_{nK}\bar{W}_n)](K + 1) \cdot 1$$

\underline{E}_n is the error vector matrix, of the form:

$$E_n = [e_n, e_{n1}, e_{n2}, \dots, e_{nK}](K + 1) \cdot 1$$

The unknown flux components in the aquifer can now be estimated by minimizing the square error sums J . By assembling the square error terms over all the cells the following equation is obtained:

$$\begin{aligned} J &= \sum_{n=1}^N (C_n q_n + D_n)^T W (C_n q_n + D_n) \\ &= \sum_{n=1}^N [E_n^T W E_n] \end{aligned} \quad (4.8)$$

where,

$()^T$ is a transpose matrix function, and

W denotes the diagonal matrix comprised of weighting values about estimated error expected for each of the terms building the mass balance for the fluid and the dissolved constituents.

The weighting matrix (W) can also reflect the level of confidence in a tracer's degree of conservation. The lower the confidence in the conservation of the tracer, the lower the weighting factor ($0 < W < 1$) assigned to that tracer. Adar (1996) states q_n is deconstructed into linear and non-linear components allowing the square error sums (J) to be minimized in order to estimate q_n . The quadratic optimization scheme used by Adar (1984) to minimize J was developed by Wolf (1967).

4.1.3. MCM Software

Adar and Küll (n.d.) developed an Excel® Add-in, based on the mathematical model presented by Adar (1984), for running a compartment or mixing cell model under in a steady state. The *Mixcel* software includes a special Mixing-cell Input Generator (MIG). The MIG serves to simplify the procedure of preparing data, as it quite literally builds an input file in Excel for the Wolf solver algorithm program which is also included in the *Mixcel* downloaded. The MIG computer code provided here is however restricted to a steady flow and steady hydrochemical flow system. The software can be downloaded free of charge from the website [<http://www.uhydro.de/doku/en/models/compartmentmodels>] and includes explanations and examples.

The latest version of the MCM developed by Adar (2012) is based in Microsoft® Access and was made available for this research project. Two codes were developed by Adar, one for steady state (MCMsf – A Mixing Cell Model for Steady Flow solver code) and one allowing for transient state (MCM_FTS – A Mixing Cell Model for Transient Flow solver code). The simpler steady state code was made use of to keep the data required and model complexity to a minimum. A step-by-step guide to using the MCMsf programme based on an example flow system is provided in Appendix A.

4.1.4. MCM adaption for the quantification of SW-GW Interactions

The Mixing Cell Model (MCM) developed by Adar (1984) is aimed at the identification and quantification of multiple recharge sources, subsurface fluxes and physical aquifer parameters, by means of representing the aquifer as mixing cells and assigning chemically-defined recharge sources to these cells to determine the recharge flux into that aquifer. To apply the MCM concept in quantifying the groundwater baseflow to a river, the main cell focus was changed from the aquifer to the river. The river section under investigation is represented as mixing cells, instead of traditionally defining the aquifer as the main mixing cells. Chemically-defined groundwater inflow sources are then assigned to the river cell to determine the groundwater flux into that river.

The conceptual representation of the surface water – groundwater system for the application into the MCM is shown in Figure 4-3. The example quaternary catchment is conceptually represented as a single box model flow system, comprising of one mixing cell. The box model flow diagram then forms the basis for both the water and chemical mass balance equations. The main river within the area is presented by a single mixing cell, the River Cell (Figure 4-3). The River Cell is assumed to have a uniform chemical composition for the considered time step (the assumption appears reasonable if the different time scales of surface and groundwater flow are considered). Water quality data from within this river segment is used to chemically-define this River Cell in the MCM model. The upstream inflow into the river (IN), the three tributary inflows and the groundwater source along the river stretch are assigned chemical compositions from water quality data, so their flow rates to the River Cell can be estimated by the MCM. The conceptual model neglects evaporation from the stream and the adjacent channel aquifer as well as surface water use, but these parameters can be easily incorporated as constant loss term in a specific MCM application.

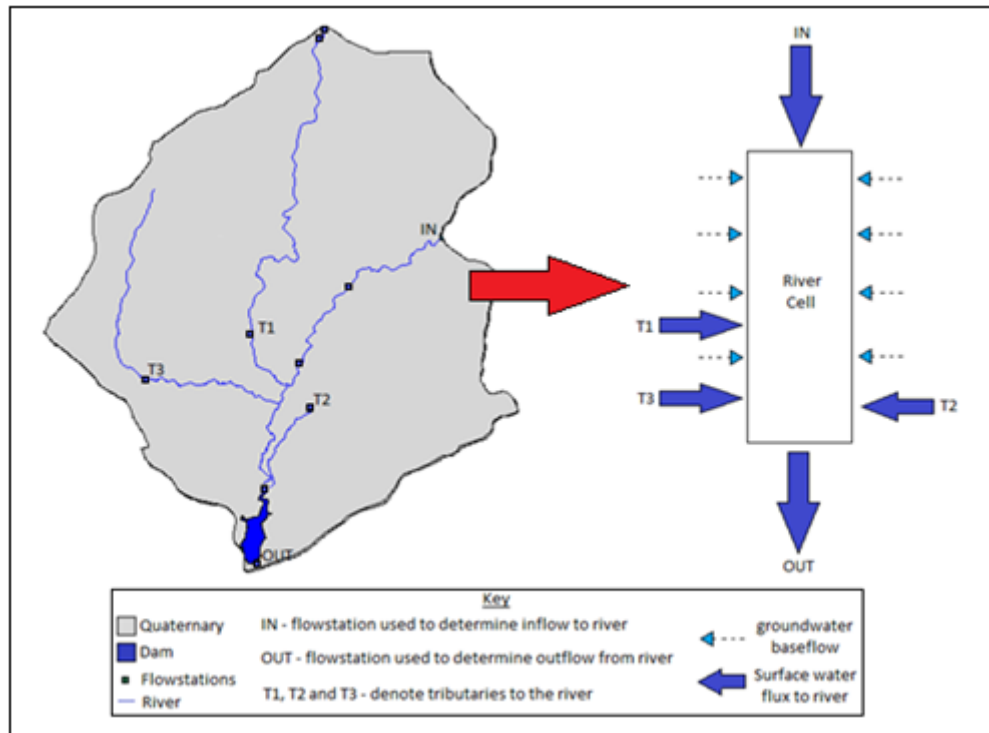


Figure 4-3 Conceptual representation of a catchment for the SW-GW interaction MCM application.

The water balance equation based on the box model flow representation of the example quaternary catchment shown in Figure 4-3, is:

$$Outflow(Q_{OUT}) = Inflow(Q_{IN}) + Tributary\ inflow(Q_T) + Groundwater\ inflow(Q_{GW})$$

The chemical mass balance equation based on the box model flow representation of the example quaternary catchment shown in Figure 4-3, is:

$$Q_{OUT} \cdot C_{OUT} = Q_{IN} \cdot C_{IN} + Q_T \cdot C_T + Q_{GW} \cdot C_{GW}$$

where,

C_{OUT} is the tracer concentration of the River Cell,

C_{IN} is the tracer concentration of the defined inflow,

C_T is the tracer concentration of the defined tributary inflow, and

C_{GW} is the tracer concentration of the defined groundwater source.

4.2. Tracer Method

4.2.1. Methodology

The Tracer method is a chemical hydrograph separation method described in an Australian Government document on surface water – groundwater interaction quantification methods (Australian Government, 2012a and 2012b). The method makes use of tracer concentrations for river flow, runoff and groundwater inflow and the total measured river flow at the point of investigation to determine the unknown groundwater discharge. The Tracer method assumes

that the total outflow volume is made up of runoff and groundwater inflow, and the total tracer concentration at this point is controlled by the ratio of input from each of these two sources. The proportion of river flow interpreted as sourced from groundwater discharge is calculated using a chemical mass balance equation (Australian Government, 2012b):

$$\frac{Q_g}{Q_t} = \frac{c - c_r}{c_g - c_r} \quad (4.9)$$

where,

c , c_r and c_g are the tracer concentrations for the river, runoff and groundwater,
 Q_t is the measured total river flow, and
 Q_g is the volume of groundwater inflow.

The data collection of this method is similar to that of a traditional hydrograph separation method, where measurements of river water quality are made at regular time intervals along with the traditional flow measurements at a single point. A continuous record of flow and water quality data allows for a continuous record of groundwater baseflow fluxes to be determined. The accuracy of the groundwater baseflow volumes determined by this method is dependent on the adequate differentiation of the source water and the assignment of the end-member concentrations, namely the river, runoff and groundwater tracer concentrations. The runoff end-member concentration can be defined by sampled rainfall or canopy throughfall. The runoff member could also be defined by the lowest tracer concentration measured in the river. Similarly, the groundwater end-member could be defined by the maximum tracer concentration measured in the river, but this is not recommended as the groundwater baseflow would tend to be over-estimated. The mean tracer concentration of groundwater sampled in boreholes is recommended for defining the groundwater end-member as it would give more accurate results (Australian Government, 2012a and 2012b).

The Tracer method can be used to determine the relative or volumetric proportion of groundwater inflow at any point along a river, where sufficient tracer data is available. The estimated groundwater baseflow at this point then includes the entire catchment upstream of this point. The method thus cannot provide specific information on where the calculated groundwater baseflow is occurring within the catchment. Considering this limitation the method would be best suited to smaller catchments where the groundwater baseflow volume could be more accurately determined (Australian Government, 2012a and 2012b).

4.3. Field Investigation

South Africa has a high density of data scarce areas and the Mixing Cell Model (MCM) requires surface water and groundwater quality data to define the various flows of a system to estimate the groundwater contribution to streamflow. To allow for a comprehensive model run of the MCM it was decided to perform a supplementary river and groundwater sampling run. The sampling run was performed along a section of the Modder River outside of Bloemfontein, South Africa. The investigated reach covers approximately 130kms of the Modder River, from the Rustfontein Dam in the south-east to the Krugersdrift Dam in the north-west (Figure 4-5).

4.3.1. Previous investigations

The Water Research Commission (WRC) and the University of the Free State (UFS) have established a surface water – groundwater interaction test site on the Modder River at the base of the Krugersdrift Dam located just outside of Bloemfontein, South Africa. The test site covers a limited area and simulates a local scale investigation of surface water – groundwater interaction. Water quality data was collected by Gomo (2011) for 15 boreholes, two river sample points and a seepage sample from alongside the river. A number of sampling runs were performed by Gomo (2011) for these sample points, but only the water quality data collected in January and August of 2011 were considered for this investigation.

4.3.2. Additional field work

A field investigation of the surface water and groundwater along the Modder River from the Rustfontein Dam to the Krugersdrift Dam was performed to increase the scale of investigation from a local scale as previously done at UFS Test Site to a large catchment scale. The various locations where river water was sampled were based on geological structures (outcrops of dolerite) as well as accessibility (Figure 4-5). The various locations at which groundwater were sampled are based on availability and accessibility of boreholes (Figure 4-5). While the sampling points were controlled by accessibility, the aim of the field work was to collect as many water quality samples as possible from both surface water and groundwater. Two sampling runs were performed, one from the 29th to the 31st October 2012 and another from the 29th January to the 1st February 2013. The first sampling run comprises 19 river water samples and 11 groundwater borehole samples, while the second sampling run comprises 23 river water samples (three additional tributary samples and one additional Modder River sample) and 13 groundwater borehole samples. Figure 4-5 indicates the location of all sampled points in the second sample run (January 2013) as well as selected images of the investigated area.

4.3.3. Surface water/groundwater sampling

River water samples were collected by means of a pole sampler to allow for the collection of the sample to be taken as close as possible to the middle of the river (Figure 4-4a). Groundwater samples from uncovered boreholes were collected by means of a bailer with no purge pumping, while groundwater from covered boreholes were either sampled from the reservoir dam or a tap directly fed by a water pumping system supplied by a borehole (Figure 4-4b). Unfiltered water samples were collected in 1L PVC plastic bottles after each bottle was rinsed twice with the sampled water. Chemical analysis of water samples was undertaken by the Institute of Groundwater Studies (IGS) Laboratories in the University of the Free State, Bloemfontein. The water samples underwent water quality analysis in which the following determinants were analysed for (Table 4-1):

Table 4-1 Physical and Chemical determinants for water quality analysis

Group	Determinant
Physical determinant	pH and EC
Major cations	Ca, Mg, Na, K
Major anions	Cl, SO ₄ , PO ₄
Aggregate determinants	PAIk, MAIk, CA Hard, Mg hard, Total Hard, TDS (sum)
Other elements	Si, NO ₂ (N), NO ₃ (N), F, Br
Metals	Cu, Fe, Mn, Zn, Al, Ni
Micro constituents	Ba, Cd, Cr, Mo, As, Be, Pb, Se, V

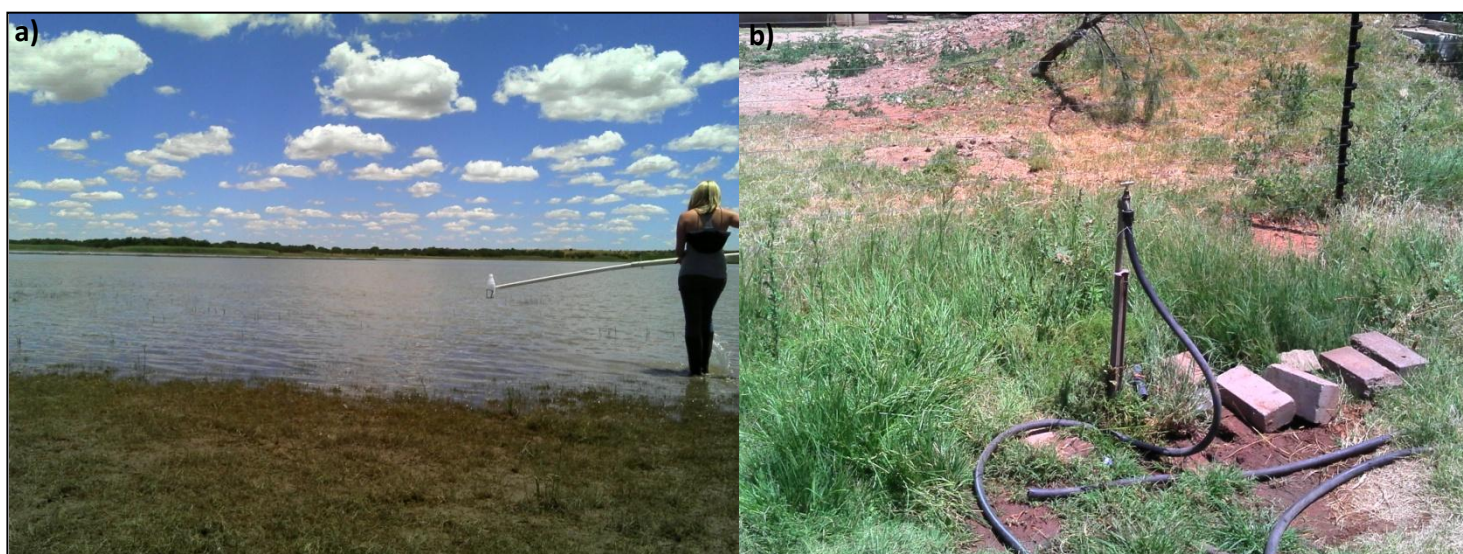


Figure 4-4 Examples of sample water collection methods used a) a pole sampler is used to collect river samples closer to the middle of the river and b) groundwater collected from a tap directly fed from a borehole.

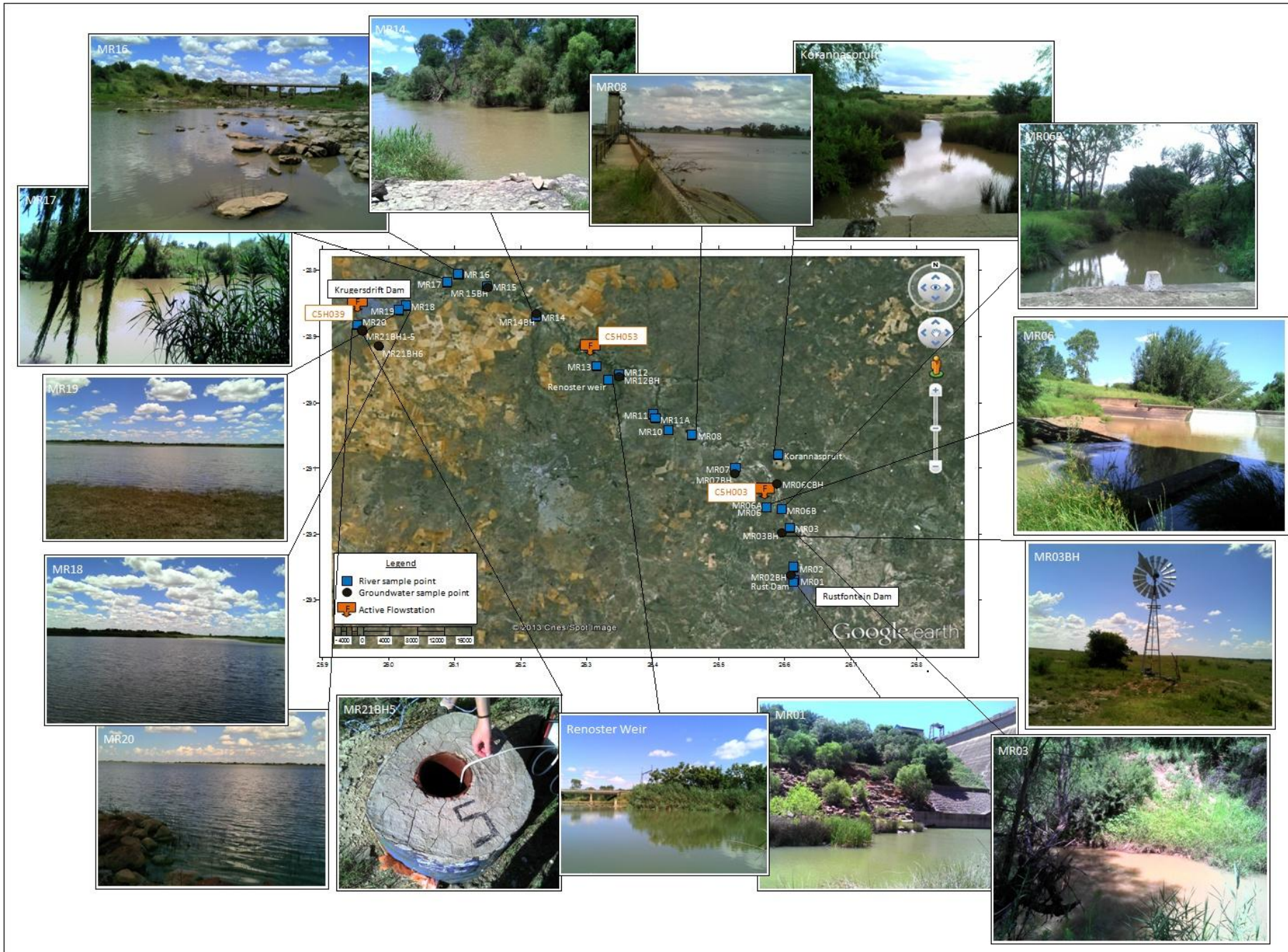


Figure 4-5 Surface water and groundwater sampling sites for the sampling run performed in January 2013. Images from selected points are shown to realistically describe the study area.

4.3.4. Additional data collection

Data collected from the following sources was utilised:

- National Chemical Monitoring Programme for Surface Water (NCMP):
The NCMP is a Department of Water Affairs initiative that aims to provide data and information on the inorganic surface water quality of South Africa's water resources. This surface water quality data was utilised to perform the additional desktop case studies and to supplement the field study investigation. The data is used in the MCM to define the water quality of the river cells. The database is available from the NCMP website: http://www.dwa.gov.za/iwqs/water_quality/NCMP/default.asp.
- Hydrological Services - Surface Water (Data, Dams, Floods and Flows):
The hydrological services provided by the Department of Water Affairs include flow volume data from a network of flowstations in South Africa. These flow volumes are made use of in both the desktop and field investigations. The database is available from the DWAF website: <http://www.dwa.gov.za/Hydrology/>.
- The National Groundwater Archive (NGA):
The National Groundwater Achieve (NGA) is the web-enabled database system provided by the Department of Water Affairs that allows capturing, viewing, modification and extraction of groundwater related data. Groundwater quality data from this database was used to define the groundwater sources of baseflow in both the desktop and field investigations. Elevation and groundwater level data was also made use of to represent the topography and groundwater levels for each study area. Information regarding registration and use of the database is available from the DWAF website: <http://www.dwa.gov.za/Groundwater/NGA.aspx>.
- WARMS database:
The Water Authorisation and Use (WARMS) database provides information on the volume of water allocated to each registered water user. This data is made use of to estimate the amount of water being abstracted from the river section in each study site. WARMS data can be requested from: warmsdatarequests@dwaf.gov.za.
- Groundwater Resource Assessment Phase II (GRA2):
The Groundwater Resource Assessment Phase II (GRA2) database of groundwater baseflow estimates for the Pitman, Hughes and Sami models is made use of for comparison purposes.

Chapter 5

Pilot Study of the Mixing Cell Model

The Mixing Cell Model (MCM) is applied to three study sites to investigate the applicability of the model to quantify the groundwater component of streamflow. The first study area is situated in the Free State Province, South Africa (Figure 5-1) and comprises quaternary catchments C52A – C52H; including the UFS surface water – groundwater interaction test site. The area is located on the edge of a regionally-defined zone of zero groundwater baseflow (Figure 5-2). The second study area is located in the Limpopo Province, South Africa and comprises quaternary catchments A42A – A42C, where calibrated Sami and Hughes model groundwater baseflow estimates are available for additional comparison and validation. The quaternary catchments are located within a regionally-defined high groundwater baseflow zone (Figure 5-2). The third study site is situated within the Northern Cape Province, South Africa and comprises of the quaternary catchment D73F, which falls within a semi-arid region (Figure 5-1). The quaternary catchment D73F is located in the middle of a regionally-defined zero groundwater baseflow zone (Figure 5-2).

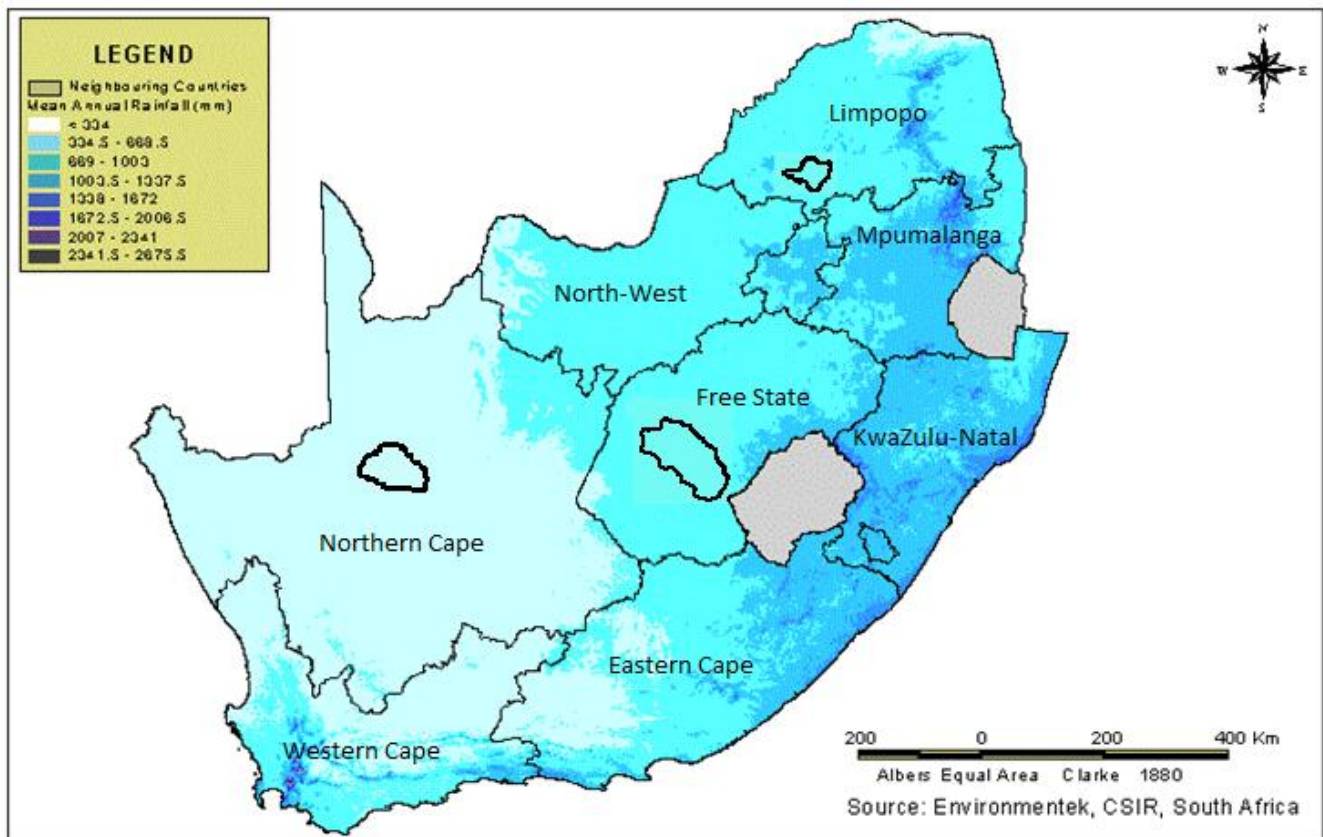


Figure 5-1 A rainfall distribution map of South Africa indicating the location of the three pilot study areas (Modified from Department of Environmental Affairs and Tourism, 1999).

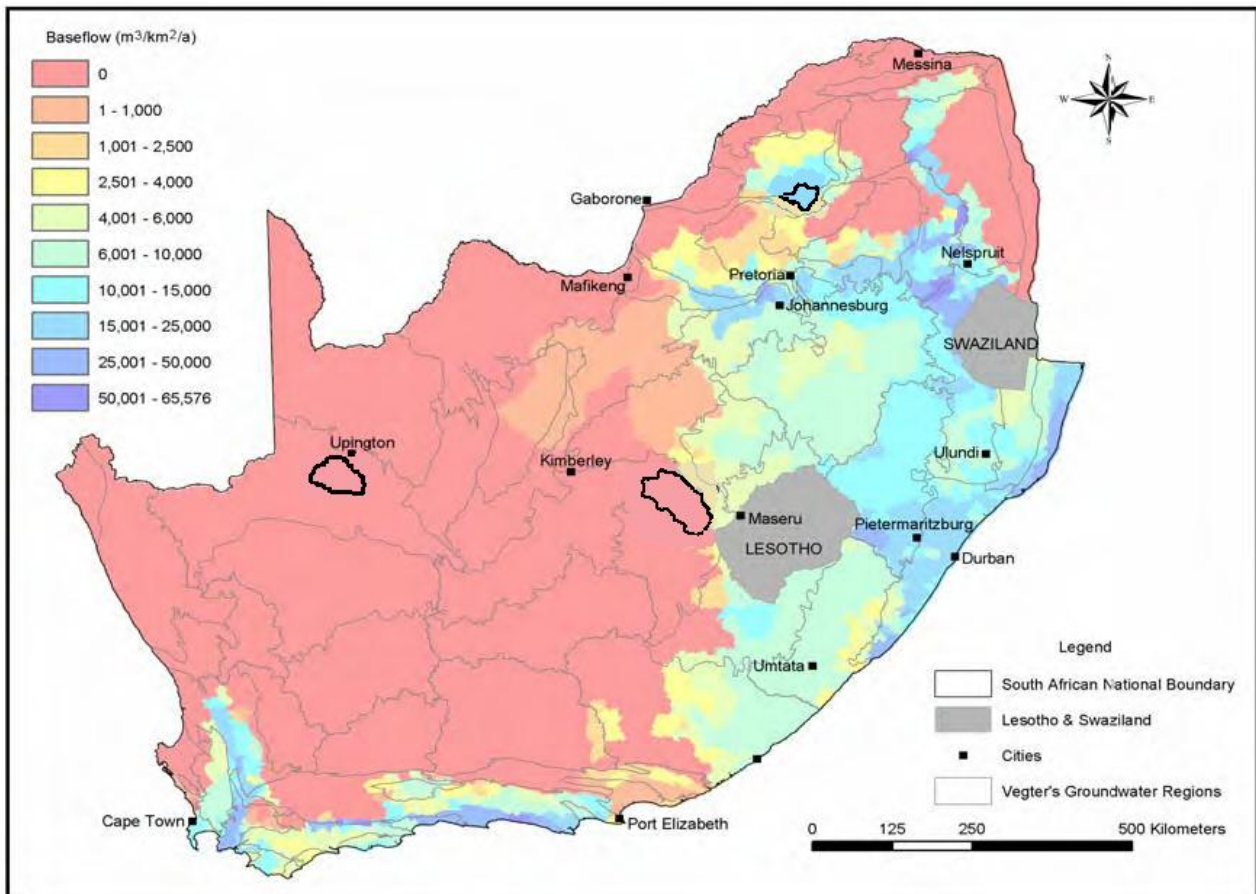


Figure 5-2 A map of South Africa indicating the GRA2 determined groundwater baseflow zones of South Africa, showing the location of the three pilot study areas (Modified from Department of Water Affairs, 2006b).

5.1. Pilot Study 1: Free State quaternary catchments C52A—C52H

5.1.1. Overview

The Water Research Commission (WRC) and the University of the Free State (UFS) have established a surface water – groundwater interaction test site on the Modder River just outside of Bloemfontein at the base of the Krugersdrift Dam, South Africa (Figure 5-1-1). River water and groundwater quality samples collected along the middle Modder River, from the Rustfontein Dam to the base of this test site formed part of a further investigation aimed at a better understanding of the surface water – groundwater interaction taking place on a larger scale.

The middle reach of the Modder River is situated within the Upper Orange Water Management Area of the Free State Province, South Africa. The Modder River is a perennial river (Welderufael and Woyessa, 2010) and is an important source of water for domestic, agricultural and industrial use to Bloemfontein, Botshabelo and Thaba N’chu areas (Seaman *et al.* 2003). The Middle Modder River covers quaternaries C52B - C52G (Figure 5-1-1), with the main tributaries being the Korannaspruit, Osspruit, Renosterspruit, Stinkhoutspruit and Doringspruit Rivers. The area has an

arid to semi-arid climate, characterised by long periods of low rainfall events in the winter months and intense thunderstorms in the summer, rainy season which promote surface runoff. The average annual rainfall is estimated at 567mm and the average annual evaporation is estimated at 1943mm, based on data from meteorological stations at the Rustfontein Dam and the Krugersdrift Dam. The average annual temperature ranges from 16°C to 26°C. The area is characterised by open flat plains with gradual hills, and koppies. The topography ranges from around 1300mamsl to 1700mamsl over most of the area, with a general decrease towards the north-west. There is an isolated area in the south east where topography rises to 2000m. The vegetation types in the area are Moist Cool Highveld Grasslands, Dry Sandy Highveld Grassland and Eastern mixed Nama Karoo (Department of Water Affairs GRDM, 2010). The warms database for the quaternaries in the study area indicate that the abstraction from surface water resources per annum is 39.7Mm³.

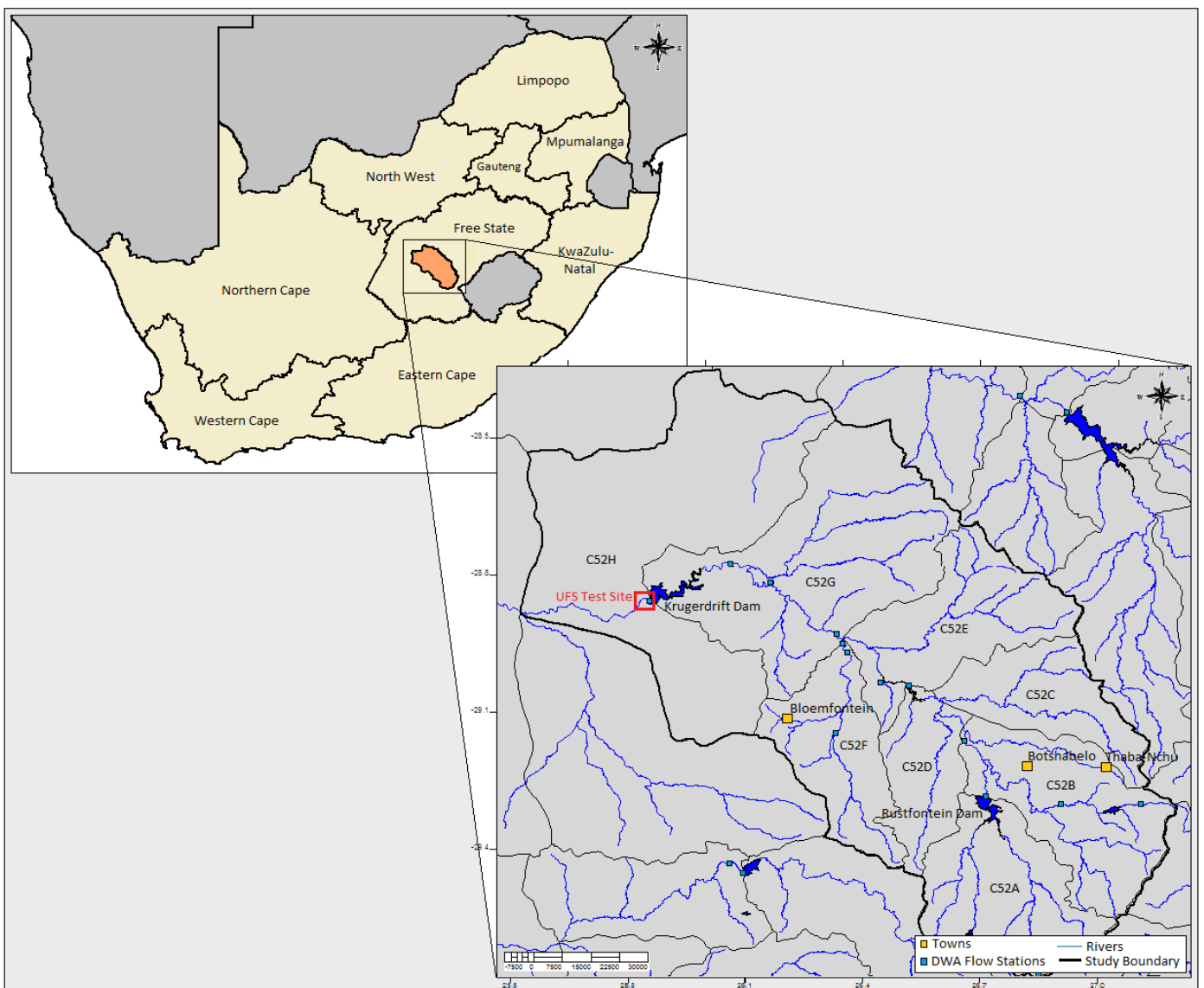


Figure 5-1-1 The location of the middle Modder River with the associated quaternaries and the UFS Surface water – groundwater interaction Test Site within South Africa.

5.1.2. Geology

The general outcrop geology of the middle Modder River area is comprised of Transvaal Supergroup rocks (Figure 5-1-2). The area is extensively intruded with dolerite dykes and sills. The Transvaal Supergroup rocks that are present in the study area form only a portion of the complete sequence, namely from the uppermost formation of the Eccca Group to the Elliot Formation of the Stormberg Group (Table 5-1-1). The overlying Drakensberg Group and underlying Dwyka Groups are not seen in the outcrop geology. The Volksrust Formation of the Eccca Group, the last of the group's sixteen formations, is a predominantly argillaceous layer that interfingers with the overlying Beaufort Group. The Volksrust Formation consists of silty shale with thin siltstone and sandstone lens. The overlying Adelaide Sub-group of the Beaufort Group consists of alternating layers of mudrock and sandstone, with sandstone becoming dominant towards the base. The Tarkastad Sub-group is characterised by a larger abundance of sandstone and red mudstone than in the Adelaide Sub-group. The boundary between these sub-groups is the only one in the Beaufort Group which can be traced throughout the Karoo Basin. The Molten Formation of the Stormberg Group consists of alternating layers of sandstone and grey mudstone. The overlying Elliot Formation is also an alternating sequence of red, green-grey mudrock and sandstone. The Elliot Formation is a typical "red bed" fluvial deposit (Johnson, Anhaeusser and Thomas, 2006).

Based on the outcrop geology of the area, a conceptual geological cross-section along the course of the middle Modder River (Rustfontein Dam to the Krugersdrift Dam) is created (Figure 5-1-3). The cross-section shows that the investigated section Modder River initially flows over the alternating sandstone and mudstone lithologies of the Adelaide Sub-group in the south-east where the topography is slightly higher, while further downstream the river flows over the lacustrine lithology of the Volksrust Fm.

The UFS surface water – groundwater interaction test site consists of 15 boreholes drilled near the base of the Krugersdrift Dam. Borehole logs from five of the test site boreholes were utilised to create a geological cross-section perpendicular to the river (Figure 5-1-4). The geological cross-section shows the conceptual understanding of the geology within 300 meters of the river. The lithology underlying the river is shale, presumably the Volksrust Shale of the Eccca Group. The surrounding river valley geology consists of this shale overlain by layers of calcrete, clay, sand and gravel.

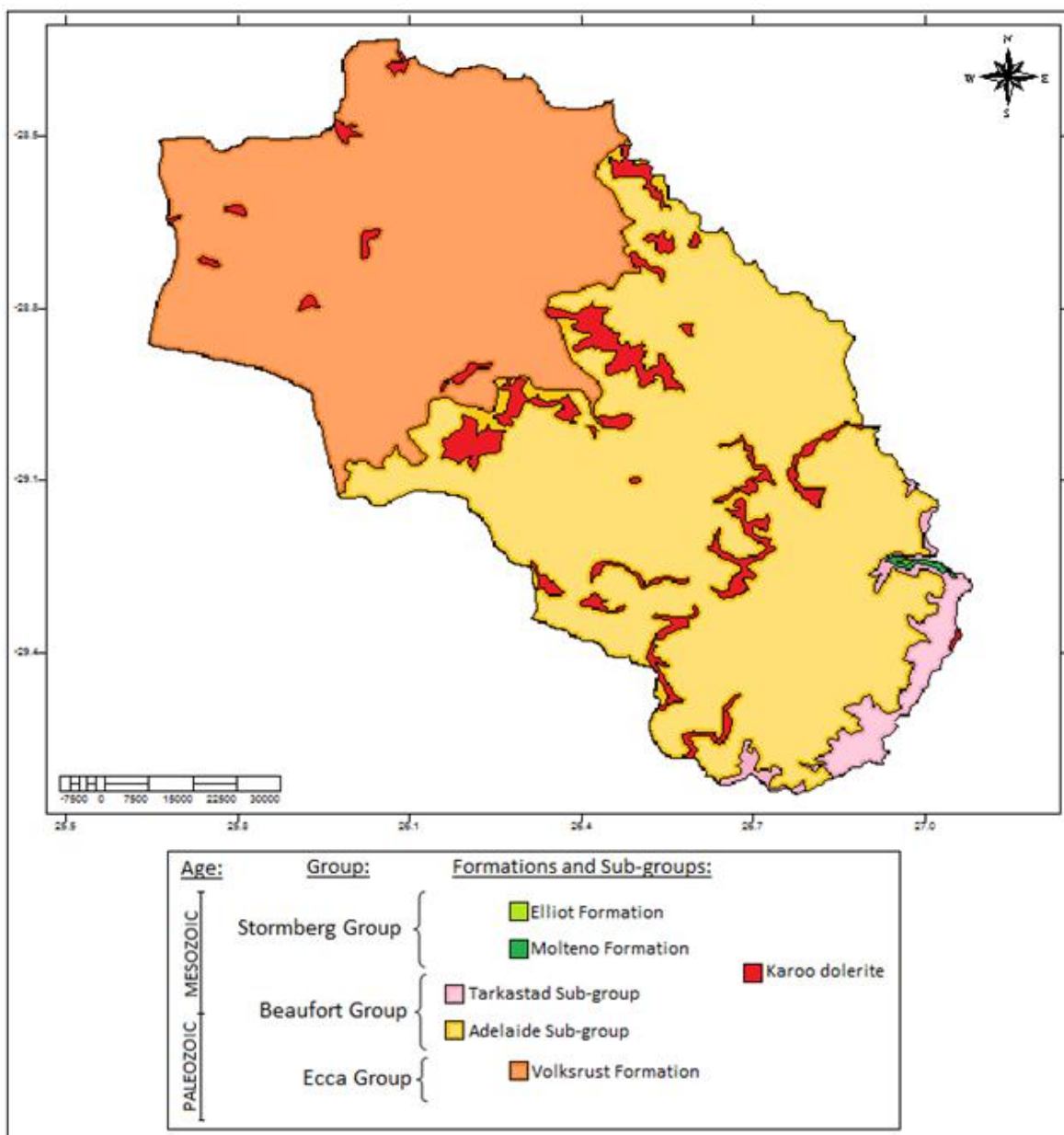


Figure 5-1-2 The outcrop geology of the C52 study area (based on the GRDM programme, 2010)

Table 5-1-1 The stratigraphic sequence present in the C52 study area

Era	Group	Sub-group	Formation
MESOZOIC	Drakensberg		
	Stormsberg		Clarens
			Elliot
			Molteno
PALEOZOIC	Beaufort	Tarkastad	
		Adelaide	
	Ecca		Volksrust

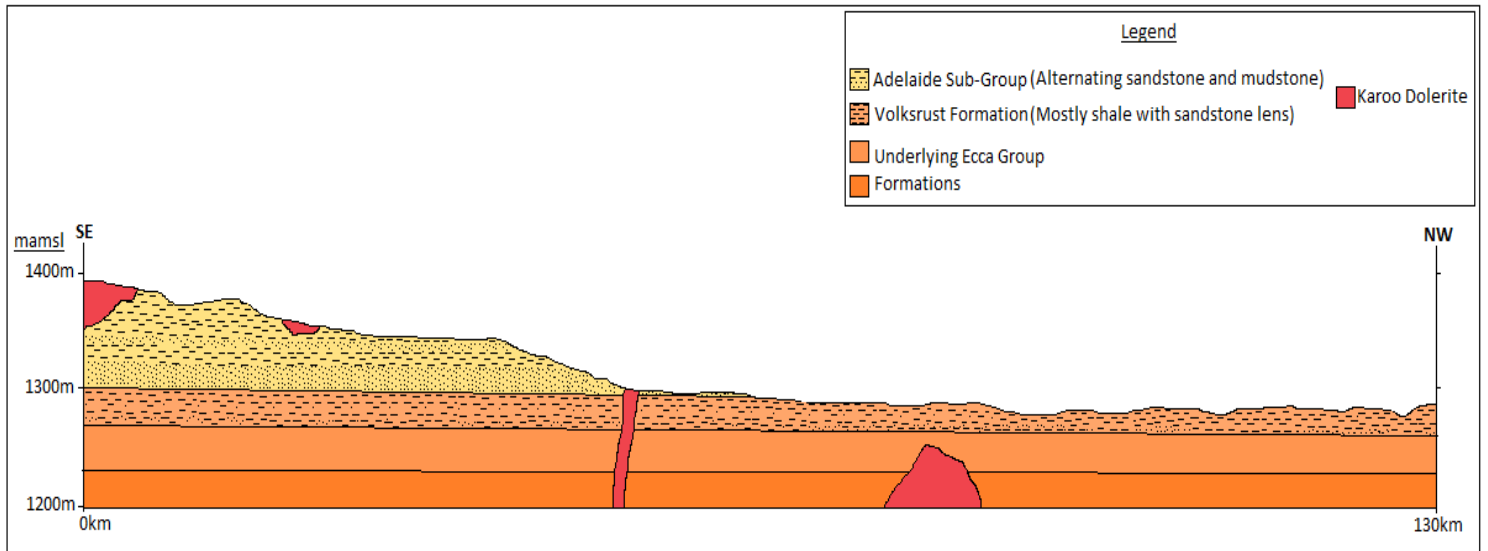


Figure 5-1-3 A conceptual geological cross-section along the Modder River from the Rustfontein Dam in the South-east to the Krugersdrift Dam in the north-west.

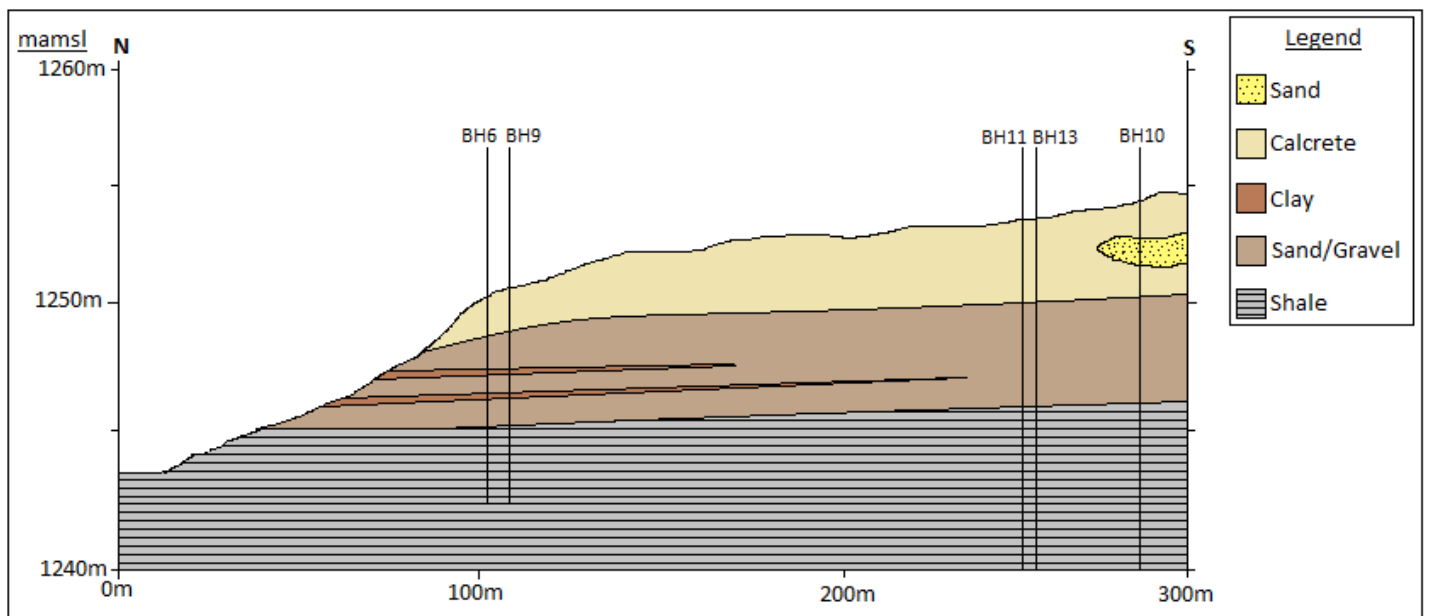


Figure 5-1-4 A conceptual geological cross-section perpendicular to the Modder River at the UFS Surface water – groundwater interaction site based on geological logs from BH10, BH13, BH11, BH9 and BH6.

5.1.3. Hydrogeology

The topography of the middle Modder River, spanning over quaternaries C52A to C52H, is higher towards the south-east with a localised high in the south-west (Figure 5-1-5). The topography gradually decreases towards the north-west, following the course of the Modder River towards the Krugersdrift Dam. The groundwater table tends to follow the topography in the area (Figure 5-1-6).

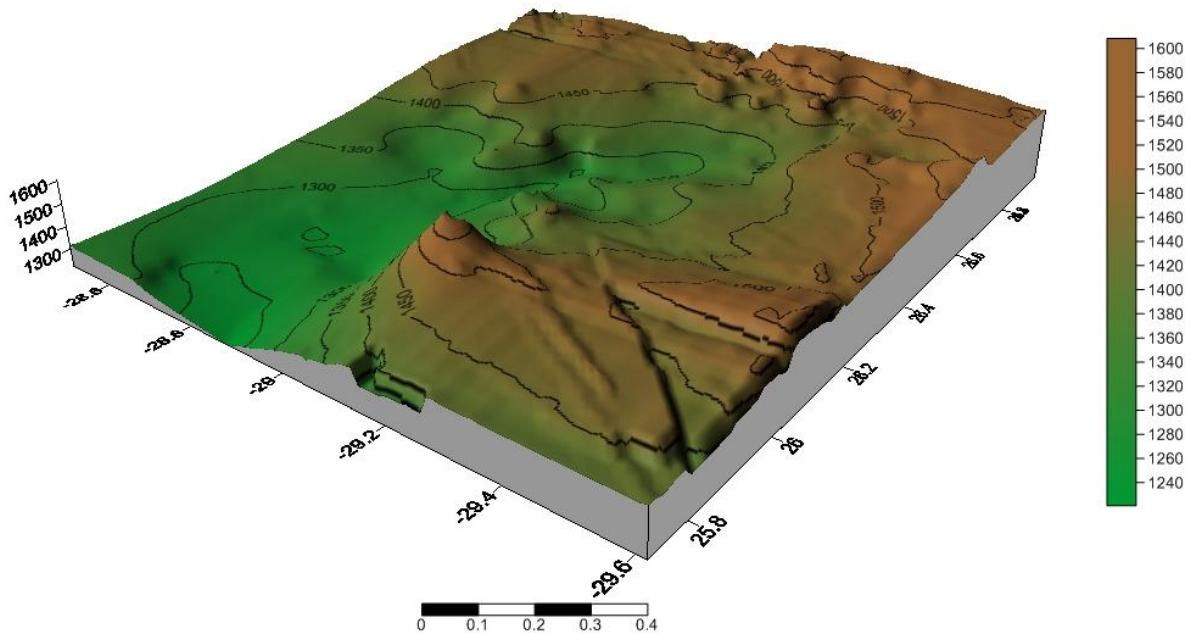


Figure 5-1-5 Generalised topography of the middle Modder River (quaternaries C52A – C52H).

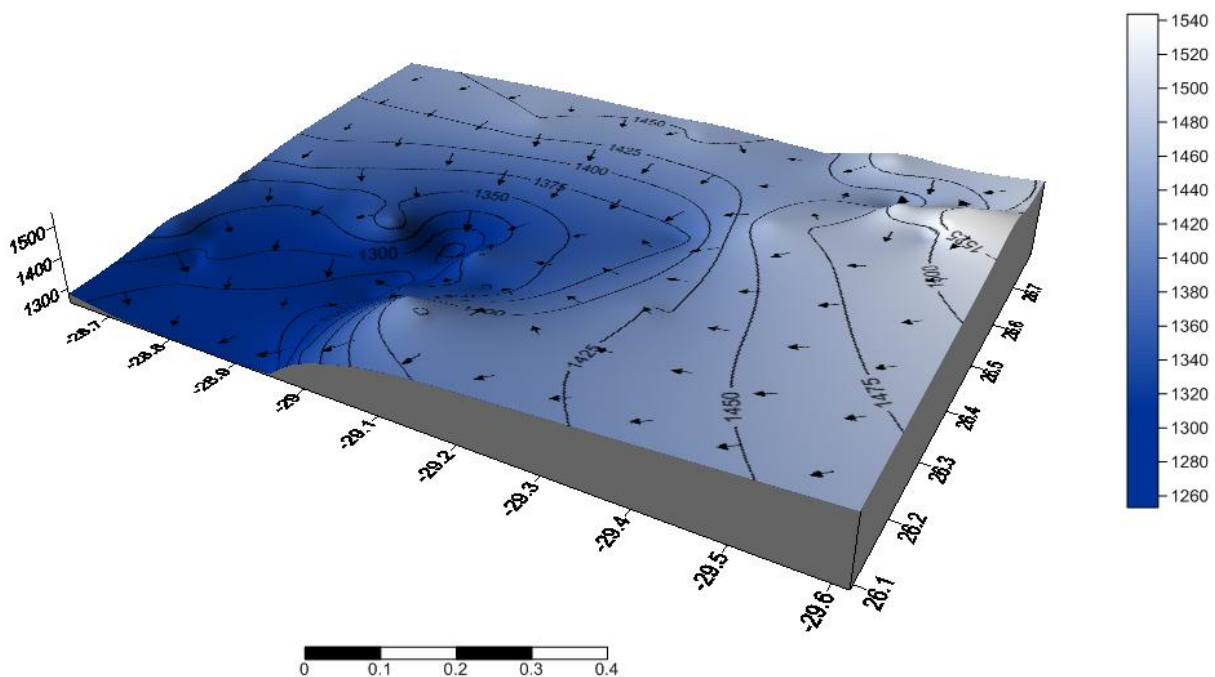


Figure 5-1-6 Generalised groundwater level within the middle Modder River course (1983-1990).

The Modder River flows over the Adelaide sub-group lithologies as well as the Volksrust shales (Figure 5-1-3). The Adelaide sub-group consists of alternating sandstone and mudstone layers. The sandstone layers are a potential source of groundwater to the river, but the mudstone layers would not be a high-yielding source of groundwater to the river due to its lower permeability. However, contact planes between the two layers and secondary features such as fractures could supply the river with groundwater through preferential pathways. The underlying Volksrust shale is an aquiclude which would only allow for slow diffusion of water, and movement along fractures and bedding planes.

The geological cross-section from geological borehole logs at the UFS surface water – groundwater interaction test site indicates that the river at this location is underlain by the impermeable shale of the Volksrust Fm (Figure 5-1-4). The contribution of groundwater to the river through this lithology would be minimal. Groundwater contributing to the river would then have to either be reaching the river through fractures and weathered areas of the shale, or from the overlying alluvial aquifer of calcrete, sand and gravel (Figure 5-1-4). Gomo (2011) first suggested this mechanism of groundwater baseflow to the Modder River along a seepage face between the overlying alluvial aquifer and the impermeable underlying shale. Gomo (2011) suggests that the groundwater discharge along the seepage face is derived from the local groundwater flow system, namely the shallow alluvial cover channel deposits. This mechanism of surface water – groundwater interaction implies that even though there is no direct hydraulic connection between the river and the underlying lithology, large amounts of groundwater can be contributing to the river flow (Gomo, 2011).

5.1.4. Mixing Cell Model (MCM)

The Mixing Cell Model (MCM) is applied on a site-specific and catchment scale. The site-specific scale application is performed using river and groundwater chemical analysis data from the UFS surface water – groundwater interaction test site. River water and groundwater chemical analysis data is available for samples collected on the 24/01/2011 and 05/08/2011, which allows the MCM to be run on two temporal scales. Figure 5-1-7 indicates the position of the 15 boreholes sampled within the test site. The catchment scale application is performed using river and groundwater chemical analysis data collected along the middle Modder River on 29/10/2012 – 31/10/2012 and 29/01/2013 – 01/02/2013, also allowing for two temporal scale applications. Figure 5-1-8 indicates the position of the river water and groundwater sample points for the January 2013 catchment scale application, and the active flowstations within the area.



Figure 5-1-7 Location of the 15 borehole samples at the UFS Surface water – groundwater interaction test site.

Application of the Mixing Cell Model to the quantification of groundwater – surface water interaction

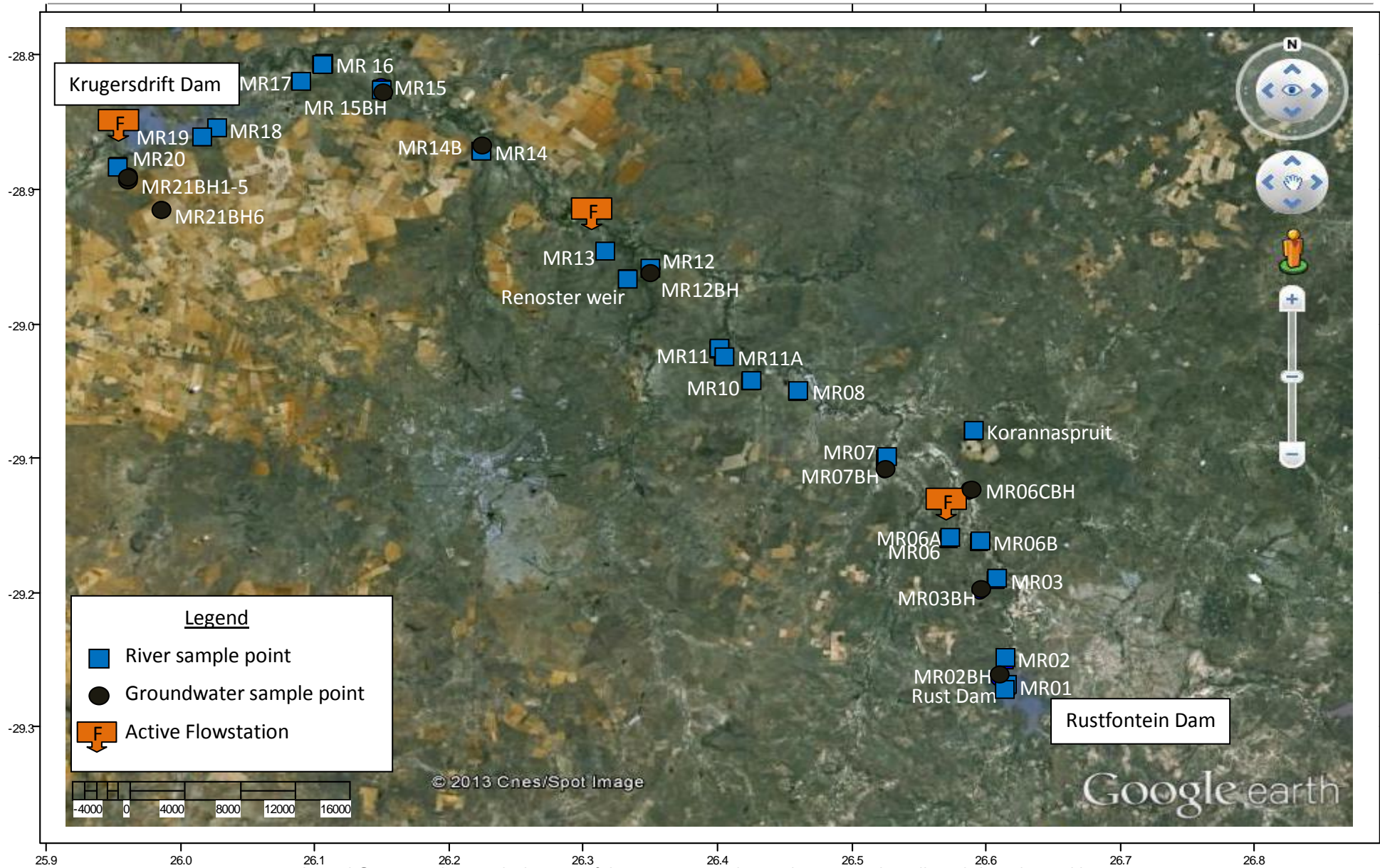


Figure 5-1-8 A Google Earth© Image indicating the location of the river water and groundwater samples collected along the Modder River in January 2013.

Site Specific Scale

5.1.5. Site Specific Scale MCM Conceptualisation and application

The various flows into and out of the river system at the UFS surface water – groundwater interaction test site are conceptualised into a box model, which forms the basis of the water balance equation for the MCM (Figure 5-1-9). The river course between the Krugersdrift Dam wall and the point of flow measurement is conceptualised as the River Cell. The MCM software requires a minimum of two mixing cells to be defined, thus a fictitious cell (o-River Cell) is created with exactly the same tracer concentrations as the River Cell, which is standard procedure for single cell applications of the model. The River Cell is defined by a set of tracer concentrations obtained from the average of two river samples taken within the defined study area (See Appendix B Table B-1). Inflow to the River Cell is defined by the average tracer concentration data from the flowstation C5H039 at Krugersdrift Dam. Groundwater sources to the River Cell are defined by chemical analysis data from each of the 15 boreholes within the study area, and the seepage source is defined by the water quality analysis data from the sampled seepage water. The river flow measurement taken downstream from the study area by Gomo (2011) is utilised in this model run as there is no flowstation at or near this point, other than the upstream Krugersdrift Dam flowstation. Abstraction and evaporation loss volumes from the section of the Modder River under investigation as well as direct rainfall volumes into the system are assumed to be negligible, as chemical data used in the model run covers a short time period and this omission ensures a mathematically simple model run. The representative EC and chloride (Cl) values for each flow represented in the flow box model are shown in Figure 5-1-10. Figure 5-1-10 indicates that the EC and Cl values, used to define the River Cell, are higher than the upstream river inflow source, which could indicate that the poorer quality groundwater could indeed be contributing to the stream.

The MCM was applied using the flow setup discussed in two model runs, each making use of a complete set of chemical data from two different time periods, namely one model run for data from January 2011 and one model run for data from August 2011. Three scenarios regarding tracers and weighting factors were performed for each time period, namely 1) all available tracer data is utilised and all tracers are assumed conservative, and thus assigned a weighting factors of 1 ($\omega = 1$), 2) tracers showing a high chemical mass balance error are assigned a lower weighting factor ($\omega = 0.3$), and 3) lower weighting factors are assigned as in scenario 2 and additionally, the tracer showing the highest error is omitted (Table 5-1-2). The outflow from the January 2011 run is $6.8 \text{ Mm}^3/\text{a}$ measured by Gomo (2011), while the outflow from the August

2011 run is determined from the upstream flowstation C5H039, because no other direct measurements of flow are available. The chemical data for each time period is given in Appendix B Table B-1 and B-2.

Table 5-1-2 The weighting factor assigned to each tracer for scenarios 1 – 3.

Weighting Factor Scenario	Weighting Factor (ω)										
	pH	EC	Ca	Mg	Na	K	TAL	F	Cl	SO4	Si
Scenario 1	1	1	1	1	1	1	1	1	1	1	1
Scenario 2	1	1	1	1	1	0.3	1	0.3	0.3	0.3	1
Scenario 3	1	1	1	1	1	0.3	1	0	0.3	0.3	1

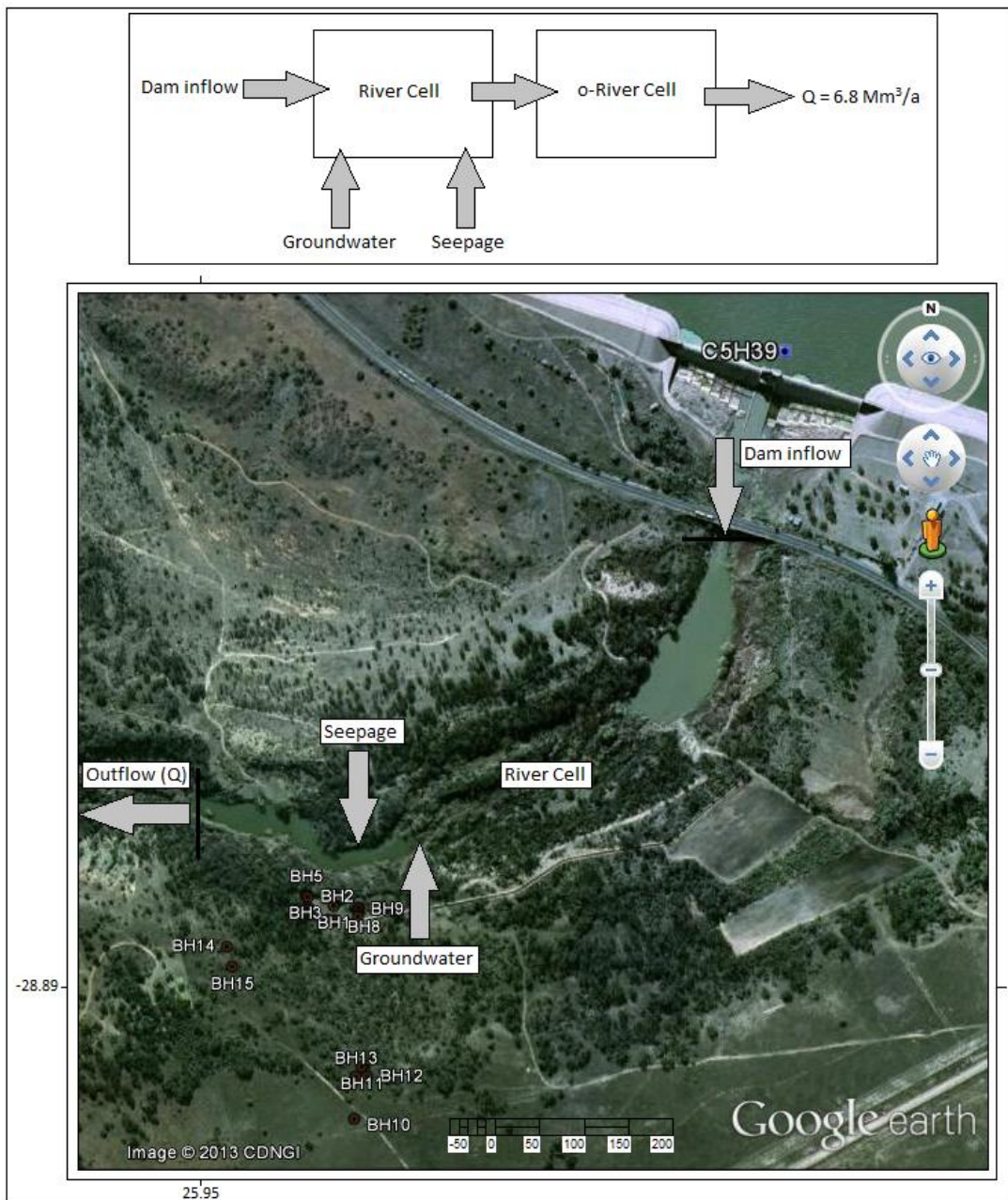


Figure 5-1-9 The box model conceptualisation of the various flows modelled at the UFS Surface water – groundwater interaction test site and a Google Earth® image showing the conceptualisation on a real scale indicating the position of the flowstation (C5H039) and sampled boreholes (BH1-15).

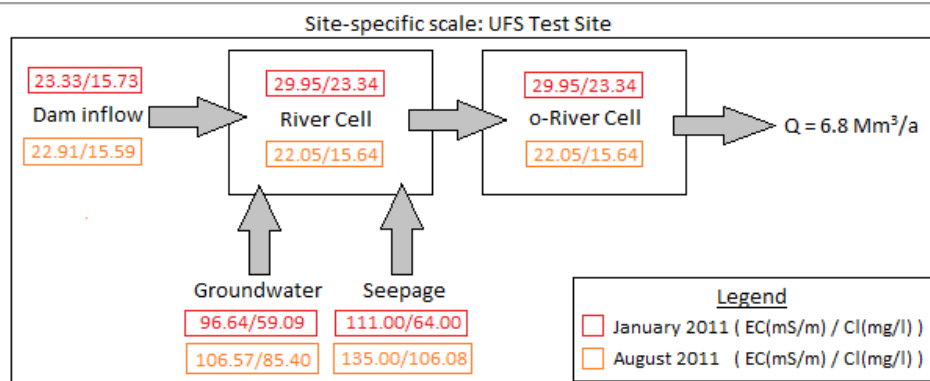


Figure 5-1-10 The representative EC and chloride (Cl) values for the box model conceptualisation of the various flows modelled at the UFS Surface water – groundwater interaction test site. Concentrations indicate that groundwater could be contributing to the streamflow because the EC-value and chloride concentration increase downstream.

5.1.6. Site Specific Scale MCM Results

The MCM run for January 2011, using scenario 1 weighting factors, indicated a groundwater contribution to baseflow from the sampled boreholes BH1-9 of $0.03\text{Mm}^3/\text{a}$, while the contribution from sampled boreholes BH10-15 was $0.84\text{Mm}^3/\text{a}$ (Table 5-1-3). No inflow from the seepage source was found. The inflow from the Dam to the River Cell was determined to be $5.23\text{Mm}^3/\text{a}$, while the outflow from the River Cell to the fictitious o-River Cell was determined at $7.32\text{Mm}^3/\text{a}$. The MCM run for January 2011, using scenario 2 weighting factors, indicated a groundwater contribution to baseflow from the sampled boreholes BH1-9 of $0.57\text{Mm}^3/\text{a}$, while the contribution from sampled boreholes BH10-15 was $0.00\text{Mm}^3/\text{a}$ (Table 5-1-3). No inflow from the seepage source was found. The inflow from the Dam to the River Cell was determined to be $6.51\text{Mm}^3/\text{a}$, while the outflow from the River Cell to the fictitious o-River Cell was determined at $7.68\text{Mm}^3/\text{a}$. The MCM run for January 2011, using scenario 3 weighting factors, indicated a groundwater contribution to baseflow from the sampled boreholes BH1-9 of $0.47\text{Mm}^3/\text{a}$, while the contribution from sampled boreholes BH10-15 was $0.00\text{Mm}^3/\text{a}$ (Table 5-1-3). No inflow from the seepage source was found. The inflow from the Dam to the River Cell was determined to be $6.89\text{Mm}^3/\text{a}$, while the outflow from the River Cell to the fictitious o-River Cell was determined at $7.74\text{Mm}^3/\text{a}$.

Table 5-1-3 Summary of the MCM output for the UFS Test Site (January 2011) for each of the weighting factor scenarios

UFS Test site (January 2011)			
Name of inflow	Rate of inflow (Mm^3/a)		
	Scenario 1	Scenario 2	Scenario 3
C5H039	5.23	6.51	6.89
BH1-9	0.03	0.57	0.47
BH10-15	0.84	0.00	0.00
Seep	0.00	0.00	0.00

The MCM run for August 2011, using scenario 1 weighting factors, indicated a groundwater contribution to baseflow from the sampled boreholes BH1-9 of $0.03\text{Mm}^3/\text{a}$, while the contribution from sampled boreholes BH10-15 was $1.6\text{Mm}^3/\text{a}$ (Table 5-1-4). Inflow to the River Cell from the seepage source along the river was estimated at $0.7\text{Mm}^3/\text{a}$. The inflow from the Dam to the River Cell was determined to be $4.45\text{Mm}^3/\text{a}$, while the outflow from the River Cell to the fictitious o-River Cell was determined at $19.13\text{Mm}^3/\text{a}$. The MCM run for August 2011, using scenario 2 weighting factors, indicated a groundwater contribution to baseflow from the sampled boreholes BH1-9 of $0.37\text{Mm}^3/\text{a}$, while the contribution from sampled boreholes BH10-15 was $0.00\text{Mm}^3/\text{a}$ (Table 5-1-4). Inflow to the River Cell from the seepage source along the river was estimated at $0.1\text{Mm}^3/\text{a}$. The inflow from the Dam to the River Cell was determined to be $12.77\text{Mm}^3/\text{a}$, while the outflow from the River Cell to the fictitious o-River Cell was determined at $20.30\text{Mm}^3/\text{a}$. The MCM run for August 2011, using scenario 3 weighting factors indicated a groundwater contribution to baseflow from the sampled boreholes BH1-9 and boreholes BH10-15 of $0.00\text{Mm}^3/\text{a}$ (Table 5-1-4). Inflow to the River Cell from the seepage source along the river was estimated at $0.35\text{Mm}^3/\text{a}$. The inflow from the Dam to the River Cell was determined to be $15.29\text{Mm}^3/\text{a}$, while the outflow from the River Cell to the fictitious o-River Cell was determined to be $21.43\text{Mm}^3/\text{a}$.

Table 5-1-4 Summary of the MCM output for the UFS Test Site (August 2011) for each of the weighting factor scenarios

UFS Test site (August 2011)			
Name of inflow	Rate of inflow (Mm^3/a)		
	Scenario 1	Scenario 2	Scenario 3
C5H039	4.45	12.77	15.29
BH1-9	0.03	0.37	0
BH10-15	1.6	0	0
Seep	0.7	0.1	0.35

The results from the two temporal scale applications, utilizing scenario 2 weighting factors are graphically represented in Figure 5-1-11 and expressed as percentages of the assigned total outflow. The January 2011 MCM application determined the inflow from the Dam to the conceptualised River Cell to be 84.7% of the total outflow from the cell. The groundwater contribution to the stream was estimated at 7.4% of the total flow, while the seepage to the river was estimated at 0.0% of the total flow. The August 2011 MCM application determined the inflow from the Dam to the conceptualised River Cell to be 62.9% of the total outflow from the cell. The groundwater contribution to stream was estimated at 1.8% of the total flow, while the seepage to the river was estimated at 0.5% of the total flow.

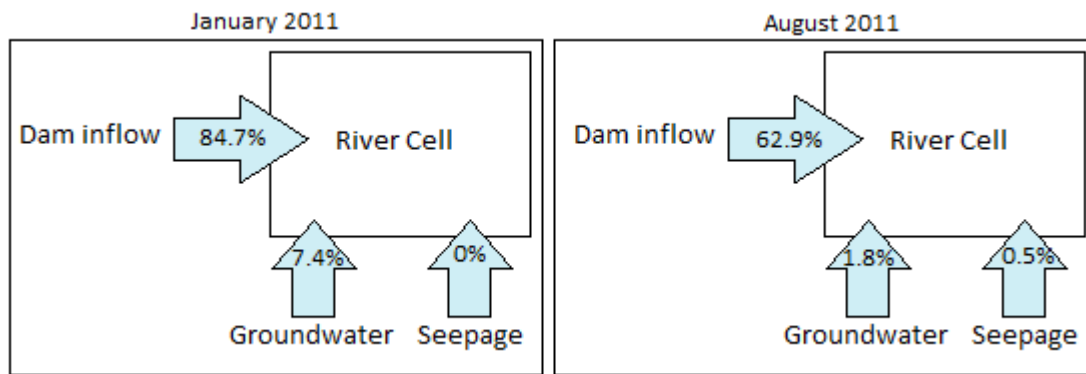


Figure 5-1-11 The unknown fluxes determined by the MCM for the January and August 2011 model runs, making use of the scenario 2 weighting factors, expressed as a percentage of the total flow volume.

5.1.7. Site Specific Scale Discussion and Comparison

The water balance percentage errors for each time period and the associated weighting factor scenario are shown in Table 5-1-6. These errors indicate that there is more water flowing out of the system than flowing in from the identified sources, which could be caused by unidentified runoff, rainfall or river inflow sources to the system. The percentage flow diagram (Figure 5-1-11) highlights this difference where in the January 2011 model run 7.9% of the flow to the system is unaccounted for, while in the August 2011 model run 34.8% of the flow is unaccounted for. These unaccounted flows result in the error seen in the water balance. The chemical mass balance percentage errors are shown in Table 5-1-5a and b. The water balance errors for the MCM run for January 2011 are much lower than the errors in the August 2011 run. The large errors associated with the August 2011 run could be attributed to the outflow volume that is used, because a flow volume from the C5H039 flowstation was utilised as no direct flow measurement was taken. The water balance error decreases when weighting factors are implemented in both instances and then even further when the tracer F (Fluoride) is omitted (Table 5-1-6). The resulting effect on the estimation of the groundwater contribution to the system in response to the three different scenarios is clear, with the change from scenario 1 to scenario 2 resulting in a drastic decrease in percentage error and overall groundwater contribution to the river. A shift in the main source of groundwater baseflow from BH10-15 to BH1-9 in the January and August model runs are (Table 5-1-3 and Table 5-1-4), and a decrease in the amount of seepage inflow for the August model run (Table 5-1-4) results from the incorporation of weighting factors. The change from scenario 2 to scenario 3 results in a decrease in groundwater contribution from BH1-9 in the January run, whereas the result in the August run is no groundwater contributes to the river and an increase in seepage inflow is seen. The chemical mass balance percentage error drastically decreases from scenario 1 to scenario 2, but slightly increases again from scenario 2 to scenario 3, as seen in the average percentage

error for each scenario (Table 5-1-5a and b). Considering these changes in percentage error for the balance equations, it seems that scenario 2 would be a reasonable selection for the final model run, because it allows for a lower water balance and chemical mass balance percentage error.

Table 5-1-5 Chemical mass balance percentage error for each tracer for a) January 2011 for each scenario, and b) August 2011 for each scenario.

a) Ion balance (January 2011)				b) Ion balance (August 2011)			
Tracer	Chemical Mass Balance Error (%)			Tracer	Chemical Mass Balance Error (%)		
	Scenario 1	Scenario 2	Scenario 3		Scenario 1	Scenario 2	Scenario 3
pH	-13.80%	0.30%	4.50%	pH	-67.70%	-35.60%	-23.90%
EC	-15.00%	-5.80%	-9.70%	EC	-47.50%	-32.30%	-27.50%
Ca	-12.40%	-10.30%	-4.30%	Ca	-28.00%	-1.30%	19.60%
Mg	16.00%	11.80%	12.40%	Mg	21.20%	23.90%	28.30%
Na	-21.30%	-12.50%	-10.90%	Na	6.80%	-29.20%	-11.00%
K	-42.80%	-33.30%	-30.60%	K	-52.60%	-38.50%	-12.20%
TAL	-4.00%	1.20%	4.30%	TAL	-37.00%	-26.00%	5.00%
F	46.70%	77.80%		F	34.40%	165.50%	
Cl	-30.90%	-24.50%	-23.00%	Cl	-46.20%	-34.40%	-13.70%
NO3	-10.50%	0.20%	3.50%	NO3	-40.00%	-29.30%	-12.50%
SO4	-38.00%	-28.40%	-26.20%	SO4	-23.30%	-21.40%	-20.20%
Average	-11%	-2%	-8%	Average	-25%	-5%	-7%

Table 5-1-6 Water balance percentage error for each scenario for each time period

Water Balance Error (%)			
Time period	Scenario 1	Scenario 2	Scenario 3
January 2011	22%	9%	6%
August 2011	70%	41%	30%

The final groundwater contribution to streamflow in the study area is estimated by the MCM at 0.57Mm³/a in January 2011 and 0.37Mm³/a in August 2011, with an additional seepage inflow of 0.1Mm³/a in August 2011. The rainfall difference between the two months is substantial (Figure 5-1-12), yet the groundwater contribution to baseflow seems to remain reasonable stable. This constant groundwater contribution to the river flow might be an indication that there is a limiting factor which inhibits the increase of groundwater baseflow with an increase in rainfall, such as a limited or conditional hydraulic connection between the river and groundwater system. An unconventional surface water – groundwater connection is plausible in light of the impermeable shale underlying the river bed. Another component to consider is the seepage inflow to the river sourced from the overlying alluvial aquifer. The MCM run for the

high rainfall time period January 2011 found no seepage inflow while the low rainfall period August 2011 model run found a seepage inflow of $0.1\text{Mm}^3/\text{a}$, which is opposite to what would be expected. However, this could be explained by high evapotranspiration rates and high borehole abstraction in the peak of the summer season resulting in little or no groundwater reaching the river from the overlying alluvial aquifer comprising the local groundwater system. The other explanation is that the MCM is essentially a numerical model, which could easily create a fictitious inflow from this source. The three weighting factor scenarios also highlight the inconsistency with regards to input data variation.

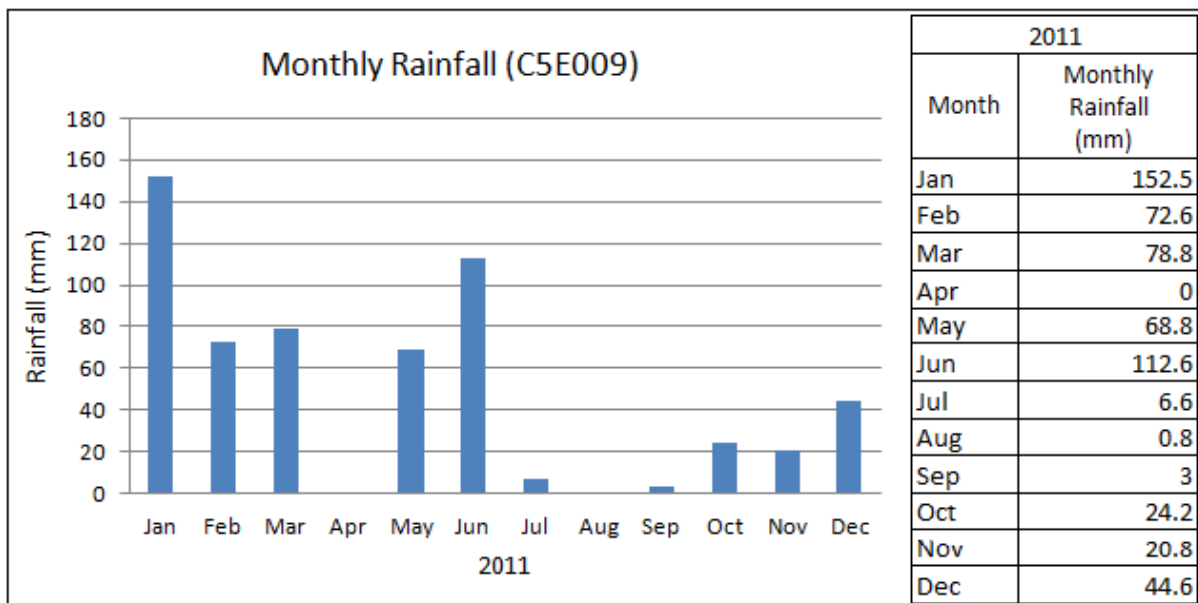


Figure 5-1-12 Rainfall in millimetres per month over the year 2011 in the UFS Test Site area from the meteorological station C5E009 at the Krugersdrift Dam.

The groundwater contribution estimates from the MCM are compared to four other methods in Table 5-1-7. The Pitman, Sami and Hughes methods all estimate the groundwater baseflow volume to be $0\text{Mm}^3/\text{a}$, while the tracer and MCM methods estimate the groundwater baseflow at $0.61\text{Mm}^3/\text{a}$ and $0.57\text{Mm}^3/\text{a}$, respectively. The average volume from a number of Tracer method application using different tracers results in a much higher estimate than when only EC is used. In terms of the methods making use of chemical data, namely the tracer and MCM methods, the MCM provides a more acceptable groundwater baseflow estimate, especially if one attempts to use more than one constituent in the tracer method. Comparing the MCM to the Pitman, Sami and Hughes methods, there is a large difference, where the conventional methods found no groundwater contribution to the river baseflow in the area, while the MCM did find that groundwater contributes to the river in small quantities. In this case, the MCM method seems to provide the most reasonable estimate of groundwater baseflow because of

the fact that estimating no groundwater baseflow might be an under-estimate. This is further substantiated when considering the results from the work performed by Gomo (2011) on the test site, where the river was found to be a gaining river and groundwater contribution to the river estimated at 4% of the total flow in the river. Gomo (2011) has reported on farmers in the area confirming seepage into the river from the alluvial aquifer along the river.

Table 5-1-7 Groundwater contribution estimates from the Pitman, Sami, and Hughes methods for the quaternary C52H & tracer and MCM methods for a) January 2011 and b) August 2011

a)

Quat	Pitman uncalibrated (Mm ³ /a)	Sami uncalibrated (Mm ³ /a)	Hughes uncalibrated (Mm ³ /a)	MCM (Mm ³ /a)	Tracer Method (Mm ³ /a)	
					EC	Average
C52H	0.00	0.00	0.00	0.57	0.61	1.96

b)

Quat	Pitman uncalibrated (Mm ³ /a)	Sami uncalibrated (Mm ³ /a)	Hughes uncalibrated (Mm ³ /a)	MCM (Mm ³ /a)	Tracer Method (Mm ³ /a)	
					EC	Average
C52H	0.00	0.00	0.00	0.37	0.57	6.72

Catchment scale

5.1.8. Catchment Scale MCM Conceptualisation and application

The catchment scale study area covers approximately 130km of the middle Modder River, stretching from the Rustfontein Dam to the Krugersdrift Dam, passing through the Mockes and Mazelspoort Dams. Nineteen river water samples and eleven groundwater borehole samples were collected in October 2012. Twenty three river water samples (three additional tributary samples and one additional Modder River sample) and thirteen groundwater borehole samples were collected in January 2013. The study area has been divided into three sections based on active flowstations (Figure 5-1-8), to allow for the quantification of groundwater baseflow within each section because the MCM requires a defined outflow from the modelled system to quantify the unknown inflows. Section 1 is defined from the Rustfontein Dam to the flowstation C5H003, Section 2 is defined from flowstation C5H003 to flowstation C5H053 and Section 3 is defined from flowstation C5H053 to the flowstation C5H039 at Krugersdrift Dam (Figure 5-1-8). A MCM run for each section is performed for each sample collection time period, namely October 2012 and January 2013.

The various flows into and out of the river system within each section are conceptualised in a box model, which forms the basis of the water balance equation for the MCM. The box model conceptualisation for Section 1 is shown in Figure 5-1-13. The system is represented as a two cell box model, with the river samples MR03 and MR06 defining the cells. Upstream river inflow to the MR03 Cell is defined by the water quality analysis results of the upstream river sample MR02 and the groundwater inflow is defined by the borehole samples, MR03BH and MR02BH. Upstream river inflow to the downstream cell (MR06 Cell) is sourced from the MR03 Cell and inflow from the Sepane tributary, represented by sample MR06B. Groundwater sources to the MR06 Cell are defined by the chemical analysis data from borehole sample points, MR06CBH and MR07BH. The flow volume recorded at the flowstation, C5H003 is used to define the downstream outflow from the system. The outflow volumes recorded at the flowstation for the month of October 2012 and January 2013 are extrapolated to yearly volumes of $6.24\text{Mm}^3/\text{a}$ and $16.05\text{Mm}^3/\text{a}$, respectively. The representative EC and chloride (Cl) values for each flow represented in the flow box model are shown in Figure 5-1-17. Figure 5-1-17 indicates both EC and Cl values increase downstream in October 2012, which could indicate that the poorer quality groundwater could be contributing to the stream. Chemical data collected in January 2013 shows an opposite trend in terms of EC as it decreases downstream, but Cl increases. The

lower EC values in January 2013 could be attributed to the increase rainfall during this period. Groundwater quality is shown to remain fairly stable between the two time periods. Data not collected in October 2012 for MR06CBH, MR07BH and MR06B prevent a complete model run.

Figure 5-1-14 is the box model conceptualisation of the flow system in Section 2 as a two cell box model, defined by river water quality samples MR12 and MR13. Upstream river inflow to the MR12 Cell is defined by the water quality analysis results of the upstream river sample MR11 and the groundwater inflow is defined by the borehole sample MR12BH. Upstream river inflow to the downstream cell (MR13 Cell) is sourced from the MR12 Cell and the groundwater inflow is defined by water quality data from the borehole sample MR12BH. The flow volume recorded at the flowstation, C5H053 is used to define the downstream outflow from the system. Outflow volume data at the flowstation C5H053 is $19.05\text{Mm}^3/\text{a}$ based on the monthly outflow data from October 2012. There is no monthly flow data for January 2013 at this flowstation. The representative EC and chloride (Cl) values for each flow represented in the flow box model are shown in Figure 5-1-18. Figure 5-1-18 shows a trend in EC and Cl values to increase downstream as well as the groundwater quality to be of a much poorer quality, which could indicate that groundwater is indeed contributing to the stream. Groundwater quality is once again seen to be fairly constant between the two time periods.

The box model conceptualisation for Section 3 is shown in Figure 5-1-15. The system is represented as a single cell box model, with the river cell defined by water quality data from the flowstation point at the Krugersdrift Dam, C5H039. The MCM requires at a minimum of two cells to be defined, thus a fictitious cell (o-C5H039 Cell) is created with exactly the same tracer concentrations as the river cell, C5H039 Cell. Inflow is defined by the average tracer concentration set of river samples MR18-MR20, and the groundwater inflow is defined by the average tracer concentration set of borehole samples MR21BH1-BH6. The box model for the January 2013 model run is slightly different from the one shown in Figure 5-1-19 because chemical data from the flowstation C5H039 sampling point is not available for this time period and thus follows a setup similar to the box flow model in Figure 5-1-16. The flow volume recorded at the flowstation, C5H039 is used to define the downstream outflow from the system. Outflow volume data at the flowstation C5H039 is $114.12\text{Mm}^3/\text{a}$ based on the monthly outflow data from October 2012. There is no monthly flow data for January 2013 at this flowstation. The representative EC and chloride (Cl) values for each flow represented in the flow box model are shown in Figure 5-1-19. The October 2012 setup shows a trend of decreasing EC and Cl values

downstream. This trend could indicate that no groundwater is contributing to the stream, but the water damming up in the Krugersdrift Dam could also be the cause for the reverse trend seen here. The January 2013 setup shows an expected trend of EC and Cl values increasing downstream. The shift in trend could be attributed to the different box model setup for this time period, where the January 2013 setup does not make use of chemical data from the C5H039 flowstation. Groundwater quality is found to be fairly constant between the two time periods.

The middle Modder River study area is additionally conceptualised as one continuous flow system to show a large scale application of the MCM and the functionality in ungauged systems (Figure 5-1-16). The continuous model run makes use of the January 2013 dataset. The area between the river samples MR15 and MR16 is insufficiently defined to allow for one continuous model, resulting in these two sections being modelled individually. Each sample point along the course of the sampled river areas is conceptualised as a river cell, optimising the number of mixing cells, to minimize the infringement on the basic assumption that total mixing occurs within each cell. Groundwater inflows are defined by borehole samples taken within the cell's vicinity.

The MCM was applied using the discussed flow system setups for Sections 1 to 3, where for each Section, two model runs were performed making use of a complete set of tracer concentration data from the two different time periods, namely one model run for data from October 2012 and one model run for data from January 2013. The water quality dataset for each section and each time period model run therein is shown in Appendix C Table C-1—C-6. Two different scenarios regarding weighting factors for individual tracers are performed also for each individual model run. Scenario 1 utilises all available tracer data and all weighting factors are assigned a value of 1, where scenario 2 assigns a lower weighting factor to tracers that show a high chemical mass balance error. The weighting factors assigned to the individual constituents for scenario are shown in Table 5-1-8. The ungauged continuous flow model setup was also run for each of the weighting factor scenarios. The water quality data used is shown in Appendix C Table C-7.

Table 5-1-8 Weighting factors assigned to each tracer for scenarios 1 and 2.

Weighting Factor Scenario	Tracer																			
	pH	EC	Ca	Mg	Na	K	MAIk	F	Cl	Br	SO4	TDS	Al	Fe	Mn	Si	Ba	Cu	Se	Zn
Scenario 1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1
Scenario 2	1	1	1	0.3	1	0.3	1	0.3	0.3	1	0.3	1	0.3	0.3	0.3	0.3	0.3	0.3	1	1

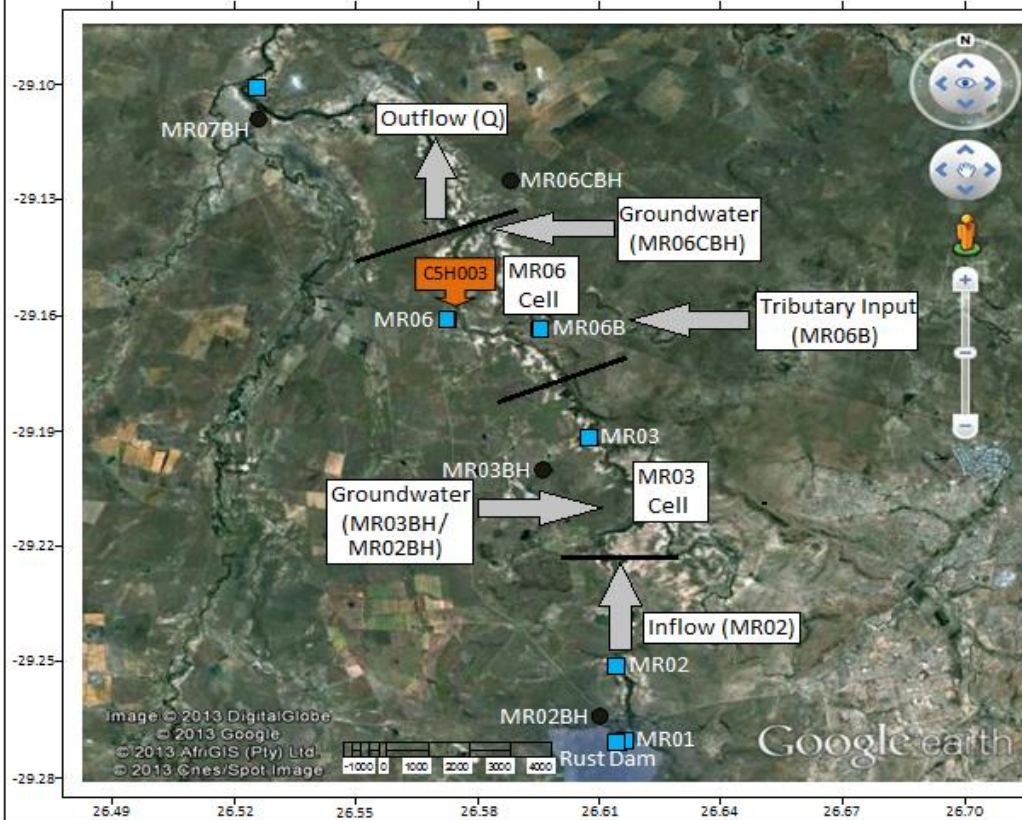
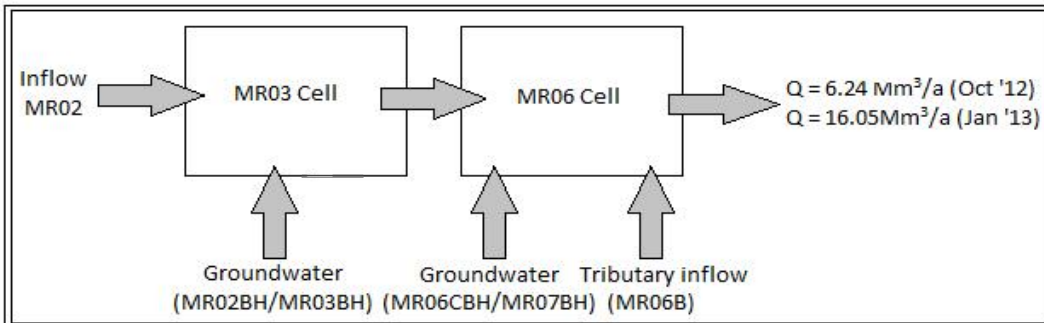


Figure 5-1-13 The box model conceptualisation of the various flows modelled for Section 1 and a Google Earth® image showing the conceptualisation on a real scale, indicating flowstation C5H003 and the various sample points.

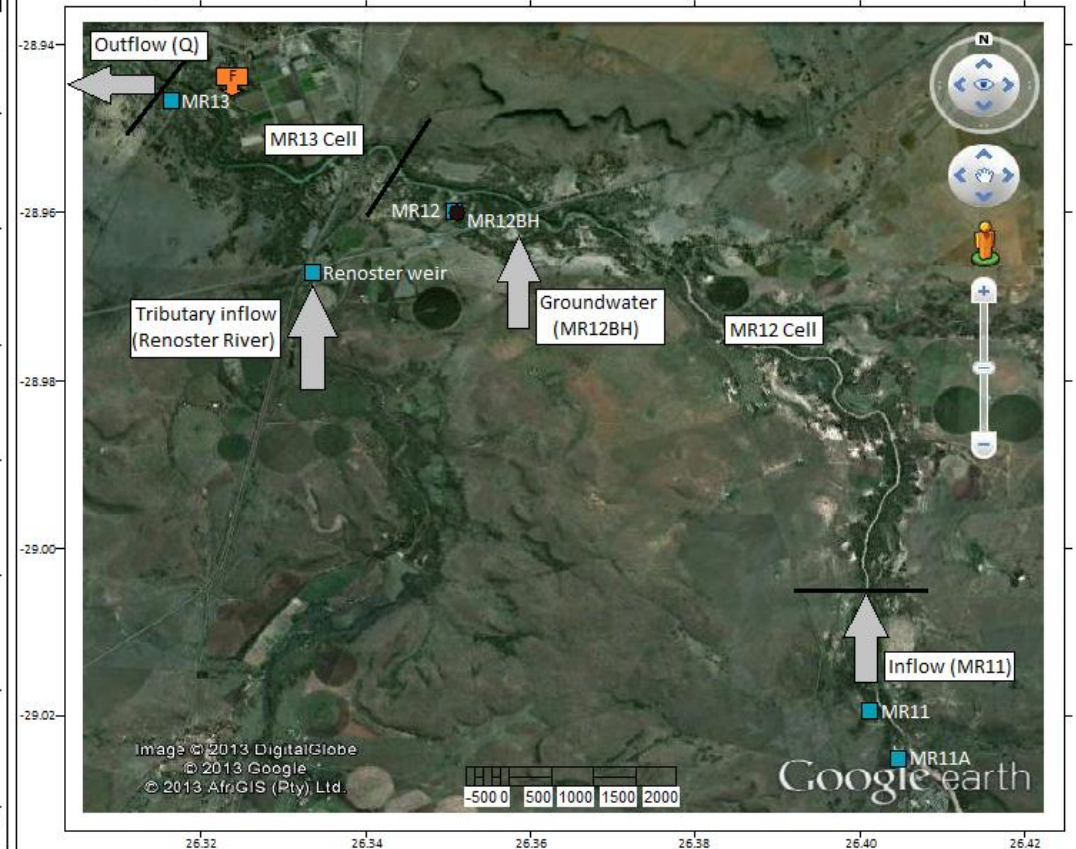
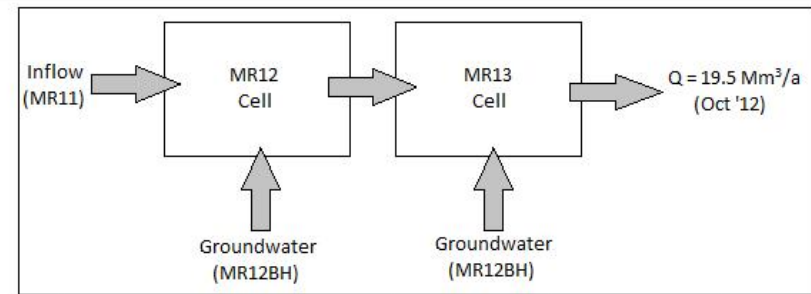


Figure 5-1-14 The box model conceptualisation of the various flows modelled for Section 2 and a Google Earth® image showing the conceptualisation on a real scale, indicating flowstation C5H053 and the various sample points.

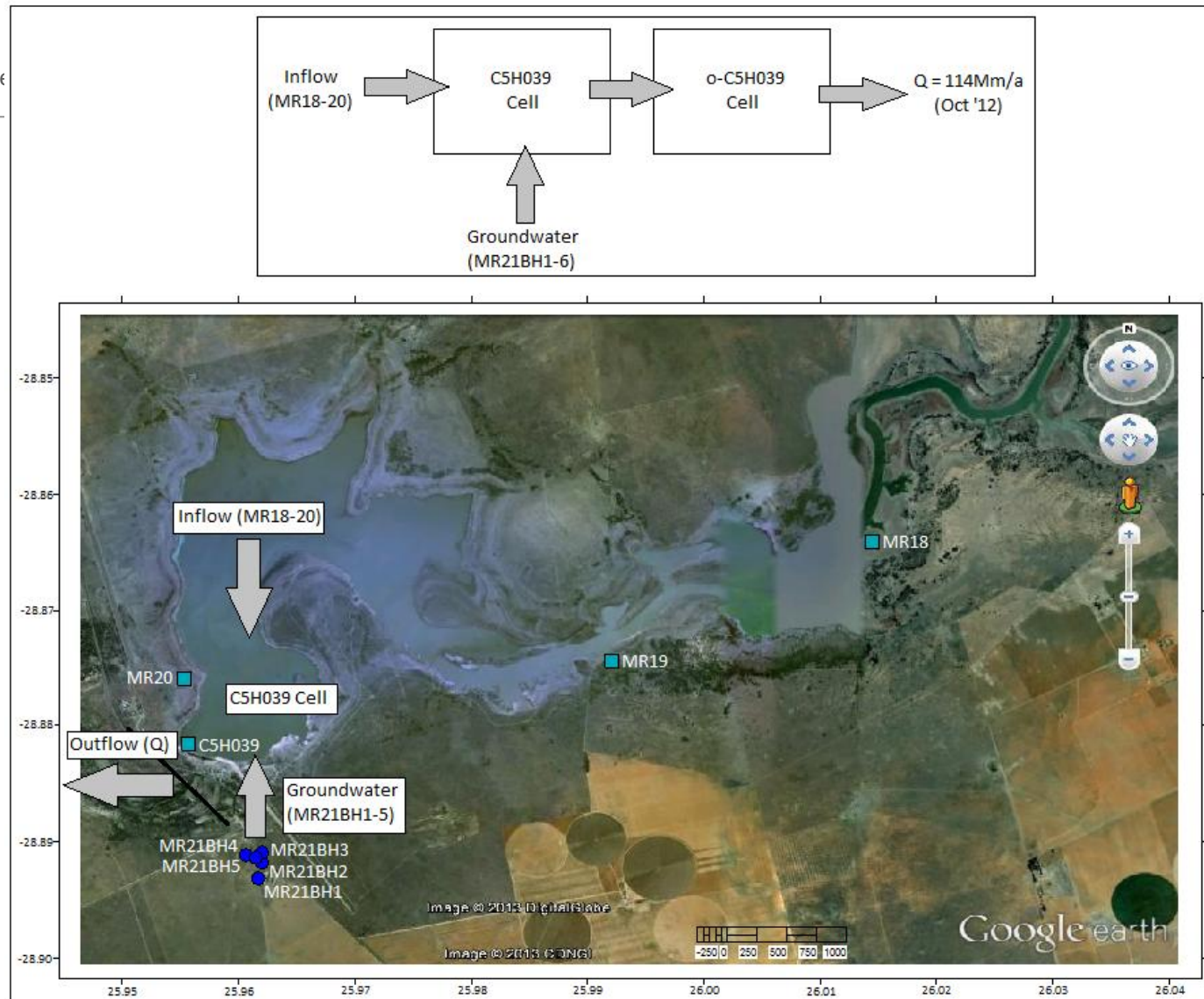
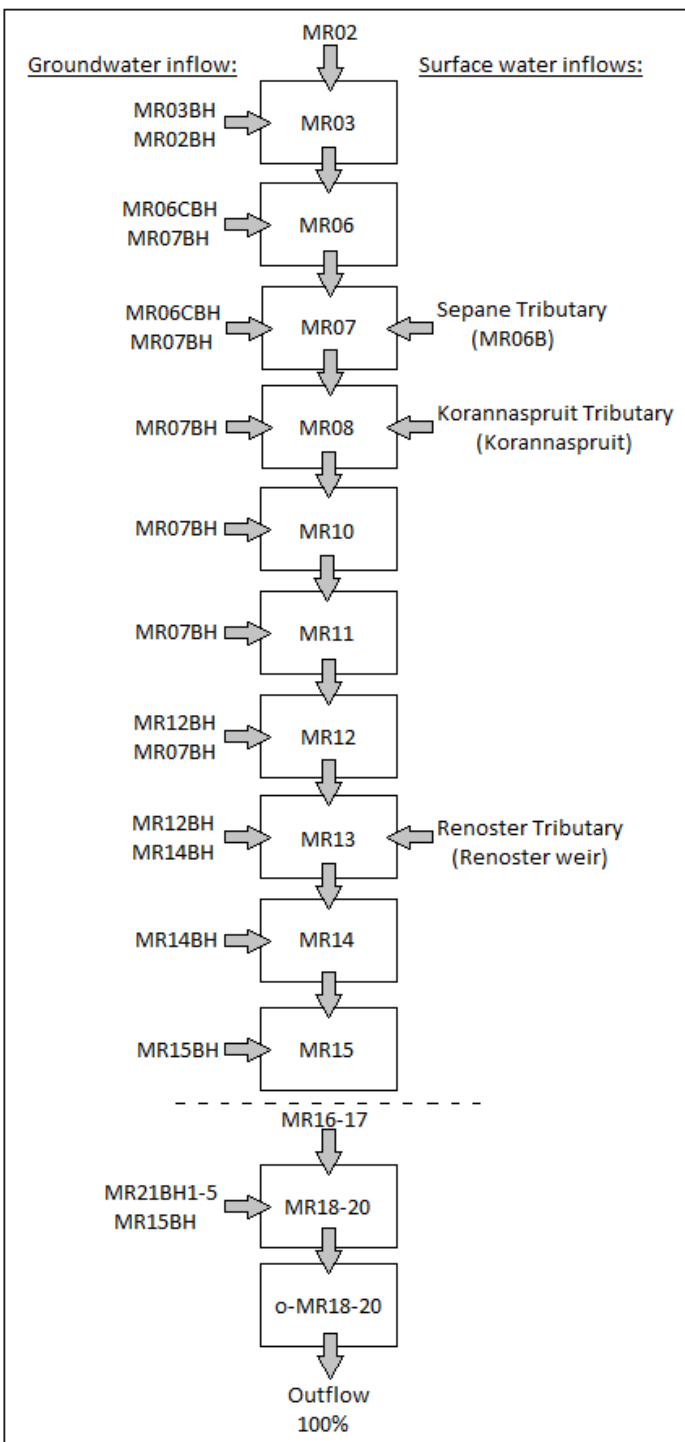


Figure 5-1-15 The box model flow conceptualisation of the various flows modelled for Section 3 and a Google Image© showing the conceptualisation on a real scale, indicating flowstation C5H039 and the various sample points.

Figure 5-1-16 The box model conceptualisation of the flow system over the entire middle Modder River study area based on sample data over January 2013.

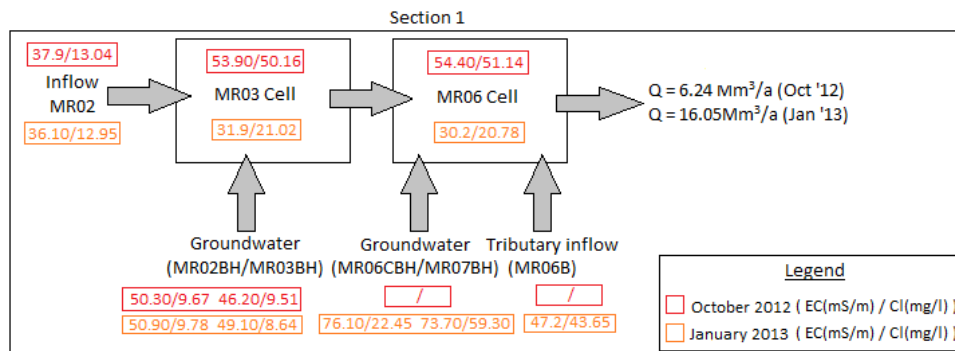


Figure 5-1-17 The representative EC and chloride (Cl) values for Section 1 flow conceptualisation. Concentrations indicate that groundwater could be contributing to the streamflow because the EC-value and chloride concentration increase as moving downstream.

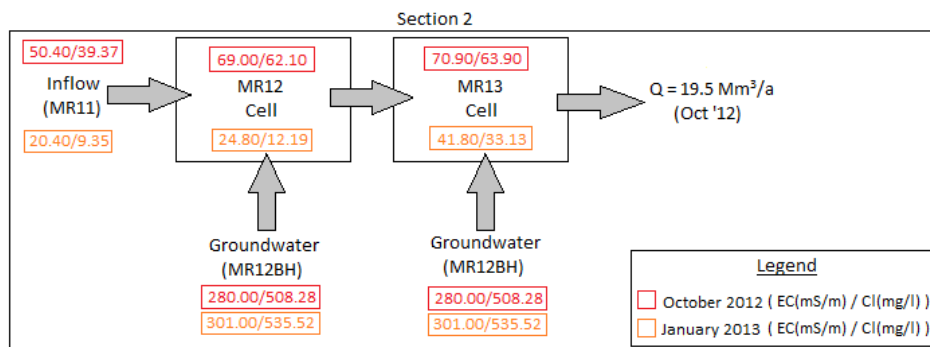


Figure 5-1-18 The representative EC and chloride (Cl) values for Section 2 flow conceptualisation. Concentration trends indicate that groundwater could be contributing to the streamflow.

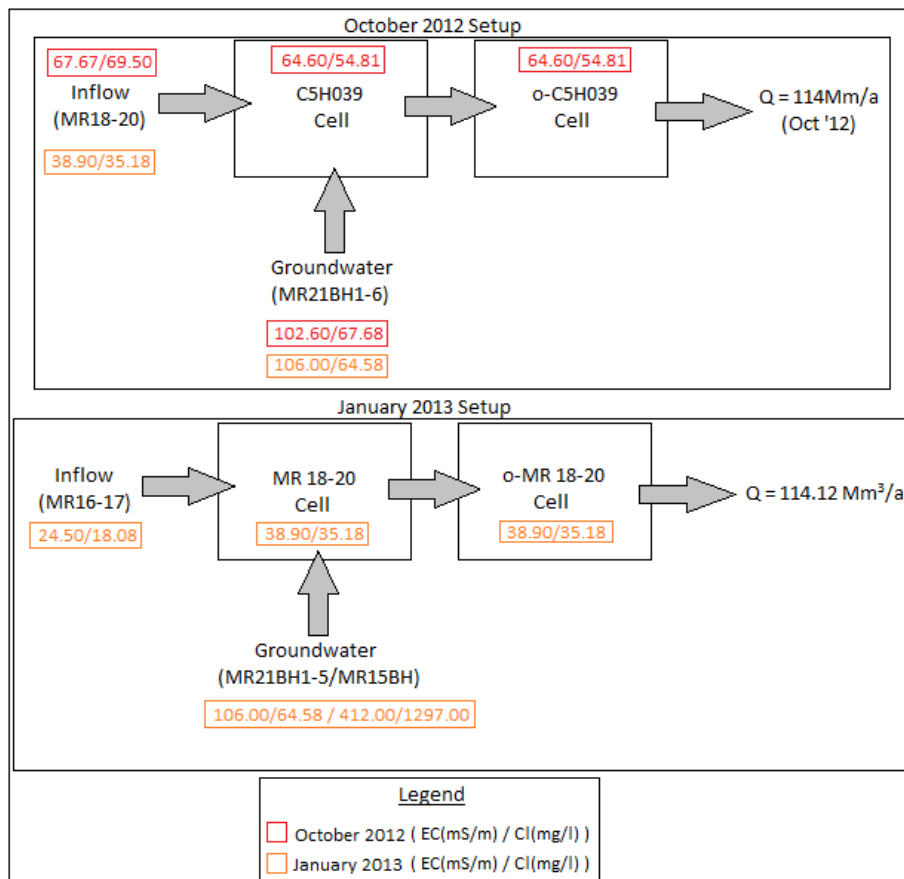


Figure 5-1-19 The representative EC and chloride (Cl) values for Section 3 flow conceptualisation, which varies for each time period due to data restrictions. Concentration trends in both conceptualisations could indicate that groundwater is contributing to streamflow.

5.1.9. Catchment Scale MCM Results

The MCM run for Section 1 using chemical data from October 2012 was unable to produce an output because the inflow sources to the defined cells were insufficient. The data collected in the January 2013 sample run did sufficiently define the inflow sources to ensure a complete MCM run. Using scenario 1 weighting factors and the January 2013 data set, the MCM estimated the groundwater inflow from the defined groundwater sources MR02BH, MR03BH, MR06CBH and MR07BH as 0.00Mm³/a, 0.04Mm³/a, 0.00Mm³/a and 0.15Mm³/a, respectively (Table 5-1-9). The sum of all the groundwater baseflow volumes estimated is 0.19Mm³/a for the Section 1 area of the Modder River. The upstream river inflow defined by the sample MR02 to Cell 1 (MR03 Cell) was determined to be 8.09Mm³/a, while the outflow from this Cell to the downstream MR06 Cell was determined to be 8.54 Mm³/a. The tributary inflow from the Sepane River to the MR06 Cell was determined to be 5.49Mm³/a. Outflow from the MR06 Cell was defined at 16.05Mm³/a from monthly flow data from flowstation C5H003. The water balance error for Section 1 using scenario 1 is 19.26% (Table 5-1-9). Using scenario 2 for the same flow setup and data, results in groundwater inflow from the defined groundwater sources MR02BH, MR03BH, MR06CBH and MR07BH as 0.00Mm³/a, 0.08Mm³/a, 0.00Mm³/a and 0.01Mm³/a, respectively (Table 5-1-9). The sum of all the groundwater inflow volumes estimated is then 0.09Mm³/a for the Section 1 area of the Modder River. The upstream river inflow defined by the sample MR02 to Cell 1 (MR03 Cell) was determined to be 8.93Mm³/a, while the outflow from this Cell to the downstream MR06 Cell was determined to be 10.33Mm³/a. The tributary inflow to the MR06 Cell was determined to be 4.33Mm³/a. Outflow from the MR06 Cell was defined at 16.05Mm³/a from monthly flow data from flowstation C5H003. The water balance error for Section 1 using scenario 2 is 21.71% (Table 5-1-9).

Table 5-1-9 Summary of the MCM output for Section 1 for the weighting factor scenarios 1 and 2, indicating the quantified inflows and the associated water balance error.

Section 1				
Sample period	Cell	Name of inflow	Rate of inflow (Mm ³ /a)	
			Scenario 1	Scenario 2
Jan '13	Cell 1	MR02	8.09	8.93
		MR02BH	0.00	0.00
		MR03BH	0.04	0.08
	Cell 2	MR06B	5.49	4.33
		MR06CBH	0.00	0.00
		MR07BH	0.15	0.01
Water balance error (%)			19.26%	21.71%

The results from the Section 1 MCM application making use of the January 2013 dataset and utilizing scenario 2 weighting factors are graphically represented in Figure 5-1-20 and expressed as percentages of the assigned total outflow. The January 2013 MCM application determined the inflow from the upstream river source MR02 to the MR03 Cell at 55.6% of the total outflow from the system. The groundwater contribution to streamflow within the MR03 Cell was estimated at 0.0% from the MR02BH defined groundwater source and 0.5% from the MR03BH defined groundwater source. Flow from the MR03 Cell to the downstream MR06 Cell was determined at 64.3% of the total flow. The groundwater contribution to streamflow within the MR06 Cell was estimated at 0.0% from both the MR06CBH and MR07BH defined groundwater sources, while the tributary inflow to the cell was estimated at 27%.

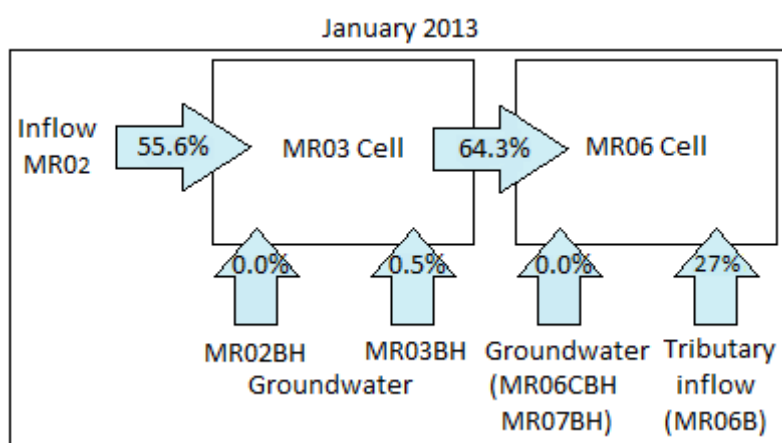


Figure 5-1-20 The unknown fluxes determined by the MCM for the January 2013 application on Section 1, making use of the scenario 2 weighting factors, expressed as a percentage of the total flow volume.

It was possible to run the MCM for both datasets from the two time periods for Section 2 as the inflows were sufficiently defined. Using the weighting factor scenario 1 and the October 2012 dataset for Section 2, the groundwater inflow from the defined groundwater source MR12BH is estimated at $1.56\text{Mm}^3/\text{a}$ and $0.07\text{Mm}^3/\text{a}$ for cells MR12 and MR13, respectively (Table 5-1-10). The sum of all the groundwater baseflow volumes estimated is $1.63\text{Mm}^3/\text{a}$ for the Section 2 area of the Modder River. The upstream river inflow defined by the sample MR11 to Cell 1 (MR12 Cell) was determined to be $15.99\text{Mm}^3/\text{a}$, while the outflow from this Cell to the downstream Cell 2 (MR13 Cell) was determined at $19.50\text{Mm}^3/\text{a}$. Outflow from the MR13 Cell was defined at $19.50\text{Mm}^3/\text{a}$ from monthly flow data from flowstation C5H053. The water balance error in October 2012 model run for Section 2 using scenario 1 is 12.11% (Table 5-1-10). Using the weighting factor scenario 2 and the October 2012 dataset for Section 2, the groundwater inflow from the defined groundwater source MR12BH is estimated at $1.33\text{Mm}^3/\text{a}$ and $0.00\text{Mm}^3/\text{a}$ for cells MR12 and MR13, respectively (Table 5-1-10). The sum of all the groundwater baseflow volumes estimated is $1.33\text{Mm}^3/\text{a}$ for the Section 2 area of the Modder

River. The upstream river inflow defined by the sample MR11 to Cell 1 (MR12 Cell) was determined to be $18.63\text{Mm}^3/\text{a}$, while the outflow from this Cell to the downstream Cell 2 (MR13 Cell) was determined at $19.67\text{Mm}^3/\text{a}$. Outflow from the MR13 Cell was defined at $19.50\text{Mm}^3/\text{a}$ from monthly flow data from flowstation C5H053. The water balance error in the October 2012 model run for Section 2 using scenario 2 is 2.63% (Table 5-1-10).

Using the weighting factor scenario 1 and the January 2013 dataset for Section 2, the groundwater inflow from the defined groundwater source MR12BH is estimated at $0.36\text{Mm}^3/\text{a}$ and $0.92\text{Mm}^3/\text{a}$ for cells MR12 and MR13, respectively (Table 5-1-10). The sum of all the groundwater baseflow volumes estimated is $1.28\text{Mm}^3/\text{a}$ for the Section 2 area of the Modder River. The upstream river inflow defined by the sample MR11 to Cell 1 (MR12 Cell) was determined to be $16.64\text{Mm}^3/\text{a}$, while the outflow from this Cell to the downstream Cell 2 (MR13 Cell) was determined at $18.96\text{Mm}^3/\text{a}$. Outflow from the MR13 Cell was defined at $19.50\text{Mm}^3/\text{a}$ from monthly flow data from flowstation C5H053. The water balance error in the January 2013 model run for Section 2 using scenario 1 is 7.23% (Table 5-1-10). Using the weighting factor scenario 2 and the January 2013 dataset for Section 2, the groundwater inflow from the defined groundwater source MR12BH is estimated at $0.32\text{Mm}^3/\text{a}$ and $0.96\text{Mm}^3/\text{a}$ for cells MR12 and MR13, respectively (Table 5-1-10). The sum of all the groundwater baseflow volumes estimated is $1.28\text{Mm}^3/\text{a}$ for the Section 2 area of the Modder River. The upstream river inflow defined by the sample MR11 to Cell 1 (MR12 Cell) was determined to be $19.78\text{Mm}^3/\text{a}$, while the outflow from this Cell to the downstream Cell 2 (MR13 Cell) was determined at $20.05\text{Mm}^3/\text{a}$. Outflow from the MR13 Cell was defined at $19.50\text{Mm}^3/\text{a}$ from monthly flow data from flowstation C5H053. The water balance error in the January 2013 model run for Section 2 using scenario 1 is 2.75% (Table 5-1-10).

Table 5-1-10 Summary of the MCM output for Section 2 for the weighting factor scenarios 1 and 2, indicating the quantified inflows and the associated water balance error.

Section 2				
Sample period	Cell	Name of inflow	Rate of inflow (Mm^3/a)	
			Scenario 1	Scenario 2
Oct '12	Cell 1	MR11	15.99	18.63
		MR12BH	1.56	1.33
	Cell 2	MR12BH	0.07	0.00
Water balance error (%)			12.11%	2.63%
Jan '13	Cell 1	MR11	16.64	19.78
		MR12BH	0.36	0.32
	Cell 2	MR12BH	0.92	0.96
Water balance error (%)			7.23%	2.75%

The results from the Section 2 MCM application on both temporal scales utilizing scenario 2 weighting factors are graphically represented in Figure 5-1-21 and expressed as percentages of the assigned total outflow. The October 2012 MCM application determined the inflow from the upstream river source MR11 to the MR12 Cell to be 97.8% of the total outflow from the system. The groundwater contribution to streamflow within the MR12 Cell, from the MR12BH defined groundwater source, was estimated at 6.9% of the total assigned flow. Flow from the MR12 Cell to the downstream MR13 Cell was determined at 103.2% of the total flow. The groundwater contribution to streamflow within the MR13 Cell, from the MR12BH defined groundwater source, was estimated at 0.0%. The January 2013 MCM application determined the inflow from the upstream river source MR11 to the MR12 Cell at 101.4% of the total outflow from the system. The groundwater contribution to streamflow within the MR12 Cell, from the MR12BH defined groundwater source, was estimated at 1.6% of the total assigned outflow. Flow from the MR12 Cell to the downstream MR13 Cell was determined at 102.8% of the total flow. The groundwater contribution to streamflow within the MR13 Cell, from the MR12BH groundwater source, was estimated at 4.9% of the total assigned outflow.

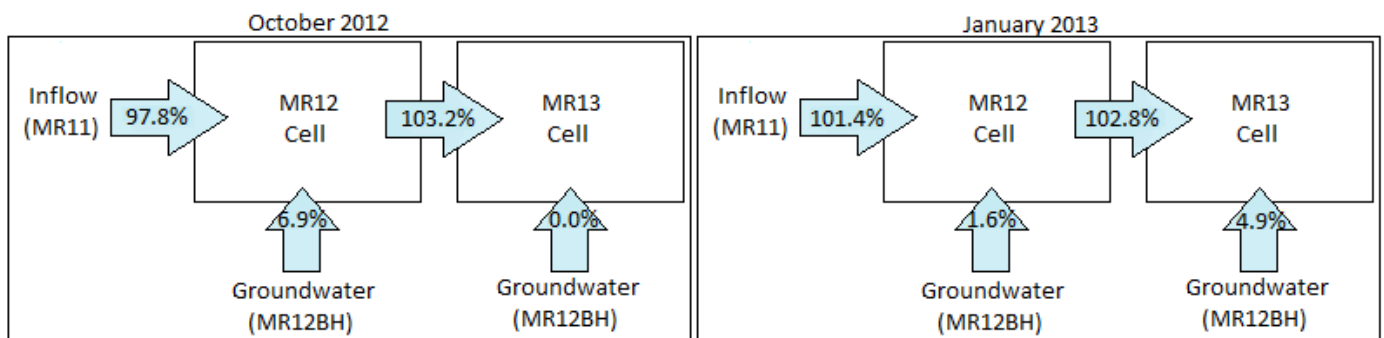


Figure 5-1-21 The unknown fluxes determined by the MCM for both the October 2012 and January 2013 applications on Section 2, making use of the scenario 2 weighting factors, expressed as a percentage of the total flow volume.

The model setup for Section 3 differs for the two time periods because water quality data from the flowstation C5H039 sample point was not available for the January 2013 time period. Using weighting factor scenario 1, the original flow setup and the October 2012 dataset for Section 3, inflow from the groundwater source defined by the average tracer concentration from sampled boreholes MR21BH1 – 6 is estimated at $2.87\text{Mm}^3/\text{a}$ (Table 5-1-11). The upstream river inflow defined by the samples MR18-20, to Cell 1 (C5H039 Cell) was determined to be $100.77\text{Mm}^3/\text{a}$, while the outflow from this Cell to the downstream fictitious Cell 2 (o-C5H039 Cell) was determined at $112.30\text{Mm}^3/\text{a}$. Outflow from the fictitious cell was defined at $114.12\text{Mm}^3/\text{a}$ from monthly flow data from flowstation C5H039. The water balance error in the October 2012 model run for Section 3 using scenario 1 is 9.88% (Table 5-1-11). Using the weighting factor

scenario 2, the original flow setup and the October 2012 dataset for Section 3, inflow from the groundwater source defined by the average tracer concentration from sampled boreholes MR21BH1-6 was estimated at $2.61\text{Mm}^3/\text{a}$ (Table 5-1-11). The upstream river inflow defined by the samples MR18-20, to Cell 1 (C5H039 Cell) was determined to be $102.66\text{Mm}^3/\text{a}$, while the outflow from this Cell to the downstream fictitious Cell 2 (o-C5H039 Cell) was determined at $113.34\text{Mm}^3/\text{a}$. Outflow from the fictitious cell was defined at $114.12\text{Mm}^3/\text{a}$ from monthly flow data from flowstation C5H039. The water balance error in the October 2012 model run for Section 3 using scenario 2 is 8.55% (Table 5-1-11).

Using the weighting factor scenario 1, the flow setup to compensate for the lack of chemical data at C5H039 and the January 2013 dataset for Section 3, inflow from the groundwater sources defined by the average tracer concentration from sampled boreholes MR21BH1 – 6 and MR15BH, are estimated at $0.00\text{Mm}^3/\text{a}$ and $2.38\text{Mm}^3/\text{a}$, respectively (Table 5-1-11). The upstream river inflow defined by the samples MR16-17, to Cell 1 (MR18-20 Cell) was determined to be $24.67\text{Mm}^3/\text{a}$, while the outflow from this Cell to the downstream fictitious Cell 2 (o-MR18-20 Cell) was determined at $79.94\text{Mm}^3/\text{a}$. Outflow from the fictitious cell was defined at $114.12\text{Mm}^3/\text{a}$ from monthly flow data from flowstation C5H039. The water balance error in the January 2013 model run for Section 3 using scenario 1 is 76.50% (Table 5-1-11).

Using the weighting factor scenario 2, the flow setup to compensate for the lack of chemical data at C5H039 and the January 2013 dataset for Section 3, inflow from the groundwater sources defined by the average tracer concentration set from sampled boreholes MR21BH1 – 6 and MR15BH, are estimated at $0.00\text{Mm}^3/\text{a}$ and $2.13\text{Mm}^3/\text{a}$, respectively (Table 5-1-11). The upstream river inflow defined by samples MR16-17, to Cell 1 (MR18-20 Cell) was determined to be $62.13\text{Mm}^3/\text{a}$, while the outflow from this Cell to the downstream fictitious Cell 2 (o-MR18-20 Cell) was determined at $92.83\text{Mm}^3/\text{a}$. Outflow from the fictitious cell was defined at $114.12\text{Mm}^3/\text{a}$ from monthly flow data from flowstation C5H039. The water balance error in the January 2013 model run for Section 3 using scenario 2 is 44.18% (Table 5-1-11).

Table 5-1-11 Summary of the MCM output for Section 3 for the weighting factor scenarios 1 and 2, indicating the quantified inflows and the associated water balance error.

Section 3				
Sample period	Cell	Name of inflow	Rate of inflow (Mm ³ /a)	
			Scenario 1	Scenario 2
Oct '12	Cell 1	MR18-20	100.77	102.66
		MR21BHave	2.87	2.61
Water balance error (%)			9.88%	8.55%
Jan '13	Cell 1	MR15BH	2.38	2.13
		MR16-17	24.67	62.13
		MR21BHave	0	0
Water balance error (%)			76.50%	44.18%

The results from the Section 3 MCM application on both temporal scales utilizing scenario 2 weighting factors are graphically represented in Figure 5-1-22 and expressed as percentages of the assigned total outflow. The October 2012 MCM application determined the inflow from the upstream river source defined by river samples MR18-20 to the C5H039 Cell at 89.95% of the total outflow from the system. The groundwater contribution to streamflow within the C5H039 Cell was estimated at 2.3% from the MR21BH1-6 defined groundwater source. Flow from the C5H039 Cell to the downstream fictitious o-C5H039 Cell was determined at 116.8% of the total flow. The January 2013 MCM setup differs from the October 2012 setup due to data restrictions. This MCM run determined the inflow from the upstream river source defined by the average of samples MR16-17 to the MR18-20 Cell at 54.4% of the total outflow from the system. The groundwater contribution to streamflow within the MR18-20 Cell was estimated at 1.8% from the MR21BH1-6 and MR15BH defined groundwater sources. Flow from the MR18-20 Cell to the downstream fictitious o-MR18-20 Cell was determined at 81.3% of the total flow.

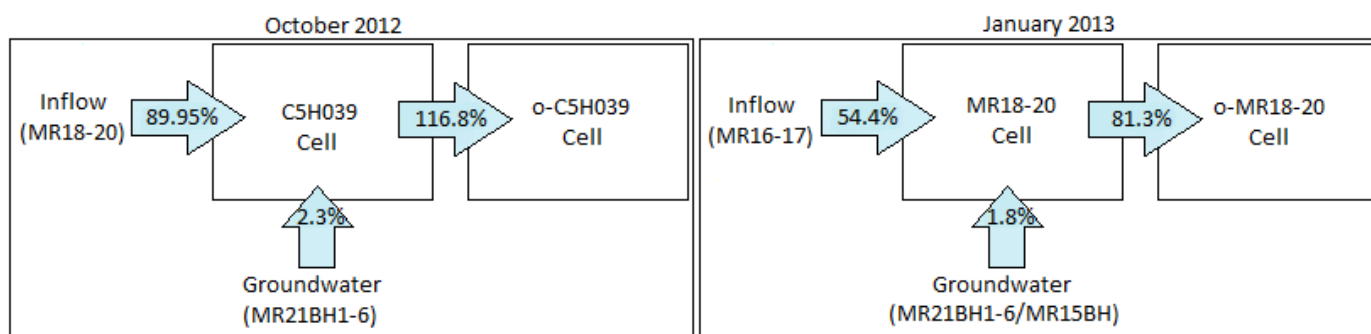


Figure 5-1-22 The unknown fluxes determined by the MCM for both the October 2012 and January 2013 applications on Section 3, making use of the scenario 2 weighting factors, expressed as a percentage of the total flow volume. The two temporal scale applications differ in setup due to data restrictions.

The continuous flow system setup from MR01 - MR15 and MR16 - MR20 using data from January 2013 was used to perform a MCM run which defined the outflow from the system at 100%, instead of assigning an outflow volume. The quantified inflows to the system are expressed as percentages relative to this 100% outflow. The results from this model run, making use of scenario 1, are graphically displayed along with the flow system diagram (Figure 5-1-23). The Renoster tributary sample is not included in the model run as this sample insufficiently defines an additional inflow to the system and results in an incomplete model run. Cell 1 (MR03 Cell) receives 2.32% inflow from the upstream MR02 source, and receives 0.02% and 0.00% from groundwater sources MR03BH and MR02BH. Cell 2 (MR06 Cell) receives 2.68% upstream inflow from the MR03 Cell, and receives 0.15% and 0.00% groundwater inflow from MR06CBH and MR07BH, respectively. Cell 3 (MR07 Cell) receives an upstream inflow of 3.20% from the MR06 Cell as well as an additional surface water inflow from the Sepane tributary at 2.3%. The groundwater inflow to the MR07 Cell is 0.00% and 0.32% from MR06CBH and MR07BH, respectively. Cell 4 (MR08 Cell) also receives an additional surface water inflow from the Korannaspruit tributary at 53.76% and an upstream inflow from the MR07 Cell at 6.42%. The groundwater inflow to this cell is 1.47% from MR07BH. Cell 5 (MR10 Cell) receives 63.94% upstream inflow from the MR08 Cell, with a groundwater contribution at 0.00% from MR07BH. Cell 6 (MR11 Cell) receives 69.60% from the upstream cell MR10 Cell and a groundwater inflow of 1.28% from the MR07BH defined groundwater source. Cell 7 (MR12 Cell) receives 74.53% inflow from the upstream cell MR11 and receives groundwater inflow from the defined sources, MR12BH and MR07BH at 1.10% and 0.00%, respectively. Cell 8 (MR13 Cell) receives 77.77% upstream river inflow from the MR12 Cell and groundwater inflow of 3.19% and 2.02% from MR12BH and MR14BH, respectively. The downstream cell 9 (MR14 Cell) receives inflow from the MR13 Cell at 77.99% and 1.48% groundwater inflow from MR14BH defined source. The last cell, MR15 Cell, receives 84.73% of upstream river inflow from the upstream MR14 Cell and 0.41% groundwater inflow from MR15BH, while the outflow from this cell is set at 100%. There is a break in the continuous flow model run at this point as the sources are insufficiently defined for a complete model run. An adjacent model run is thus performed for the remaining sample data (MR16 – MR20), with the outflow of this section also set to 100%. Cell 1 of the adjacent model run is MR18-20, which receives an upstream inflow of 54.51% from the MR16-17 defined source and groundwater inflow from MR21BH1-5 and MR15BH at 0.00% and 1.87%, respectively. This cell flows into a fictitious last cell (o-MR18-20) at a rate of 81.44%. The outflow from the fictitious cell is set to 100%.

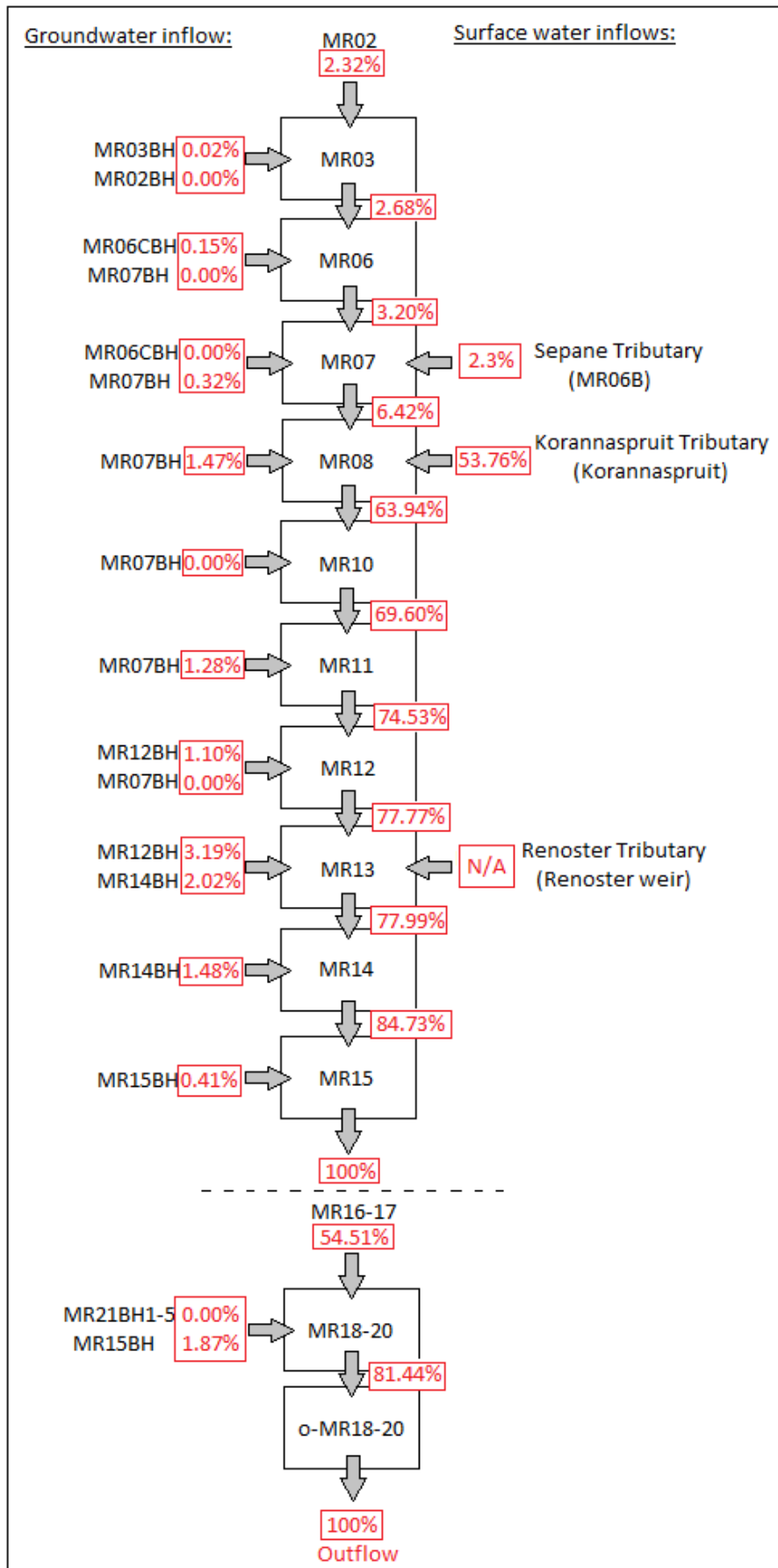


Figure 5-1-23 The box model diagram of the flow system conceptualised for the continuous flow MCM run, with the quantified percentage flows.

5.1.10. Catchment Scale MCM Discussion and Comparison

The total sum of the quantified groundwater inflows into each section of the Modder River are summarised in Table 5-1-12 for each weighting factor scenario and time period. Table 5-1-12 includes the water balance and average chemical mass balance errors associated with the quantified groundwater baseflow volume for each model run. Appendix C Table C-8 is a detailed chemical mass balance percentage error report on each chemical constituent for all model runs. There are no results for Section 1 making use of the October 2012 data as the chemical data sampled was not sufficient for a MCM run. The MCM quantification of groundwater baseflow for Section 1 using January 2013 data shows a decrease from $0.19\text{Mm}^3/\text{a}$ to $0.09\text{Mm}^3/\text{a}$ when changing from scenario 1 to scenario 2. This decrease in groundwater baseflow is however accompanied with an increase in the water and chemical balance percentage error. This increase in both water and chemical mass balances is unexpected and opposite to the response of all other model runs, where a decrease in water and chemical mass balance percentage errors are seen when scenario 2 is implemented. This increase in percentage error could be attributed to the fact that an important tracer constituent has been given a lower weighting factor, or that the flow system has been insufficiently defined with either the omission of an inflow source or inaccurate river cell definition, among other reasons. The MCM quantification of groundwater baseflow for Section 2 using October 2012 data shows a decrease from $1.63\text{Mm}^3/\text{a}$ to $1.33\text{Mm}^3/\text{a}$ when changing from scenario 1 to scenario 2 accompanied with a decrease in both the water and chemical mass balance percentage errors. However, the groundwater baseflow estimate for this section using January 2013 data does not show a change in the sum of the groundwater inflows when changing from scenario 1 to scenario 2, even though the portion of the inflow from the two different sources changes slightly (Table 5-1-10). The water and chemical balance percentage errors both decrease for this model application though, when scenario 2 is implemented. The MCM quantification of groundwater baseflow for Section 3 using October 2012 data shows only a slight decrease from $2.87\text{Mm}^3/\text{a}$ to $2.61\text{Mm}^3/\text{a}$ when changing from scenario 1 to scenario 2. The change from scenario 1 to scenario 2 is also accompanied with a slight decrease in both the water and chemical mass balance percentage errors. The groundwater baseflow estimate for this section using January 2013 data also shows a slight decrease from $2.38\text{Mm}^3/\text{a}$ to $2.13\text{Mm}^3/\text{a}$, but is accompanied by a large decrease in the water and chemical mass balance percentage errors. The percentage errors are extremely high for this model run, which could be due to the fact that a differently defined flow setup was used, because data used in the October 2012 model run was not

available for January 2013. The Section 3 study area mainly comprises the Krugersdrift Dam the Modder River where one would expect estimations of flow volumes to become more variable as natural flow is no longer taking place. Considering the various changes in the groundwater baseflow estimates and the water and chemical mass balance percentage errors, scenario 2 seems a reasonable selection of weighting factors as it results a lower associated error in most cases and by incorporating lower weighting factors attempts to decrease the infringement of the natural system on one of the main assumptions of the MCM, that of conservative tracers.

Table 5-1-12 The total groundwater baseflow estimates for each section and time period using both scenarios with the associated water and chemical mass balance percentage errors.

Section	GW Baseflow (Mm ³ /a)		Water balance error (%)		Average chemical mass balance error (%)	
	Scenario 1	Scenario 2	Scenario 1	Scenario 2	Scenario 1	Scenario 2
Section 1 (Jan '13)	1.49	0.96	19.3%	21.7%	6.2%	11.3%
Section 2 (Oct '12)	1.63	1.33	12.1%	2.6%	13.4%	9.5%
Section 2 (Jan '13)	1.28	1.28	7.2%	2.8%	7.4%	5.0%
Section 3 (Oct '12)	2.87	2.61	9.9%	8.6%	4.2%	3.7%
Section 3 (Jan '13)	2.38	2.13	76.5%	44.2%	60.3%	26.7%

Section 1, the segment of the Modder River from the Rustfontein Dam to the flowstation C5H003, falls within the quaternary catchment C52B. Similarly, Section 2 and Section 3 fall within the quaternary catchments C52E and C52G, respectively. Groundwater baseflow estimates from the Pitman, Hughes and Sami models for each of the quaternary catchments as well as an estimate from another method incorporating water quality data, namely the Tracer method, using the same data as the MCM are shown in Table 5-1-13. The Tracer method estimate is shown for both the use of only the electrical conductivity (EC) tracer as well as an average value from using a number of tracers. The groundwater baseflow estimate from the MCM and the Tracer method is based on water quality data from both October 2012 and January 2013. The Pitman, Sami and Hughes model groundwater baseflow estimates for the quaternary catchment C52B are 0.00Mm³/a, 0.03Mm³/a and 5.03Mm³/a, respectively (Table 5-1-13). The Tracer method groundwater baseflow estimate for Section 1 using data from October 2012 and only EC as a tracer is 0.51Mm³/a, while for an average from a number of tracers is 1.24Mm³/a. Similarly, using data from January 2013 results in a groundwater estimate of 2.12Mm³/a for EC alone and 1.61Mm³/a for an average from a number of tracers. The MCM groundwater baseflow estimate for section 1 is 0.09Mm³/a based on data from January 2013.

From the various groundwater estimates for Section 1, it can be seen that the Tracer method produces variable results between the two time periods as well as between using EC alone and using an average of a number of tracers. The MCM and the Tracer method found groundwater baseflow to be contributing to the Modder River within this section unlike the Pitman model that estimated zero groundwater baseflow. The MCM and Tracer method found slightly more baseflow than estimated by the Sami model at $0.03\text{Mm}^3/\text{a}$, but much less groundwater baseflow than was found in the Hughes model at $5.03\text{Mm}^3/\text{a}$.

The Pitman, Sami and Hughes model groundwater baseflow estimates for the quaternary catchment C52E are $0.00\text{Mm}^3/\text{a}$, $0.00\text{Mm}^3/\text{a}$ and $2.22\text{Mm}^3/\text{a}$, respectively (Table 5-1-13). The MCM estimates the groundwater baseflow for this section at $1.33\text{Mm}^3/\text{a}$ in October 2012 and $1.28\text{Mm}^3/\text{a}$ in January 2013. The Tracer method estimates the groundwater baseflow for EC alone and multiple tracers at $0.21\text{Mm}^3/\text{a}$ and $1.35\text{Mm}^3/\text{a}$, respectively for October 2012 and $1.35\text{Mm}^3/\text{a}$ and $2.13\text{Mm}^3/\text{a}$ for January 2013. The tracer method proves to be quite variable with the MCM producing a fairly constant value for both time periods. The MCM and the tracer method found groundwater baseflow to be contributing to the Modder River within this section unlike the Pitman and Sami models that both estimated $0.00\text{Mm}^3/\text{a}$ groundwater baseflow, but found less groundwater baseflow than the Hughes model that estimates the groundwater baseflow at $2.22\text{Mm}^3/\text{a}$.

Pitman, Sami and Hughes model groundwater baseflow estimates for the quaternary catchment C52G are $0.00\text{Mm}^3/\text{a}$, $0.00\text{Mm}^3/\text{a}$ and $5.35\text{Mm}^3/\text{a}$, respectively (Table 5-1-13). The MCM estimates the groundwater baseflow for this section at $2.61\text{Mm}^3/\text{a}$ in October 2012 and $2.13\text{Mm}^3/\text{a}$ in January 2013. The Tracer method estimates the groundwater baseflow for EC alone and multiple tracers at $0.04\text{Mm}^3/\text{a}$ and $0.61\text{Mm}^3/\text{a}$, respectively in October 2012 and $4.24\text{Mm}^3/\text{a}$ and $4.25\text{Mm}^3/\text{a}$ in January 2013. The Tracer method seems to be more stable between using EC alone and a number of tracers for this section, but quite variable between time periods. The MCM produces similar groundwater estimates for both time periods. Both water quality methods found groundwater baseflow contributing to the river, while the Pitman and Sami model estimate zero groundwater baseflow to this section of the Modder River. On the other hand, the water quality methods indicate much less groundwater baseflow than the Hughes model at $5.35\text{Mm}^3/\text{a}$.

Table 5-1-13 Groundwater baseflow estimates from the Pitman model, Sami model, Hughes model, Tracer method and MCM for Section 1 (C52B), Section 2 (C52E) and Section 3 (C52G).

Quat/ Section	Pitman uncalibrated (Mm ³ /a)	Sami uncalibrated (Mm ³ /a)	Hughes uncalibrated (Mm ³ /a)	MCM (Mm ³ /a)		Tracer Method			
				Oct '12	Jan '13	Oct '12		Jan '13	
						EC	Average	EC	Average
C52B/ Section 1	0.00	0.03	5.03	/	0.09	0.51	1.24	2.12	1.61
C52E/ Section 2	0.00	0.00	2.22	1.33	1.28	0.21	1.35	1.35	2.13
C52G/ Section 3	0.00	0.00	5.35	2.61	2.13	0.04	0.61	4.24	4.25

Considering the results from these three sections, the trend seen is for the MCM groundwater baseflow volume to seem like an over-estimate when compared to the Sami and Pitman model estimates, but to seem like an under-estimate when compared to the Hughes model estimate. Considering work done by Welderufael and Woyessa (2010) which found that baseflow contributed on average 71% of the total streamflow in the quaternary catchment C52A using four baseflow separation techniques and work by Gomo (2011) also reporting the Modder River is a gaining stream at the base of the Krugersdrift Dam, it seems reasonable to assume that there is some groundwater contributing to the baseflow of the river, even if it is in small quantities.

The continuous flow model run was performed as if the study area was an ungauged catchment. The model run shows an overall groundwater contribution to this approximately 130km section of the Modder River of 11.44% of the total river flow at MR15, and an additional 1.87% of the total flow at the Krugersdrift dam (Figure 5-1-23). The water and chemical mass balance errors associated with these model runs are high at 30.87% and 13.89% for MR01-MR15, and 44.18% and 26.71% for MR16-MR20, respectively. The contribution of the Korannaspruit River tributary seems to have been over-estimated at 53.76% of the total river flow. This over-estimate could be attributed to insufficiently defined runoff sources which have resulted in a large % being attributed to this single source. However, the MCM does seem to give reasonable groundwater inflow estimates to the river system. The MCM could thus serve as an initial estimate method for the contribution of groundwater to baseflow in ungauged flow systems.

5.2. Pilot Study 2: Limpopo Quaternary Catchments A42A – A42C

5.2.1. Overview

The quaternary catchments A42A, A42B and A42C, fall within the Limpopo Water Management Area (WMA) of the Limpopo Province, South Africa (Figure 5-2-1). The quaternary catchments form the upper most source area of the Moloko River, which is the highest yielding river in the WMA. The Sand River in quaternary A42A, the Grootspuit and Sandspruit Rivers in A42B and the Klein Sand River in A42C are tributaries to the Moloko River. The average annual rainfall within this area is between 400mm and 700mm. The total annual rainfall was 647mm and the average annual evaporation was 1582.9mm in the year 2005, the selected period of investigation. The average annual temperature ranges from 14°C to 20°C. The area is characterised by a flat open landscape surrounded by mountains, where streams flow through steep, rocky areas. The topography varies from 1200mamsl in the river valley to 1700mamsl in the surrounding mountains. The predominant vegetation is Waterberg Moist Mountain Bushveld and Mixed Bushveld (Department of Environmental Affairs and Tourism, 2006). The warms database for the A42 quaternaries indicate that the abstraction from surface water resources in the area per annum is 33.4Mm³.

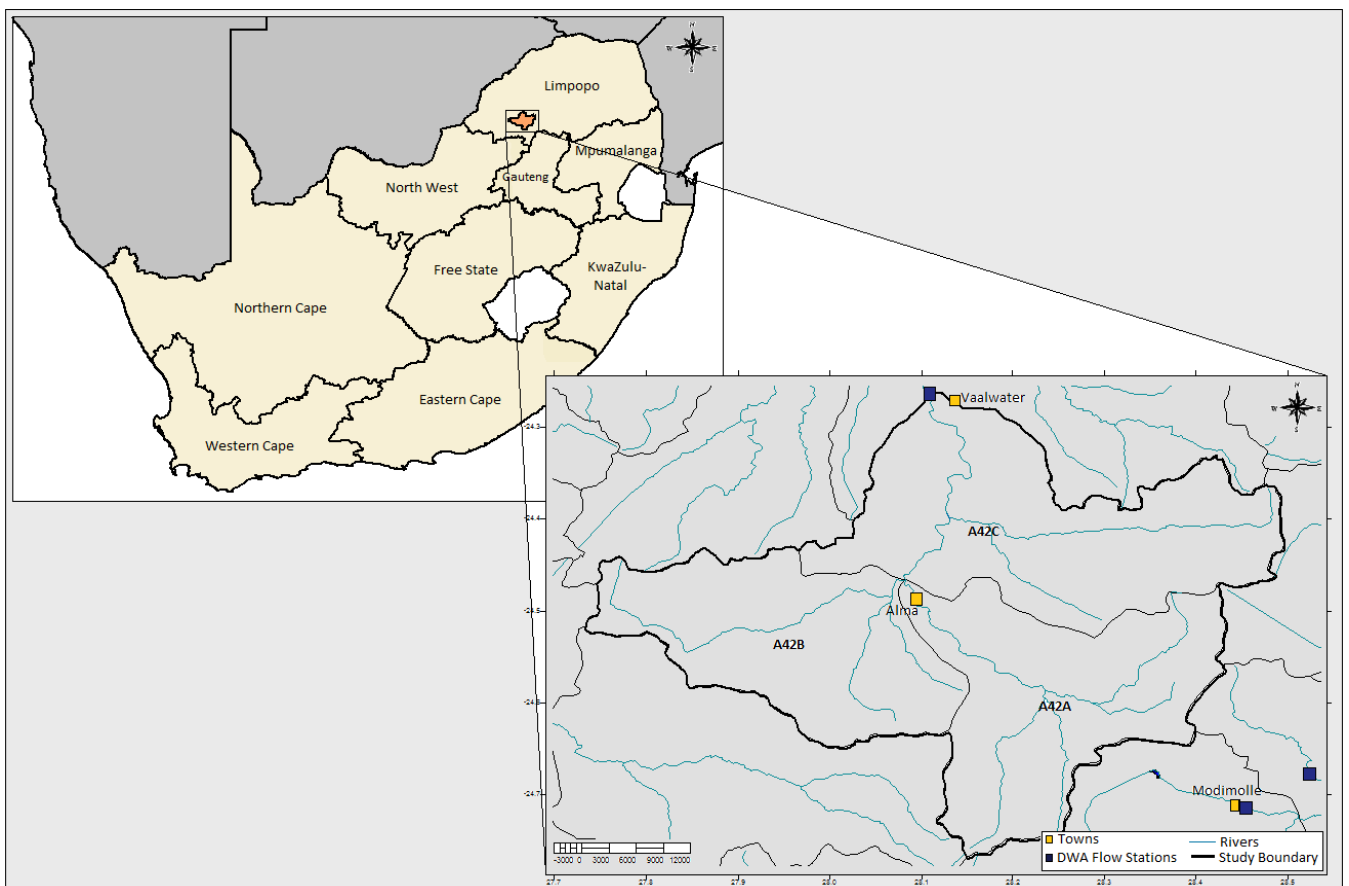


Figure 5-2-1 The position of the quaternaries A42A, A42B and A42C within South Africa.

5.2.2. Geology

The outcrop geology of the area mainly consists of Waterberg and Rooiberg Group lithologies (Figure 5-2-2). Minor intrusions of the Lebowa Granite Suite are also present in the study area along with the Rooiberg Group they make up part of the Bushveld Magmatic Province. The Shrikklouf and Kwaggasnek Formations, representing the volcanic Rooiberg Group, are the oldest rocks in the study area (Table 5-2-1). A quartzite layer forms the base of the Kwaggasnek Formation, overlain by a thick layer of siliceous lavas (rhyolite) and capped by a laterally extensive shale/tuff layer that is underlain by a layer of volcanic breccias. The Shrikklouf Formation consists mainly of siliceous lavas (flow-banded rhyolite), with a layer of ash-flow tuff marking the top of the unit. The Lebowa Granite Suite intrudes above the Rustenburg Layered Suite of the Bushveld Complex, but does not outcrop with the study area. The Lebowa Granite Suite consists of several granite types, namely Nebo, Bobbejaankop, Klipklouf and Makhutso Granite. The small and isolated Glentig Formation outcropping in the study area consists of predominantly argillaceous, clastic sedimentary rocks with interbedded lavas and a basal conglomerate of reworked volcanic material. This formation used to be classed as the uppermost beds of the Transvaal Supergroup, but is now considered as proto-Waterberg deposits (Johnson, Anhaeusser and Thomas, 2006).

The Waterberg Group lithologies which outcrop within the study area form a complete stratigraphic column of the group (Table 5-2-1). The Waterberg Group lies unconformably on the rocks of the Bushveld Complex in the study area. The rocks of the Waterberg Group are usually dark greyish-red in colour. The oldest formation within the Waterberg group is the Swaershoek Formation which consists of mainly fractured arenites and rudites, and thought to have been deposited as a fan-delta, or inter fan-delta tidal flats. The Alma Formation overlies the Swaershoek Formation and is made up of a succession of various arkoses and feldspathic arenites, deposited as a series of alluvial fans. The overlying Skilpadkop Formation consists of thickly bedded immature lithic arenites, pebble rudites and minor quantities of arkose. The Aäsvaelkop Formation coarsens from arenaceous lutites at its base, overlying the Skilpadkop Formation, to arenites higher up, indicating deposition in a shallow inland lake environment. The Sandriviersberg Formation consists of arenite, rudite and interbedded pebble rudites. The Cleremont Formation is thought to have been deposited as arenaceous sediments along a shoreline, consisting of medium-grained well-sorted arenites. The Formation maintains a constant thickness of approximately 125m. The Vaalwater Formation, the topmost formation of

the Waterberg Group, comprises poorly exposed feldspathic arenites and lutites thought to have been deposited in a lower energy shelf setting (Johnson, Anhaeusser and Thomas, 2006).

A conceptual geological cross-section of the area is created based on the outcrop geology map (Figure 5-2-3). A general cross-section from the south to the north of the area is conceptualised as a folded landscape to account for the reverse geological sequence along with a decrease in topography. The outcrop geology also indicates the presence of a large fault to the south. The area is intruded by Lebowa granites and various basic intrusive lithologies, making for a complicated geology.

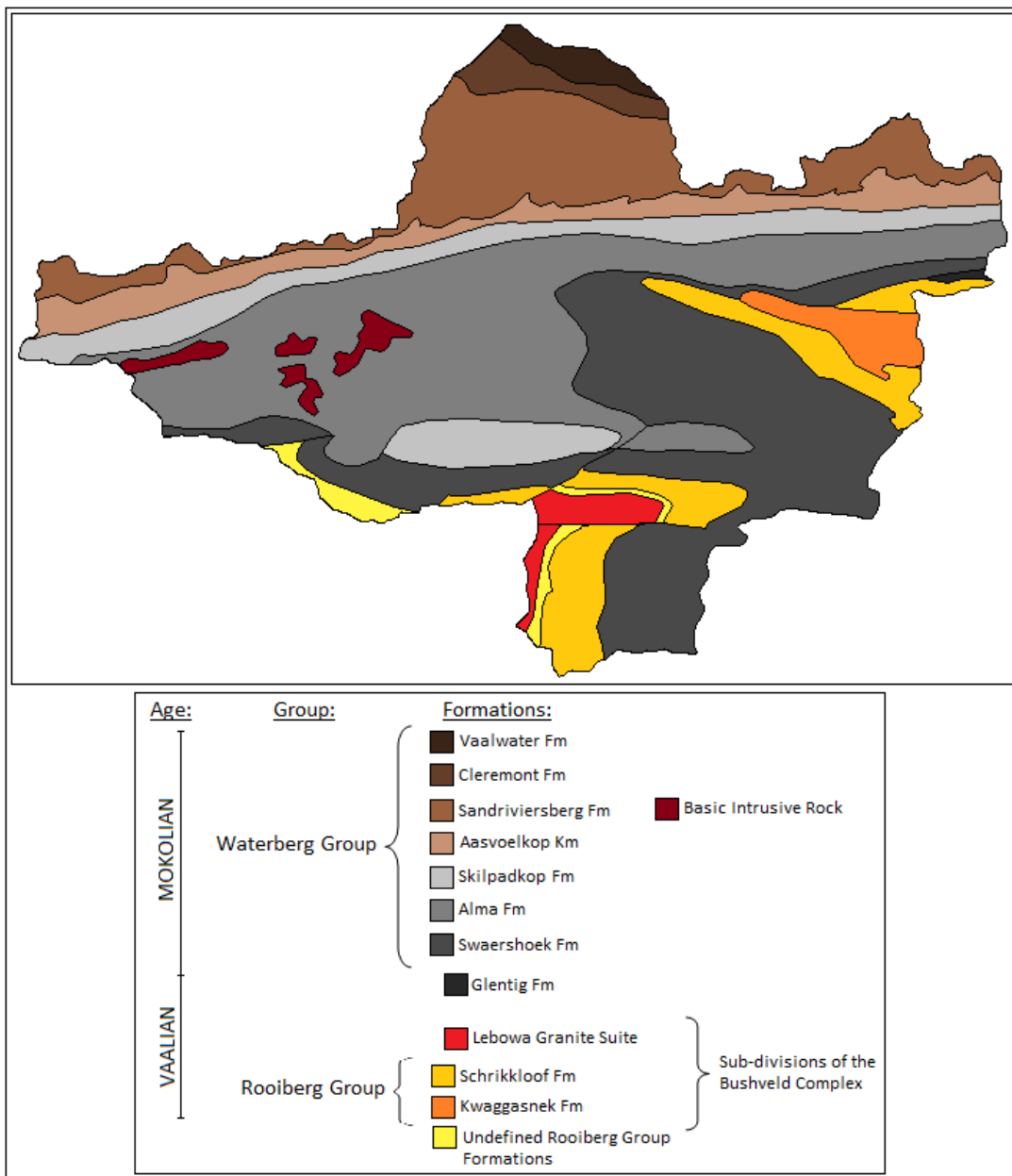


Figure 5-2-2 The outcrop geology of the A42 study area. (Based data from GRDM, 2010).

Table 5-2-1 The stratigraphic sequence within the A42 study area

Era	GROUP	SUB-GROUP	FORMTATIONS	THICKNESS
MESOZOIC	Waterberg	Kransberg	Vaalwater	<475m
			Cleremont	~125m
			Sandriversberg	1250m
		Matlasbas	Aasvoelkop	<600m
			Skilpad	<600m
		Nylstroom	Alma	<3000m
Swaershoek	<1000m			
PALEOZOIC	Proto-Waterberg		Glentig	
	Rooiberg		Schrikkloof	>1000m
			Kwaggasnek	>1000m

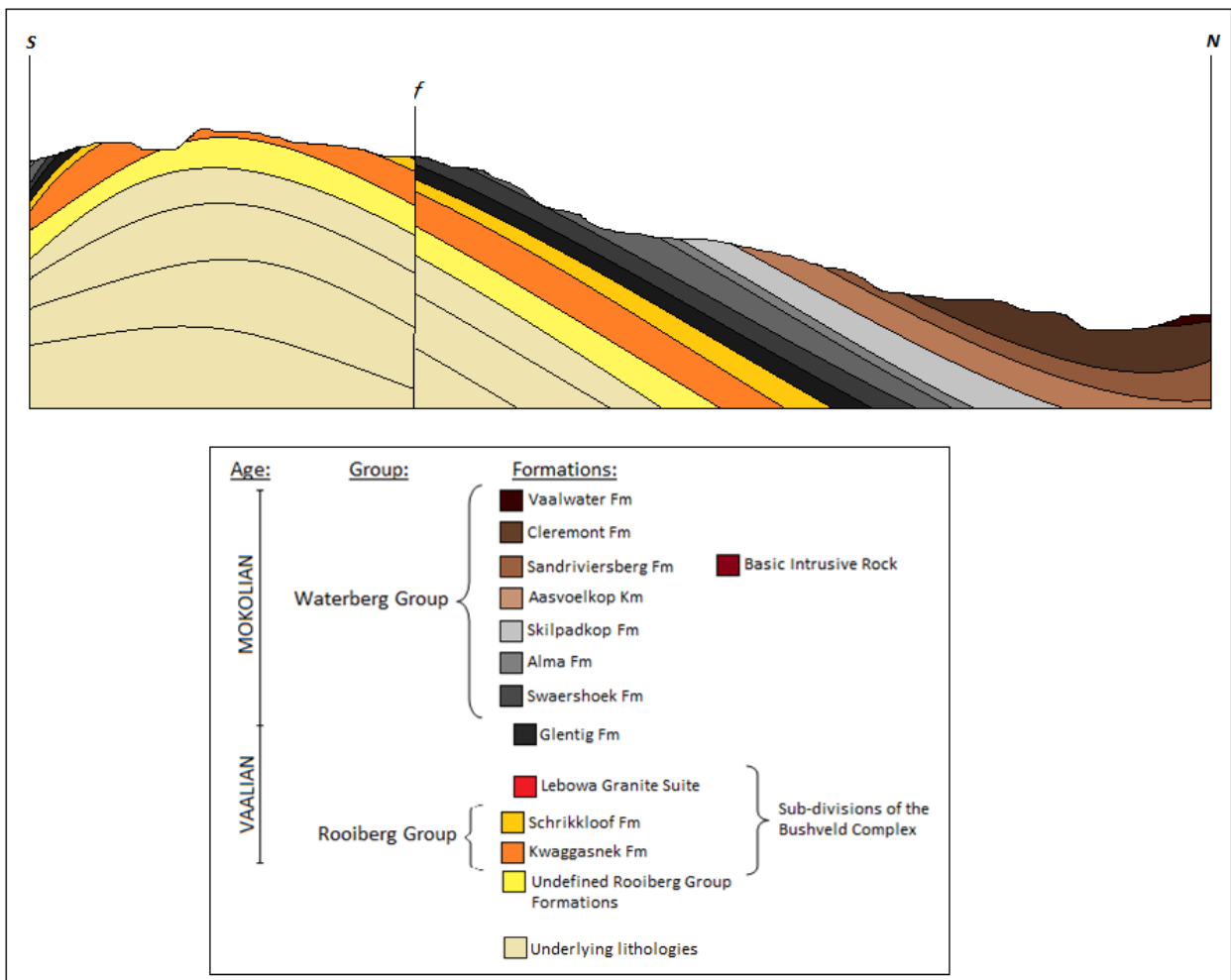


Figure 5-2-3 A conceptual geological cross-section of the A42 study area based on the outcrop geology

5.2.3. Hydrogeology

The topography of the A42 area (quaternary catchments A42A, A42B and A42C) varies between 1100mamsl to 1680mamsl above sea level. The western and eastern borders are mountainous with steeper topography that evens out towards the river valley in the centre (Figure 5-2-4). The topography to the southern border of the area indicates three river valleys which form one larger river valley towards the north which can be assumed to be the Moloko River. There is a general trend of decreasing topography from the south to the north. The groundwater level seems to follow the topography and there is a general gradient towards the river valley as well as towards the north (Figure 5-2-5).

The underlying Rooiberg Group lithologies within the study area consist of volcanic rocks, mostly rhyolite. Rhyolite is a fine-grained extrusive rock which would have little primary porosity due to the nature in which the rock is crystallised. Secondary features such as fractures would however allow for the movement of water through this lithology. The proto-Waterberg formation, Glentig, consists of mainly argillaceous sedimentary rocks which would also be an aquiclude or leaky aquifer based on the primary porosity of such lithologies. The overlying Waterberg Group lithologies on the other hand could provide good aquifers because the formations comprise of mostly arenites, rudites and arkoses which are sedimentary in nature.

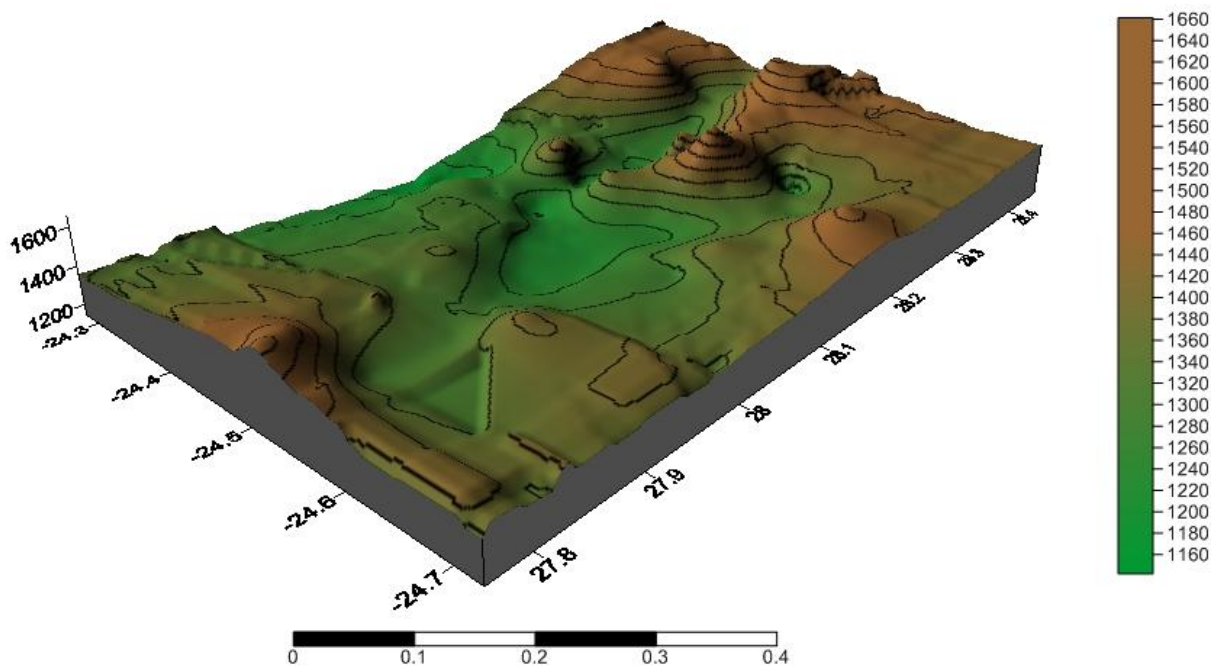


Figure 5-2-4 Generalised topography of the A42 area (quaternaries A42A – A42C).

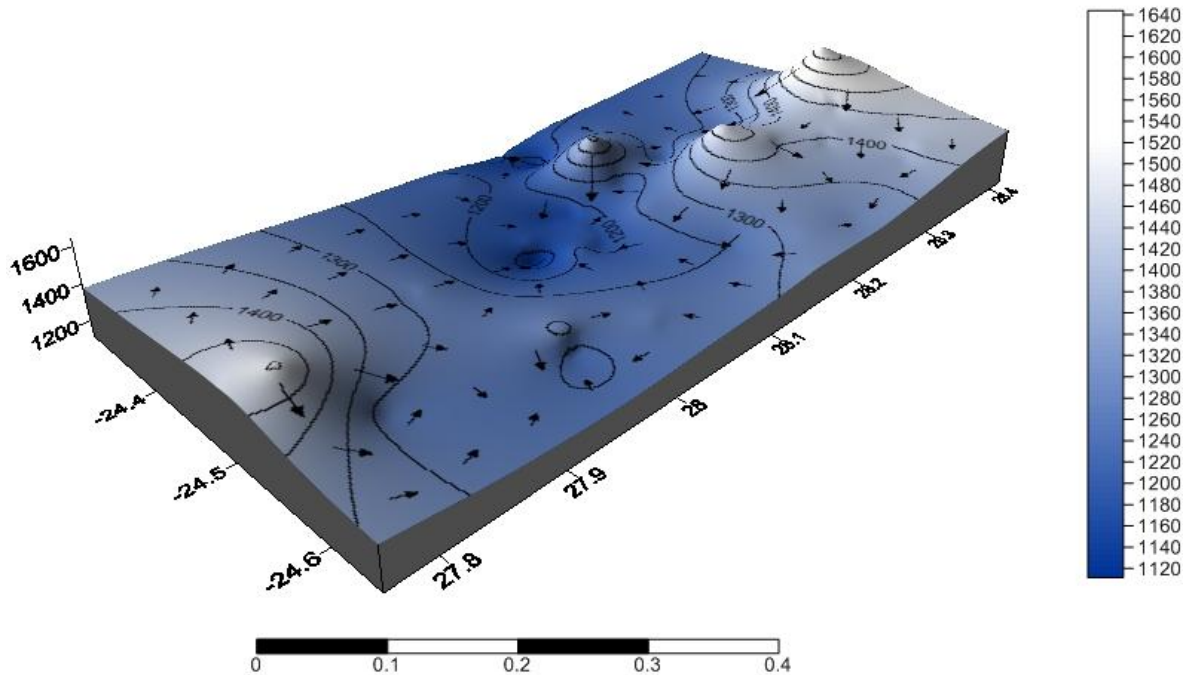


Figure 5-2-5 Generalised groundwater level within the A42 area based on data from 1993-1994.

5.2.4. Mixing Cell Model

The quaternary catchments A42A, A42B and A42C have an uneven distribution of river water and groundwater water quality data. The available groundwater quality data in the quaternary catchment A42B consists of a total of four borehole samples, one sample in 1979 and three in 1983 (Figure 5-2-6). The river water and groundwater quality data available for this time period in the other two quaternary catchments consists of one river sample point in A42C and one borehole sample in A42B. There are additional river water quality samples in the quaternary catchment A42C in 2005 as well as two borehole sample points, but there is no borehole or river water quality data in either A42A or A42B quaternary catchments during this time period (Figure 5-2-7). The MCM application is thus limited to a small area within the A42C quaternary catchment, where there is sufficient surface water and groundwater quality data from 2005.

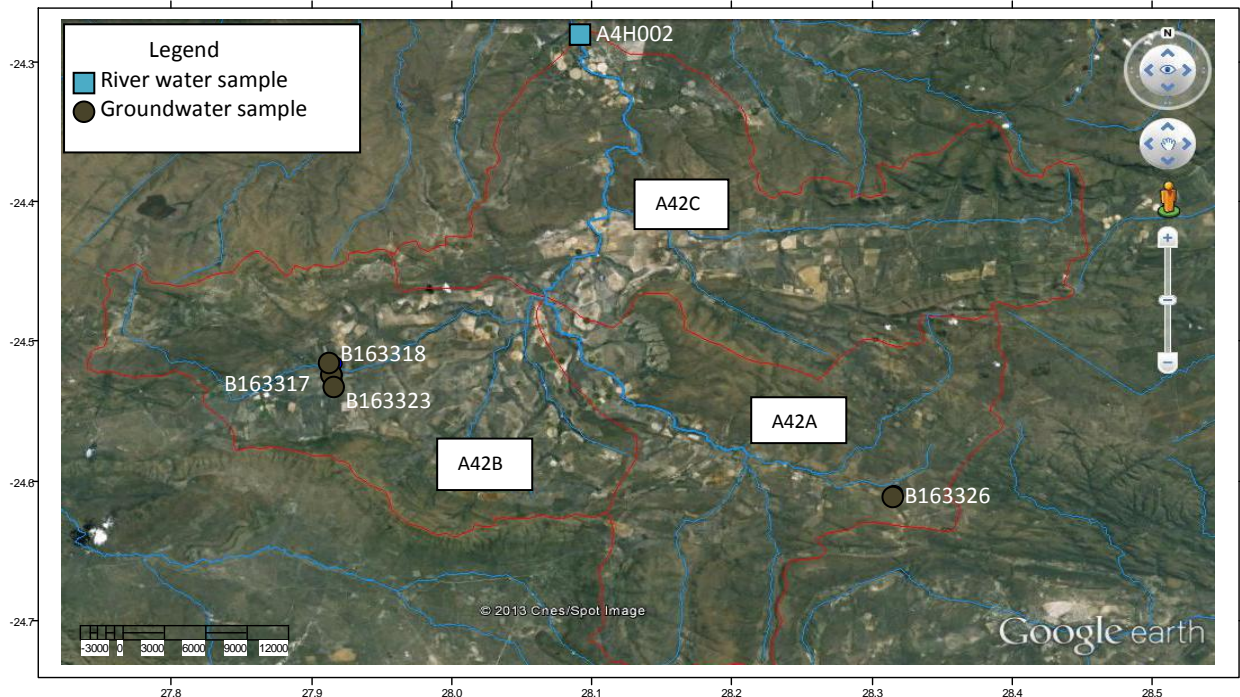


Figure 5-2-6 The river water and groundwater water quality data sample points for 1983 in the A42 area.

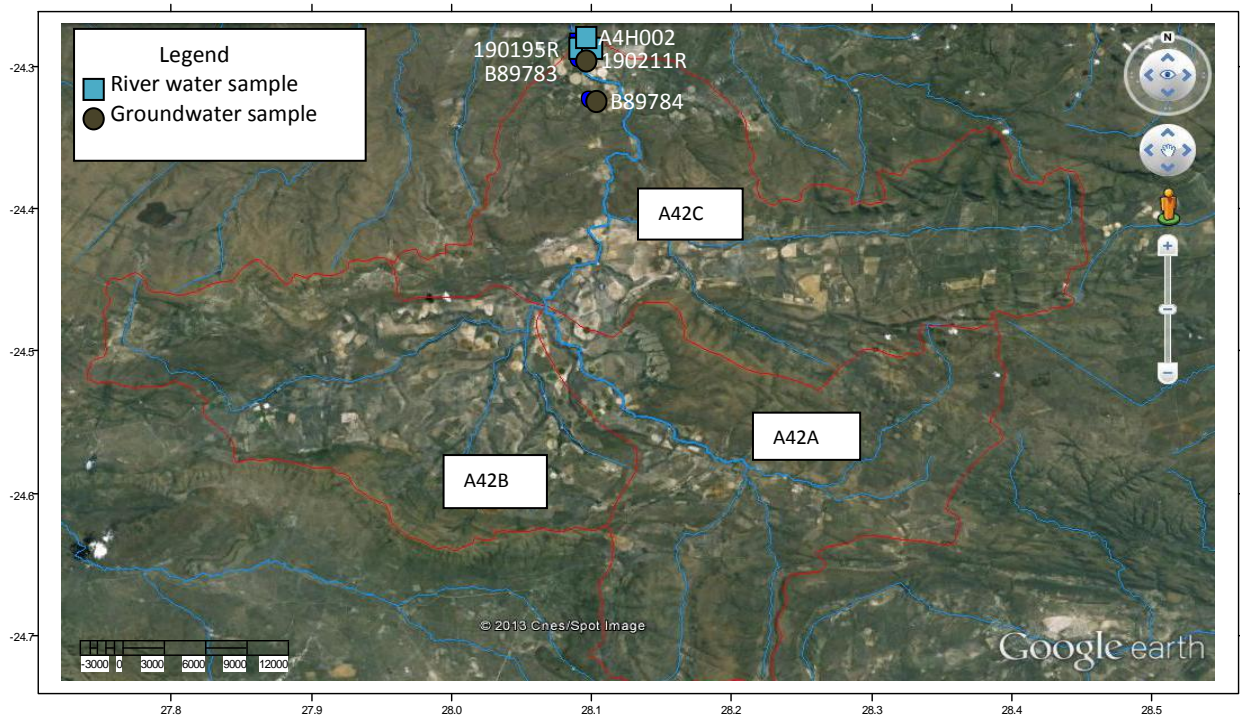


Figure 5-2-7 The river water and groundwater water quality data sample points for 2005 in the A42 area.

5.2.5. MCM Conceptualisation and application

The last segment of the Moloko River within the quaternary catchment A42C at the flowstation A4H002 is conceptualised as the river cell A4H2 Cell for the MCM application. The various flows into and out of this river system are conceptualised into a box model which forms the basis of the water balance equation for the MCM (Figure 5-2-8). A fictitious cell (o-A4H2 Cell) with exactly the same tracer concentrations as the river cell A4H2 Cell was created because the MCM software requires a minimum of two mixing cells to be defined in each model run. The A4H2 Cell is defined by a set of tracer concentrations from water quality data sampled at the flowstation A4H002. Inflow to the river cell is defined by water quality data also sampled at the flowstation due to data restrictions. Groundwater sources are defined by chemical analysis data for each of the 2 boreholes within the study area. The river flow measurements taken at the flowstation A4H002 are used to define the outflow from the flow system. All the water quality data used to define the various cells and inflows are shown in Appendix D Table D-1.

The representative EC and chloride (Cl) values for each flow represented in the flow box model are shown in Figure 5-2-9. Figure 5-2-9 shows a trend for EC and Cl values to decrease downstream. However, tracer concentrations for F, K, NO₃ and TAL increase downstream. The limited surface water quality data for this area has led to a poorly defined flow box model which is most likely the cause of the ambiguous chemical concentration trends. Groundwater is found to be of a much poorer quality when compared to surface water.

Abstraction and evaporation loss volumes from the section of the Moloko River under investigation as well as direct rainfall volumes into the system are assumed to be negligible because chemical data used in the model run covers a short time period and this omission ensures a mathematically-simple model run. A MCM model run is performed for each of the different weighting factor scenarios. Scenario 1 assigns the maximum weighting factor ($\omega = 1$) to all tracers, while scenario 2 assigns a lower weighting factor ($\omega = 0.3$) to tracers indicating a high chemical mass balance error (Table 5-2-2).

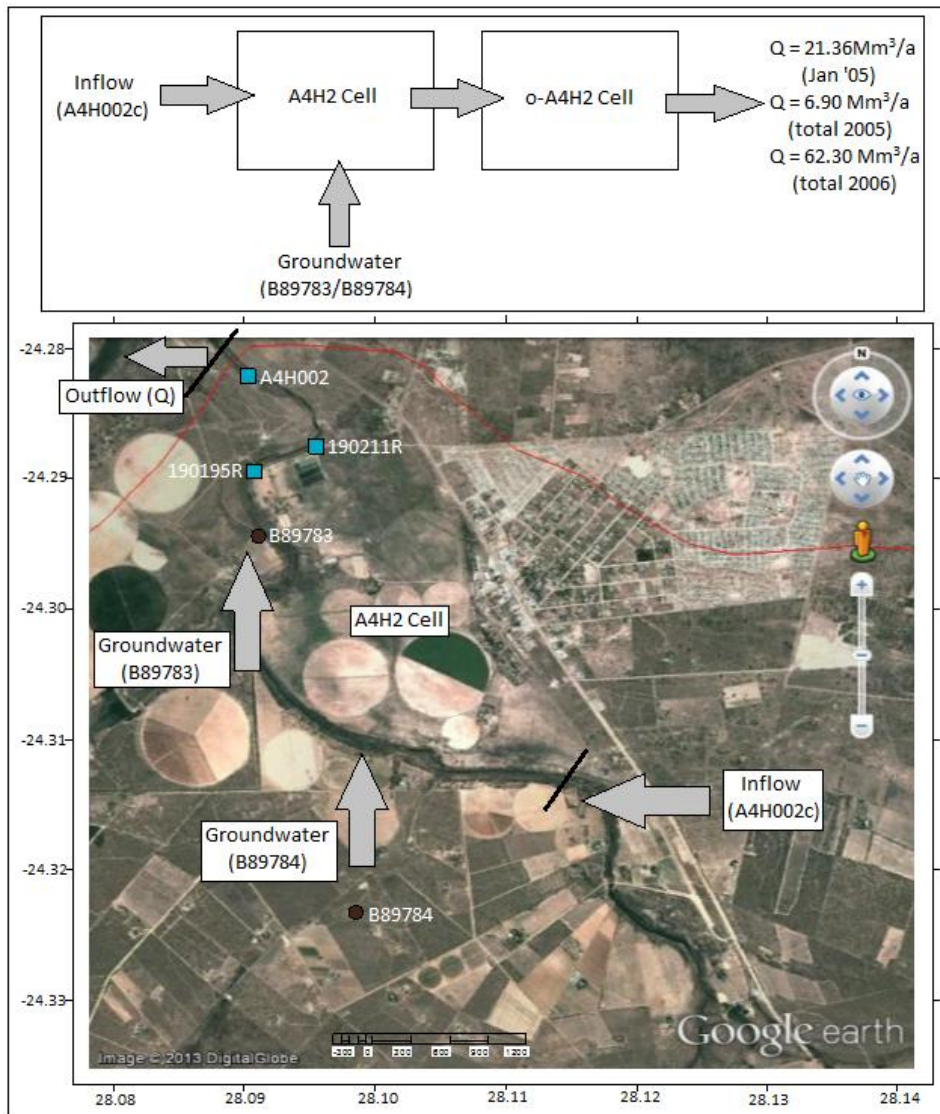


Figure 5-2-8 The box model conceptualisation of the various flows modelled for the A42 area and a Google Earth® image showing the conceptualisation on a real scale, indicating the position of the flowstation and borehole samples.

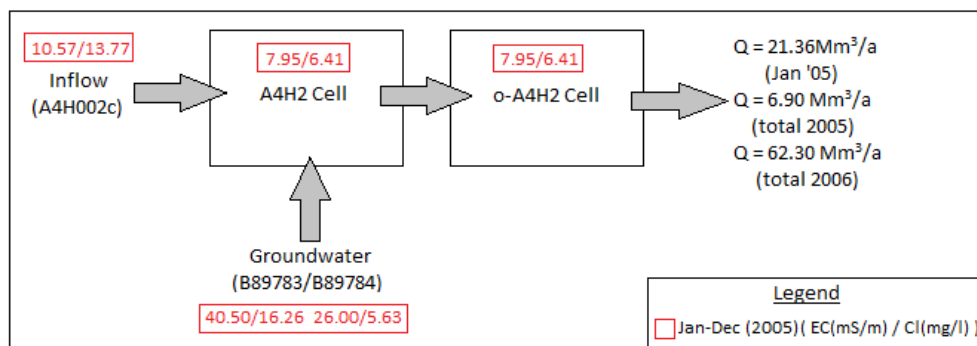


Figure 5-2-9 Representative EC and chloride (Cl) values for the A42 area flow conceptualisation. Concentration trends show a decrease in the EC-value and chloride concentration as moving downstream indicating that groundwater is not contributing to streamflow, but the area is in a high groundwater baseflow zone.

Table 5-2-2 Weighting factor assigned to each tracer for scenarios 1 and 2.

Weighting Factor Scenario	Tracer											
	Ca	Cl	EC	F	K	Mg	Na	NO3	pH	PO4	SO4	TAL
Scenario 1	1	1	1	1	1	1	1	1	1	1	1	1
Scenario 2	1	0.3	1	0.3	0.3	1	1	1	1	1	0.3	1

5.2.6. MCM Results

The MCM run, using scenario 1 weighting factors and defining the outflow from the system with the monthly river flow measured in January 2005 extrapolated to an annual flow volume, estimated the groundwater contribution to baseflow from the groundwater source defined by the borehole sample B89783 at $0.60\text{Mm}^3/\text{a}$, while the contribution from the source defined by the borehole sample B89784 was estimated at $1.31\text{Mm}^3/\text{a}$ (Table 5-2-3). The inflow to the A4H2 Cell, defined by the river sample A4H002c, was determined to be $11.69\text{Mm}^3/\text{a}$, while the outflow from the A4H2 Cell to the fictitious o-A4H2 Cell was determined to be $20.72\text{Mm}^3/\text{a}$. The MCM was run for the same setup, but using scenario 2 weighting factors, resulting in a estimation of the groundwater contribution to baseflow from the groundwater source defined by the borehole sample B89783 of $0.58\text{Mm}^3/\text{a}$, while the contribution from the source defined by the borehole sample B89784 was estimated at $0.63\text{Mm}^3/\text{a}$ (Table 5-2-3). The inflow to the A4H2 Cell, defined by the river sample A4H002c, was determined to be $13.70\text{Mm}^3/\text{a}$, while the outflow from the A4H2 Cell to the fictitious o-A4H2 Cell was determined to be $21.43\text{Mm}^3/\text{a}$.

The results from the A42 area MCM application, using scenario 2 weighting factors, are graphically represented in Figure 5-2-10. The unknown fluxes determined by the model are expressed as percentages of the assigned total outflow, which is the monthly flow volume measured at the flowstation in January 2005 extrapolated to an annual flow volume. The MCM application determined the inflow from the upstream river source A4H002c to the A4H2 Cell at 63.9% of the total outflow from the system. The groundwater contribution to streamflow within the A4H2 Cell was estimated at 5.6% from the defined groundwater sources. Flow from the A4H2 Cell to the downstream fictitious o-A4H2 Cell was determined at 100.3% of the assigned total flow. The outflow of more than 100% to the fictitious o-A4H2 Cell indicates that that the assigned total outflow volume could have been under-estimated.

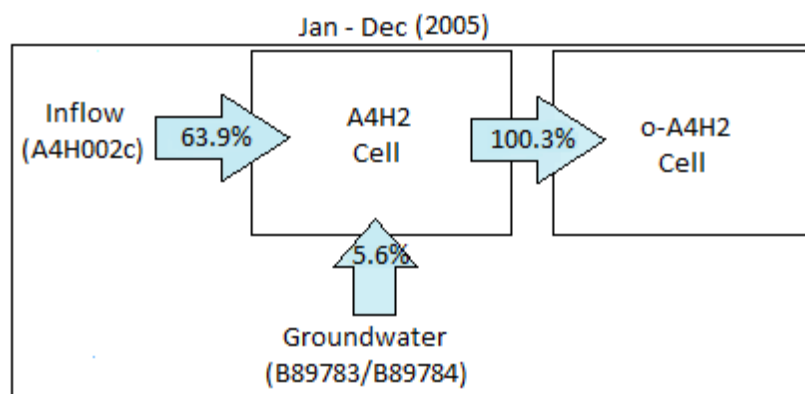


Figure 5-2-10 The unknown fluxes determined by the MCM for the A42 area application, making use of the scenario 2 weighting factors, expressed as a percentage of the total flow volume.

Two additional MCM runs were performed using two different volumes to define the outflow from the modelled system, to demonstrate the variability of the model. The total sum of the monthly river flow volumes measured at the flowstation A4H002 in 2005 and 2006 are used. The total sum of monthly flow volumes for 2005 is $6.9\text{Mm}^3/\text{a}$, while the total sum of monthly flow for 2006 is $62.30\text{Mm}^3/\text{a}$. The MCM run, using scenario 1 weighting factors and defining the outflow from the system with the total sum of monthly flow volumes from 2005, estimated the groundwater baseflow from the groundwater source defined by the borehole sample B89783 at $0.21\text{Mm}^3/\text{a}$, while the contribution from the groundwater source defined by the borehole sample B89784 was estimated at $0.46\text{Mm}^3/\text{a}$ (Table 5-2-3). Inflow to the A4H2 Cell, defined by the river sample A4H002c, was determined to be $4.13\text{Mm}^3/\text{a}$, while the outflow from the A4H2 Cell to the fictitious o-A4H2 Cell was determined to be $7.32\text{Mm}^3/\text{a}$. The MCM was run for the same setup, but making use of scenario 2 weighting factors, resulting in the groundwater baseflow from the groundwater source defined by the borehole sample B89783 to be estimated at $0.20\text{Mm}^3/\text{a}$, while the contribution from the groundwater source defined by the borehole sample B89784 was estimated at $0.22\text{Mm}^3/\text{a}$ (Table 5-2-3). The inflow to the A4H2 Cell, defined by the river sample A4H002c, was determined to be $4.84\text{Mm}^3/\text{a}$, while the outflow from the A4H2 Cell to the fictitious o-A4H2 Cell was determined to be $7.57\text{Mm}^3/\text{a}$.

The MCM run, using scenario 1 weighting factors and defining the outflow from the system with the total sum of monthly flow volumes in 2006, estimated the groundwater contribution to streamflow from the groundwater source defined by the borehole sample B89783 at $1.70\text{Mm}^3/\text{a}$, while the contribution from the groundwater source defined by the borehole sample B89784 was estimated at $3.72\text{Mm}^3/\text{a}$ (Table 5-2-3). The inflow to the A4H2 Cell, defined by the river sample A4H002c, was determined to be $33.09\text{Mm}^3/\text{a}$, while the outflow from the A4H2 Cell to the fictitious o-A4H2 Cell was determined to be $58.65\text{Mm}^3/\text{a}$. The MCM was run for the same setup, using scenario 2 weighting factors, resulting in a estimation of the groundwater contribution to streamflow from the groundwater source defined by the borehole sample B89783 of $1.63\text{Mm}^3/\text{a}$, while the contribution from the source defined by the borehole sample B89784 was estimated at $1.78\text{Mm}^3/\text{a}$ (Table 5-2-3). The inflow to the A4H2 Cell, defined by the river sample A4H002c, was determined to be $38.79\text{Mm}^3/\text{a}$, while the outflow from the A4H2 Cell to the fictitious o-A4H2 Cell was determined to be $60.66\text{Mm}^3/\text{a}$.

Table 5-2-3 Summary of MCM results for the A42 area model run for each scenario using three different outflows.

Name of inflow	Rate of inflow (Mm ³ /a)					
	Outflow (Jan '05)		Outflow (Sum 2005)		Outflow (Sum 2006)	
	Scenario 1	Scenario 2	Scenario 1	Scenario 2	Scenario 1	Scenario 2
A4H002c	11.69	13.70	4.13	4.84	33.09	38.79
B89783	0.60	0.58	0.21	0.20	1.70	1.63
B89784	1.31	0.63	0.46	0.22	3.72	1.78

5.2.7. MCM Discussion and comparison

The water and chemical mass balance errors for each model run and the associated weighting factor scenario are shown in Table 5-2-4. From Table 5-2-4 it can be seen that the use of different outflow volumes has no effect on the associated water and average chemical mass balance errors. A detailed, tracer-specific table of all the chemical mass balance percentage errors can be seen in Appendix D Table D-2. The change in outflow volumes does affect the estimated groundwater inflow, where a larger assigned outflow volume results in a higher groundwater baseflow estimate (Table 5-2-3). The water balance error is 39.2% using scenario 1 and 33.3% using scenario 2, for all the model runs. The chemical mass balance error is 12.7% using scenario 1 and 9.9% using scenario 2, for all the model runs (Table 5-2-4). The groundwater baseflow estimates from the MCM run using the weighting factors from scenario 2 are used for comparison because this scenario results in a decrease in both the water and chemical mass balance errors.

Table 5-2-4 Water and chemical mass balance percentage errors associated with each model run and scenario.

Balance	Outflow 1 (Jan '05)		Outflow 2 (Sum 2005)		Outflow 3 (Sum 2006)	
	Scenario 1	Scenario 2	Scenario 1	Scenario 2	Scenario 1	Scenario 2
Water balance error (%)	39.2%	33.3%	39.2%	33.3%	39.2%	33.3%
Chemical mass balance error (%)	12.7%	9.9%	12.7%	9.9%	12.7%	9.9%

The total groundwater baseflow was estimated at 1.21Mm³/a, 0.42Mm³/a and 3.41Mm³/a for the model runs using outflow 1, 2 and 3, respectively. Groundwater baseflow estimates from the Pitman, Hughes and Sami models as well as an estimate from another method incorporating water quality data are shown in Table 5-2-5. The Tracer method estimate is shown for both the use of electrical conductivity (EC) as the sole tracer, as well as an average value from applying the Tracer method with a number of tracers. The Sami and Hughes models were calibrated for quaternary catchments A42A and A42B, while an estimated A42C calibration volume is inferred from the ratio between calibrated and uncalibrated volumes. The MCM and Tracer method were only applied within the A42C quaternary catchment due to data restrictions. The

uncalibrated Pitman, Sami and Hughes models groundwater baseflow estimates for the quaternary catchment A42A are 17.99Mm³/a, 8.70Mm³/a and 15.62Mm³/a, respectively (Table 5-2-5). The calibrated groundwater baseflow estimates for the Sami and Hughes models are 8.99Mm³/a and 11.90Mm³/a, respectively. For the quaternary catchment A42B the uncalibrated Pitman, Sami and Hughes models groundwater baseflow estimates are 16.86Mm³/a, 6.56Mm³/a and 15.39Mm³/a, respectively. The calibrated groundwater baseflow estimates for the Sami and Hughes models are 8.50Mm³/a and 11.72Mm³/a, respectively. The uncalibrated Pitman, Sami and Hughes model groundwater baseflow estimates for the quaternary catchment A42C are 22.41Mm³/a, 8.83Mm³/a and 20.26Mm³/a, respectively while the inferred calibrated volumes for the Sami and Hughes model are 10.28Mm³/a and 15.43Mm³/a, respectively. The Tracer method groundwater baseflow estimate for the quaternary A42C, using only EC is 9.83Mm³/a, while for an average from a number of tracers is 20.83Mm³/a. The MCM groundwater baseflow estimate for A42C was 1.21Mm³/a for outflow 1, 0.42Mm³/a for outflow 2 and 3.41Mm³/a for outflow 3, based on data from the 2005 time period.

The trend seen in these results for the Pitman, Sami and Hughes models is for the groundwater baseflow estimated volume to decrease from the Pitman model to the Hughes model to the Sami model, with the difference between the Pitman model and the Hughes model being less dramatic than between the Hughes and Sami model. The smaller difference between the Hughes and Pitman models is expected because the Hughes model is a modified version of the Pitman model. However, the difference between the Sami and Hughes models is fairly substantial. This difference becomes less after calibration with the Sami model groundwater baseflow estimate increasing after calibration and the Hughes model estimate decreasing after calibration. Considering these changes after calibration, the initial Sami model groundwater baseflow volume is an under-estimation of the groundwater baseflow, while the initial Hughes model volume is an over-estimation. Figure 5-2-11 is a graphical representation of the estimates for each method, illustrating the discussed trends. The Tracer method groundwater baseflow estimates seems to be in agreement with the general amount of groundwater contributing to the river, with the EC-only estimate similar to that of the Sami model and the multiple-tracer estimate similar to that of the Hughes model initial estimate.

Table 5-2-5 The groundwater baseflow estimates from the Pitman model, Sami model (calibrated and uncalibrated), Hughes model (calibrated and uncalibrated), Tracer method and MCM for the quaternary catchments A42A – A42C.

Quat	Pitman Uncalibrated (Mm ³ /a)	Sami Uncalibrated (Mm ³ /a)	Sami Calibrated (Mm ³ /a)	Hughes Uncalibrated (Mm ³ /a)	Hughes Calibrated (Mm ³)	MCM (Mm ³ /a)			Tracer Method (Mm ³ /a)	
						Outflow 1	Outflow 2	Outflow 3	EC	Average
A42A	17.99	8.70	8.99	15.62	11.90	/	/	/	/	/
A42B	16.86	6.56	8.50	15.39	11.72	/	/	/	/	/
A42C	22.41	8.83	10.28 (E)	20.26	15.43 (E)	1.21	0.42	3.41	9.83	20.83

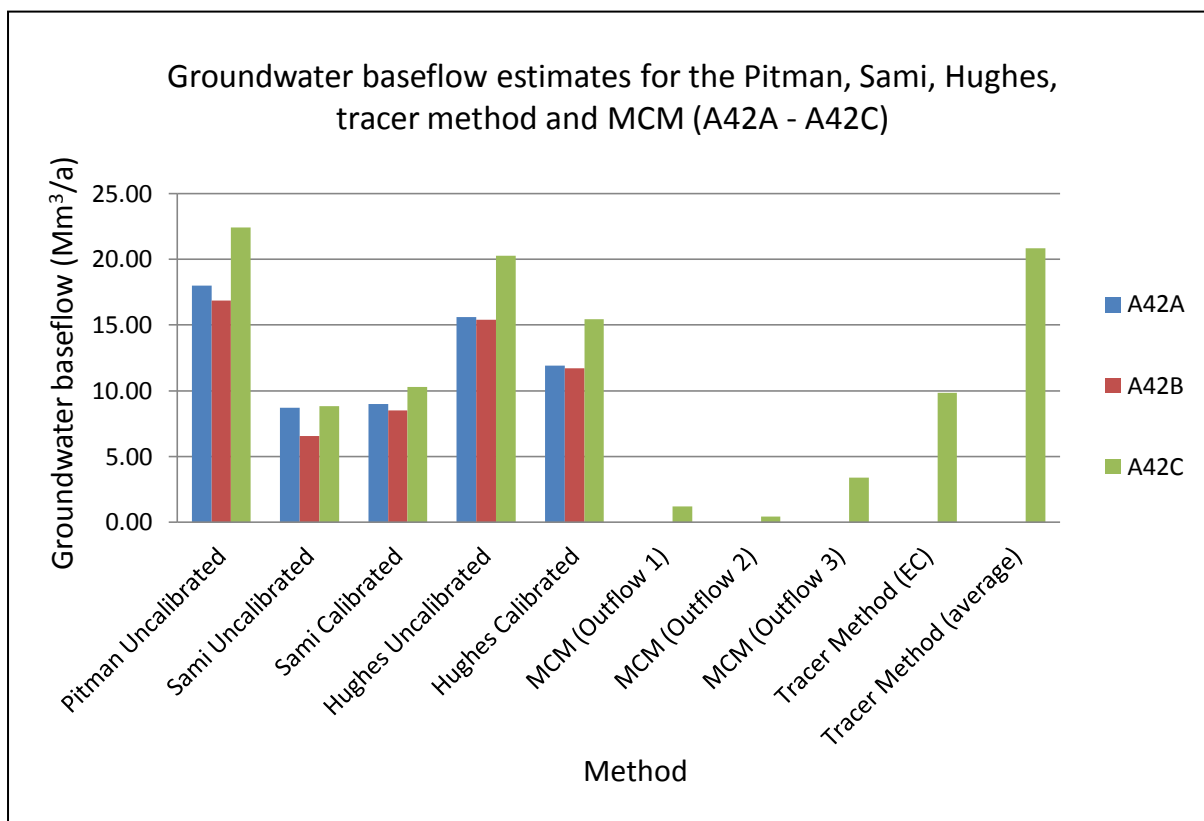


Figure 5-2-11 The groundwater baseflow estimates from the Pitman model, Sami model (calibrated and uncalibrated), Hughes model (calibrated and uncalibrated), Tracer method and MCM for the quaternary catchments A42A – A42C.

The groundwater baseflow estimates from the MCM are considerably lower than all the other methods. One explanation for the lower groundwater contribution estimated by the MCM is that the data used in the model was collected during a particularly dry year with the rainfall being less than that of the corresponding month in 2006. The measured river flow volumes measured at A4H002 in 2005 show a drastic difference when compared with the corresponding monthly flow volumes in 2006 (Figure 5-2-12). The difference in rainfall is not as drastic as the difference in measured flow volumes, which might indicate that the low river flow is not related to the amount of rainfall, but perhaps the amount of water use. If groundwater resources were

over utilised along with the surface water resources this would lead to a drastic decrease in the amount of groundwater reaching the river. The MCM run using the outflow volume measured in 2006 still results in a low groundwater baseflow estimate which could mean that the chemical data is restricting, or that the low flow in 2005 is not the reason for the low estimate. Another explanation is the rough definition of the inflow sources and the river cell due to a general lack of water quality data. The MCM results also have a large percentage error associated with the estimates which could indicate that an important source has been omitted or that the river cell was incorrectly defined.

Considering the results from this study area, the Tracer method seems to be better suited for quantifying the groundwater baseflow volume than the MCM. The Tracer method estimate from the use of EC alone is in agreement with the calibrated Sami model estimate and the estimate from multiple tracer use is in agreement with the initial Hughes model estimate. In this instance, it could be deduced that the average Tracer method estimate from a number of tracers is an over-estimation of the groundwater baseflow, as is seen with the initial Hughes model estimate. The Tracer method estimate from the use of EC alone gives the most reasonable estimate of groundwater baseflow, between the two water quality methods.

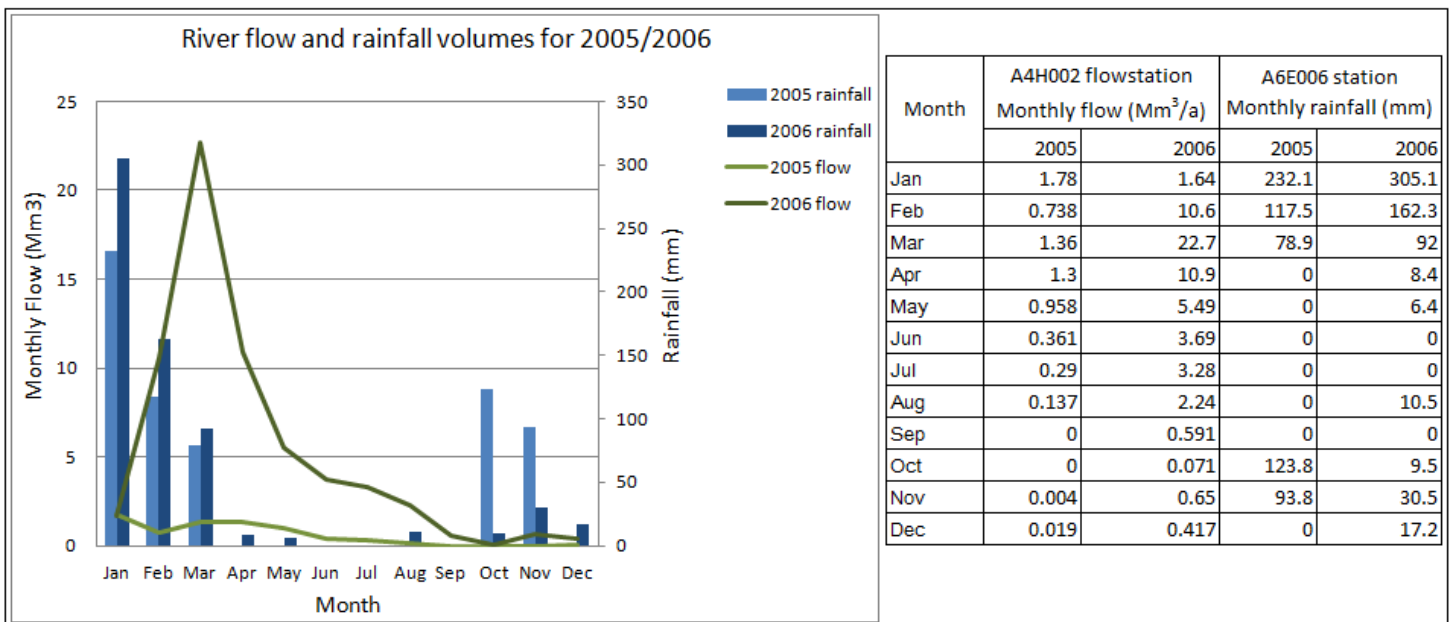


Figure 5-2-12 Rainfall at the meteorological station A6E006 and river flow volumes at the flowstation A4H002 for 2005 and 2006.

5.3. Pilot Study 3: Quaternary Catchment D73F

5.3.1. Overview

The quaternary catchment D73F is situated within the Northern Cape Province of South Africa (Figure 5-3-1). This area falls within the Lower Orange Water Management Area and includes a section of the Orange River between the towns of Upington and Kakamas. The Orange River is a perennial river and the longest in South Africa. It is fed by a number of non-perennial tributaries within the study area, namely the Neuspruit, Kameel, Brak, Vaalputs, Kareeboom, Olienhout and Donkerhoek Rivers. The average annual rainfall, based on data from the Department of Water Affairs meteorological station D7H003, is 190mm, and the total measured rainfall for the year of investigation (1989) was 136mm. The average annual evaporation, based on S-pan measurements from the meteorological station, is 2527mm. The average monthly temperature for the area ranges from 4°C to 35°C. The area is classified as a semi-desert and characterised by a flat landscape. The topography ranges from 700mamsl to 1000mamsl, with no drastic changes in topography. The predominant vegetation is Nama Karoo which consists of low shrubs and grasses, with the northern point of the quaternary catchment D73F dominated by Southern Kalahari vegetation. The WARMS database indicates that most of the surface water abstraction is for irrigation schemes and amounts to approximately 226Mm³/a.

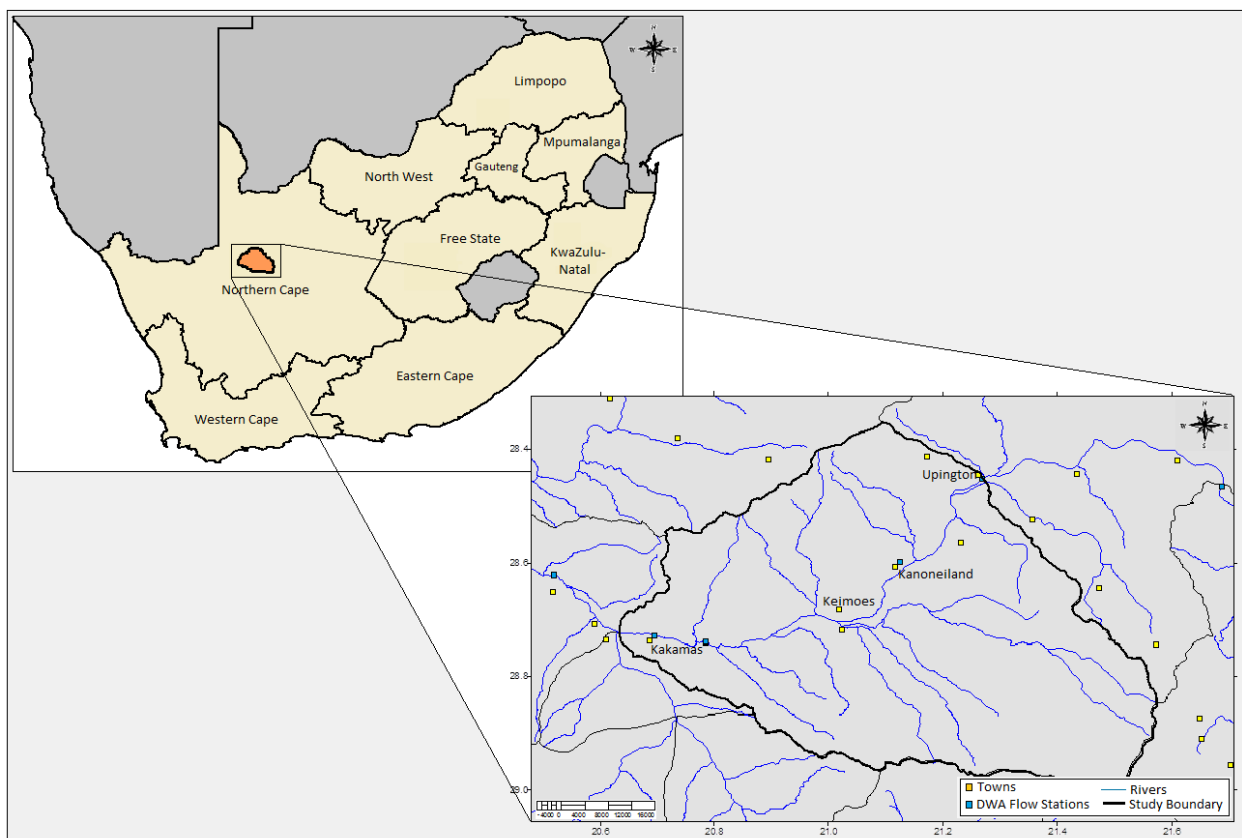


Figure 5-3-1 The position of the quaternary catchment D73F within South Africa.

5.3.2. Geology

The Namaqua-Natal Metamorphic Province covers most of the Northern Cape and the quaternary catchment under investigation falls within the Namaqua sector of this tectono-stratigraphic province. The Namaqua-Natal Province comprises a number of tectono-stratigraphic terranes bounded by shear zones. The five main subdivisions in the Namaqua sector are the Richtersveld Subprovince, the Bushmanland Terrane, Kakamas Terrane, Areachap Terrane and the Kaaien Terrane (Johnson, Anhaeusser and Thomas, 2006). Formations from the Kakamas, Areachap and Bushmanland Terranes are found in the outcrop geology of the quaternary catchment D73F (Figure 5-3-2). The outcrop geology of the area indicates the structural, metamorphic and intrusive complexity of the area which leads to uncertainty in regards to stratigraphic relations. The Bushmanland Terrane is the largest of the crustal blocks in the Namaqua sector, and represented in the quaternary catchment by the supracrustal Hoogoor Suite and intrusive rocks of the Little Namaqualand Suite and the Eendoorn Granite. The Kakamas Terrane lies to the East of the Bushmanland Terrane and is represented within the study area by the supracrustal Korannaland Group, Witwater Gneiss and Toeslaan Formation and intrusive rocks of the Eendoorn Suite, Keimos Suite and Friersdale Charnockite. The Bethesda and Jannelsepan Formations of the Areachap Terrane are found in the outcrop geology of the area as well as intruded Keimos Suite granites (Johnson, Anhaeusser and Thomas, 2006).

The Hoogoor Suite of the Bushmanland Terrane, often referred to as the pink gneiss, is a large concordant to semi-concordant body of red-weathering quartzofeldspathic gneisses. The intrusive Little Namaqua Suite consists of igneous lithologies ranging from granite to rocks with sheet-like bodies of quartz-microcline-biotite augen gneiss. The Eendoorn Granite resembles the augen gneiss in the Little Namaqua Suite, but geochronological restraints are currently insufficient to confirm a direct relation. The Korannaland Group of the Kakamas Terrane comprises several lithologies, which can be generally described as high-grade supracrustal rocks consists of mostly arenite and calc-arenite lithologies. Witwater Gneiss rocks are also found within this Terrane. The Toeslaan Formation is included into the Korannaland Group, but its stratigraphic position is uncertain and might be correlated to the Areachap Terrane. The formation consists of a thick succession of quartz-feldspar-cordierite-spinel-garnet meta-pelitic gneisses. The Kakamas Terrane is intruded by syn- to late-tectonic granitoids, including the Eendoorn Suite. The Keimos Suite, a collection of the syn- to post-tectonic granitoids east of the

Neusberg Shear Zone, is also an intrusive rock of the Kakamas Terrane. The Friersdale Charnockite consists of late-tectonic bodies of undeformed lithologies which intrude earlier members of the Keimoes Suite and has been interpreted as the intrusive equivalent of the Jannelsepan Formation of the adjacent Areachap Terrane. The Jannelsepan Formation is found in the northern part of the Areachap Group and consists of mainly migmatitic amphibolite and calc-silicate rocks. The Bethesda Formation occurs along the western side of the Group forming metapelitic schist similar to that of the Sprigg Formation, but devoid of a conglomerate component and partly interlayered with amphibolites. The Keimoes Suite and the charnockite intrusive of the Friersdale Charnockite intrude into the Areachap Terrane (Johnson, Anhaeusser and Thomas, 2006).

The Kubis Sub-group of the Nama Group from the Namibian successions, outcrops in a small section of the study area near Upington. The sub-group consists of alternating layers of sandstone and shale. Cenozoic age sediments of the Kalahari Group are also evident in the outcrop geology of D73F. The Kalahari Group is the most extensive body of Cenozoic age terrestrial sediments and has been divided into a number of formations, ranging from clayey gravels to calcretes. The Kalahari group deposits tend to coincide with the occurrence of Dwyka Group rocks at their thickest parts, which is seen in the study area (Johnson, Anhaeusser and Thomas, 2006).

The geology within quaternary catchment D73F is highly complicated which makes an accurate stratigraphic column or conceptual cross-section impractical and is thus not attempted.

5.3.3. Hydrogeology

The topography of the quaternary catchment D73F ranges from 700mamsl to 1000mamsl. There is a gradual decrease in topography towards the river valley as well as a gradual decrease in topography from east to west along the Orange River (Figure 5-3-3). The water table seems to follow the topography, but the groundwater level is far below the terrestrial surface (Figure 5-3-4). The geology of the area is highly complicated with multiple intrusions, metamorphosed lithologies and extreme deformation. This complex geology tends to limit the existence of continuous aquifers. The amount of groundwater that would have a direct path towards the river through a complicated geology as shown here is negligible. The groundwater level is far below the land surface which also limits the probability of groundwater contributing to the river.

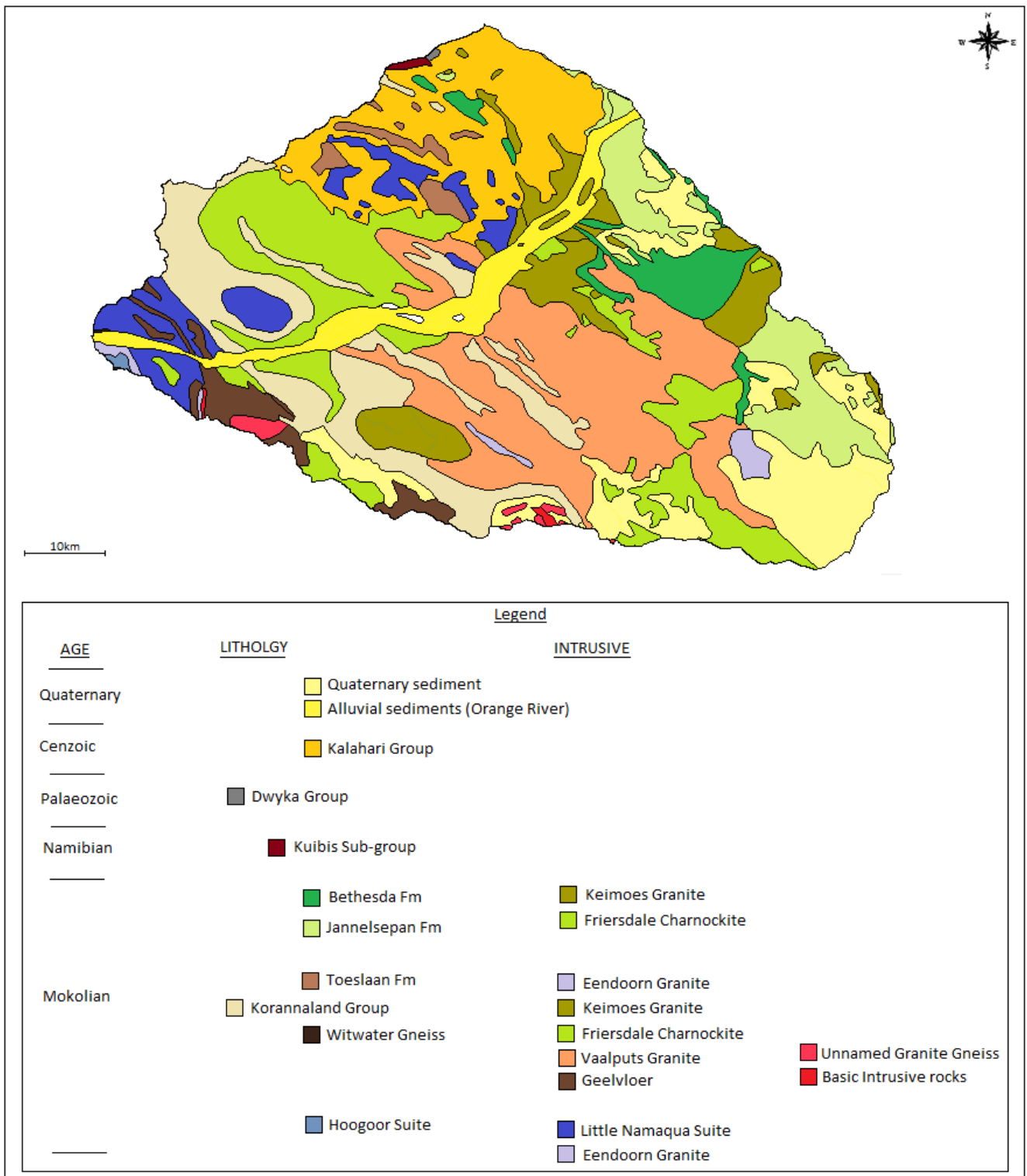


Figure 5-3-2 The outcrop geology of the quaternary catchment D73F (Based on data from GRDM, 2010).

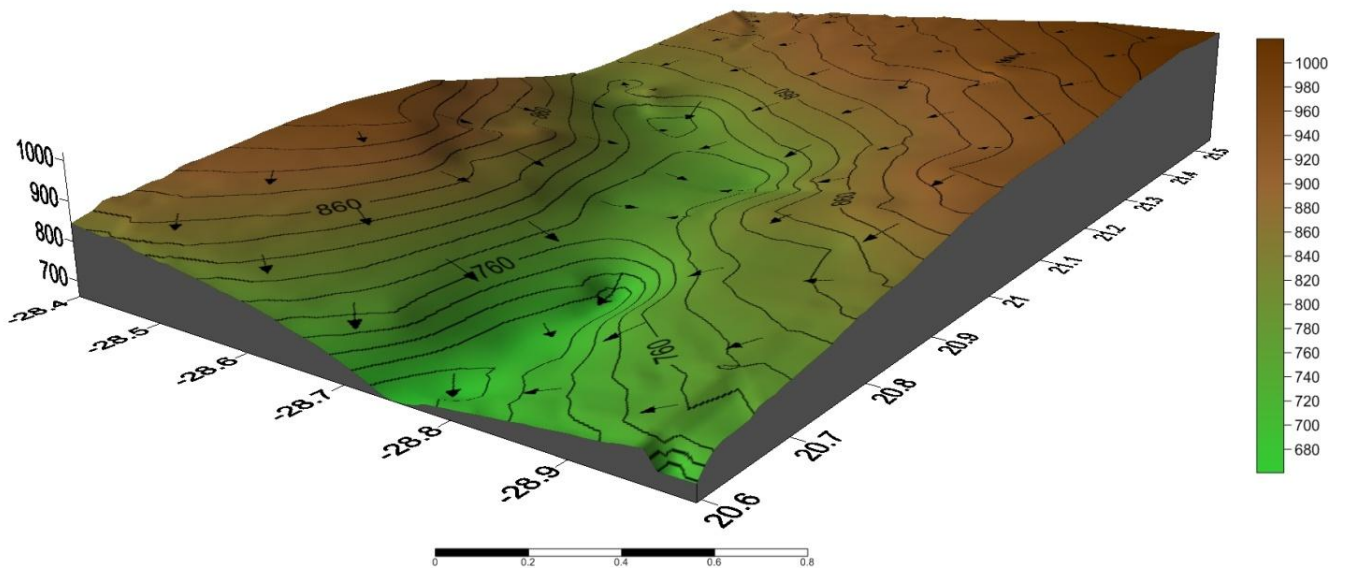


Figure 5-3-3 Generalised topography of the quaternary catchment D73F

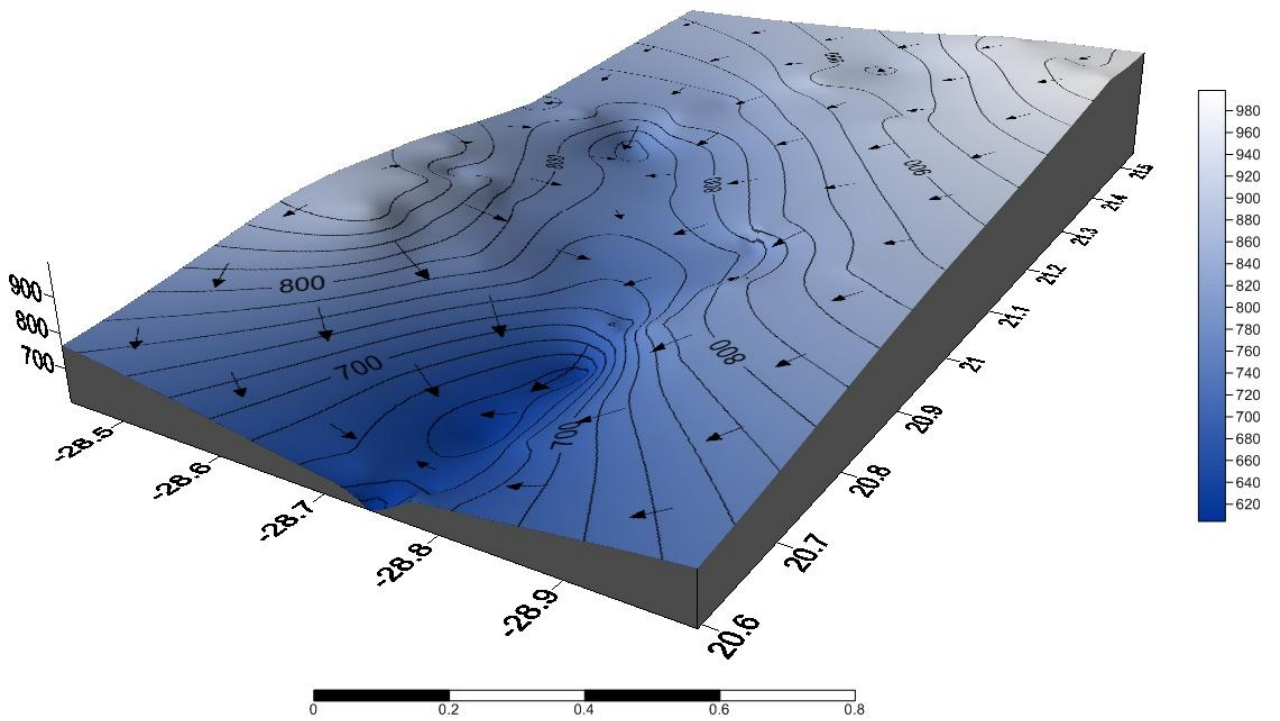


Figure 5-3-4 Generalised groundwater level within the quaternary catchment D73F area.

5.3.4. Mixing Cell Model

The MCM does not directly use physical, geohydrological principles in estimating the groundwater component of baseflow, but estimates the groundwater baseflow by means of a mathematical algorithm. The MCM thus has the potential to allocate a fictitious groundwater baseflow volume to an area that in reality receives no groundwater baseflow. The quaternary catchment D73F falls in the middle of a regionally-defined zero groundwater baseflow zone. The application of the MCM to this catchment will be useful to investigate how the MCM will perform when groundwater sources to a system are defined in the MCM, but it is known that these sources are not contributing to the system.

5.3.5. MCM Conceptualisation and application

The flowstation D7H003 at the downstream boundary of the quaternary catchment D73F would be the best sample point for surface water quality data to define the river mixing cell. Water quality data for this station is only available from 1965 to 1993. Additional data for a MCM model run is thus selected from within this time period as to allow for the use of the chemical data from flowstation D7H003. Additional surface water and groundwater quality data was the most suitable during the year 1989, resulting in the MCM run making use of surface water and groundwater data from this time period.

The section of the Orange River within the quaternary catchment D73F is conceptualised as the river cell D7H003 Cell for the MCM run. The various flows into and out of this river system are conceptualised into a box model which forms the basis of the water balance equation in the MCM (Figure 5-3-5). The MCM requires a minimum of two cells to be defined, thus a fictitious cell (o-D7H003 Cell) is created with exactly the same tracer concentrations as the river cell D7H003 Cell. The D7H003 Cell is defined by a set of tracer concentrations from water quality data sampled at the flowstation D7H003. Groundwater sources are defined by the chemical analysis data from each of the 14 sampled boreholes within the study area over the time period of investigation. The river flow measurements taken at the flowstation D7H003 are used to define the outflow from the flow system. A monthly flow volume is used to define the outflow from the system because the yearly volume of water passing point this point is extremely large. The monthly flow volumes are extrapolated to yearly flow volumes after computation. All the water quality data used to define the various cells and inflows are shown in Appendix E Table E-1.

The representative EC and chloride (Cl) values for each flow represented in the flow box model are shown in Figure 5-3-6. Figure 5-3-6 shows a trend for the EC value to increase downstream. However, the Cl concentration decreases downstream. The EC trend could indicate that groundwater or some poorer quality water is reaching the stream, while the Cl concentration trend could indicate that no groundwater is contributing to the stream or some lower Cl concentration source of water is reaching the stream. Groundwater was found to be of a poorer quality than surface water, but large ranges of tracer concentrations are found between the various boreholes.

Abstraction and evaporation loss volumes from the section of the Moloko River under investigation as well as direct rainfall volumes are assumed to be negligible for the initial MCM application, but an abstraction volume is assigned to the river cell in an additional model run. Both model runs were applied using different weighting factor scenarios. The first scenario (scenario 1) assigns the maximum weighting factor to all tracers ($\omega = 1$), while scenario 2 assigns a lower weighting factor ($\omega = 0.3$) to tracers showing a high chemical mass balance error (Table 5-3-1).

Table 5-3-1 The weighting factor assigned to each tracer for scenarios 1 and 2.

Weighting Factor Scenario	Tracer														
	Ca	Cl	DMS	EC	F	K	Mg	Na	NH4	NO3_NO2	pH	PO4	Si	SO4	TAL
Scenario 1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1
Scenario 2	1	1	1	1	1	1	1	0.3	0.3	0.3	1	1	1	1	0.3

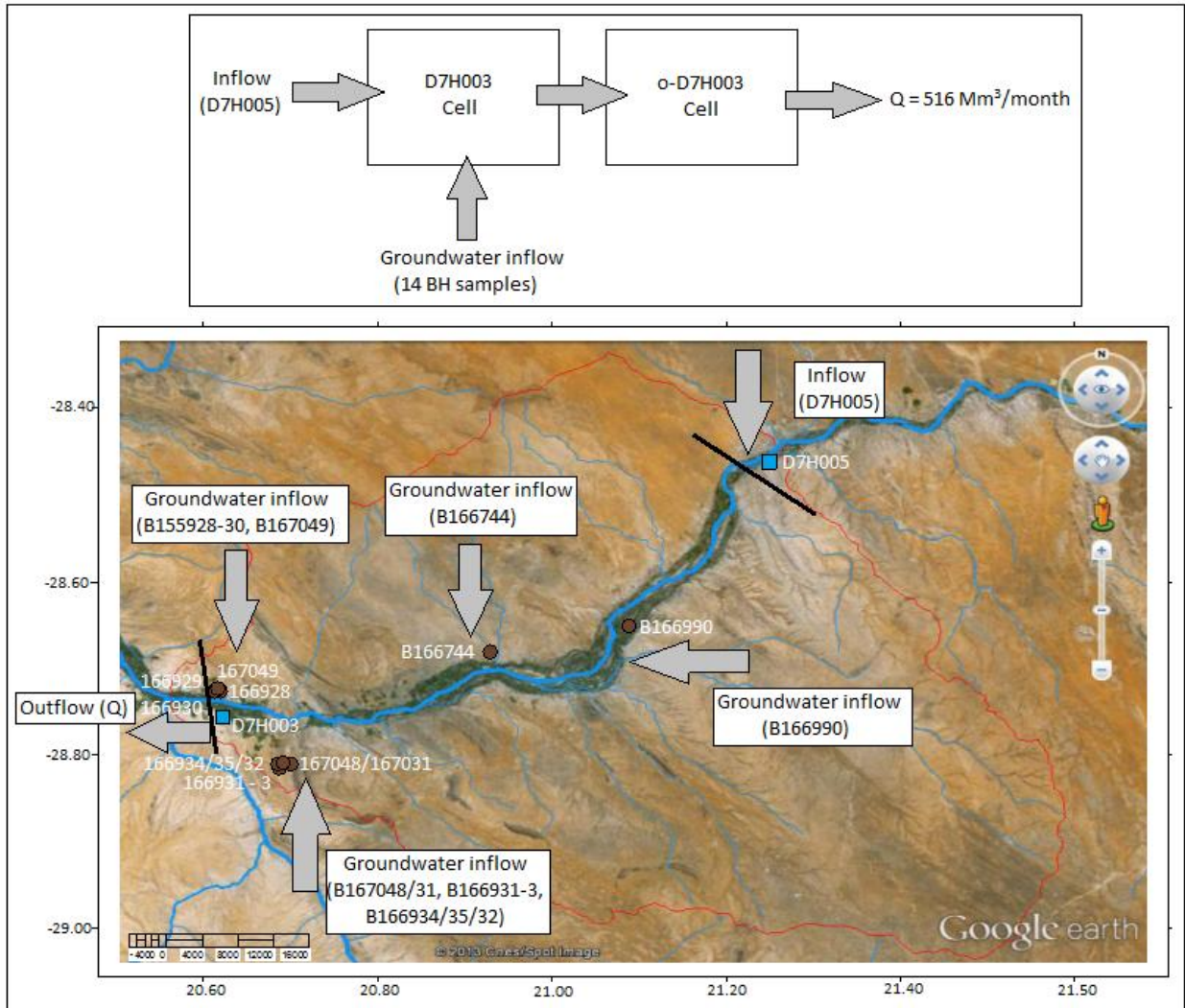


Figure 5-3-5 The box model conceptualisation of the various flows modelled for the quaternary catchment D73F area and a Google Earth® image showing the conceptualisation on a real scale indicating the position of the flowstations (D7H003 and D7H005) and the 14 borehole sample points.

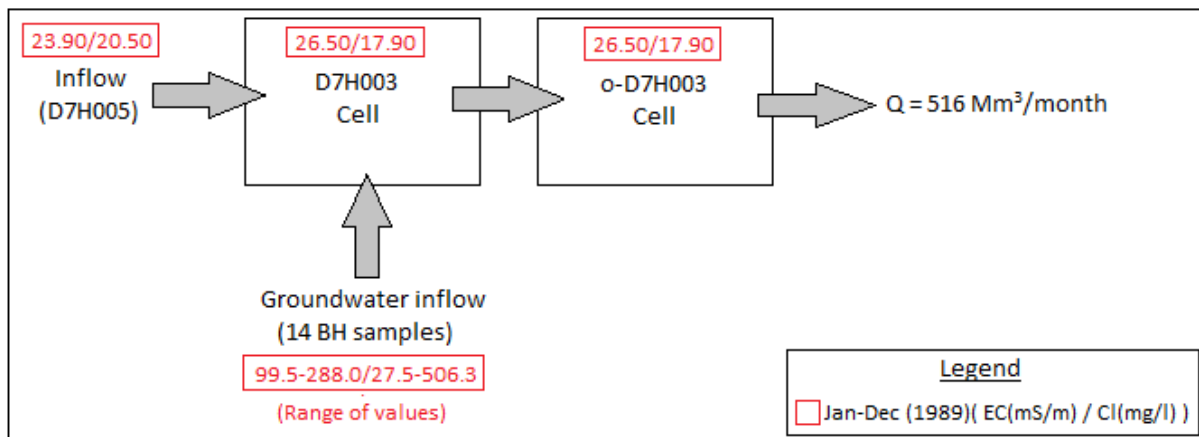


Figure 5-3-6 Representative EC and chloride (Cl) values for the quaternary catchment D73F flow conceptualisation. Concentration trends show an increase in the EC-values, but a decrease in the Cl concentrations moving downstream.

5.3.6. MCM Results

The MCM run, using scenario 1 weighting factors and assigning no abstraction volume to the river cell, estimated the groundwater contribution to baseflow from the groundwater sources defined by the borehole samples B166990 and B166744, located in the middle of the river cell, at $0.00\text{Mm}^3/\text{month}$ (Table 5-3-2). The groundwater contribution to baseflow from the groundwater sources located on the northern side of the river and defined by the borehole samples B166928, B166929, B166930 and B167049 were all estimated at $0.00\text{Mm}^3/\text{month}$. The groundwater contribution to baseflow from the groundwater sources located on the southern side of the river and defined by the borehole samples B166931, B166932, B166933, B166934, B166935, B167031 and B167048 were all estimated at $0.00\text{Mm}^3/\text{month}$, while the groundwater source defined by the borehole sample B167032 was estimated at $0.62\text{Mm}^3/\text{month}$, which extrapolated to a yearly flow volume becomes $7.44\text{Mm}^3/\text{a}$. The inflow to the D7H003 Cell, defined with water quality data from the flowstation D7H005, was determined to be $430.35\text{Mm}^3/\text{month}$ which extrapolated to a yearly flow volume becomes $5164.20\text{Mm}^3/\text{a}$, while the outflow from the D7H003 Cell to the fictitious o-D7H003 Cell was determined to be $487.32\text{Mm}^3/\text{month}$ which extrapolated to a yearly flow volume becomes $5847.84\text{Mm}^3/\text{a}$. The MCM was run for the same setup, using scenario 2 weighting factors, resulting in an estimation of the groundwater contribution to baseflow, from all the defined groundwater sources, at $0.00\text{Mm}^3/\text{a}$ (Table 5-3-2). The inflow to the D7H003 cell, defined with water quality data from flowstation D7H005, was determined to be $479.73\text{Mm}^3/\text{month}$ or $5756.76\text{Mm}^3/\text{a}$, while the outflow from the D7H003 Cell to the fictitious o-D7H003 Cell was determined to be $506.22\text{Mm}^3/\text{month}$ or $6074.64\text{Mm}^3/\text{a}$.

The results from the quaternary catchment D73F MCM application, utilizing scenario 2 weighting factors and assigning no abstraction volume to the river cell, are graphically represented in Figure 5-3-7. The unknown fluxes determined by the model are expressed as percentages of the assigned total outflow. The MCM application determined the inflow from the upstream river source, defined by a water sample taken at the flowstation D7H005, to the D7H003 Cell at 94.7% of the total outflow from the system. The groundwater contribution to streamflow within the D7H003 Cell was estimated at 0.0% from the defined groundwater sources. Flow from the D7H003 Cell to the downstream fictitious o-D7H003 Cell was determined at 98.1% of the total flow.

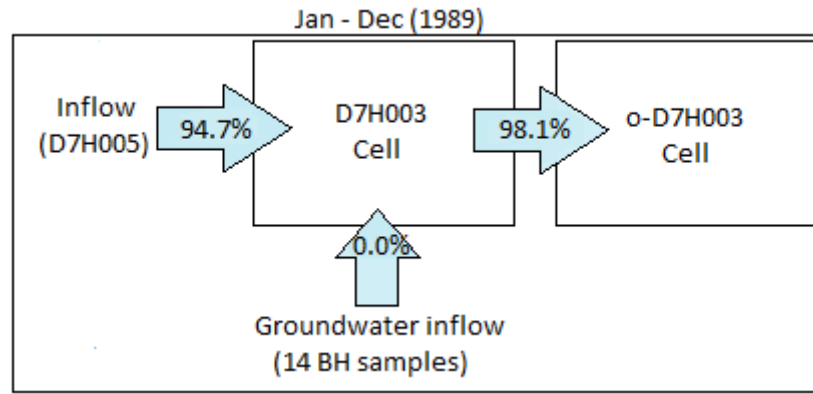


Figure 5-3-7 The unknown fluxes determined by the quaternary catchment D73F MCM application, making use of the scenario 2 weighting factors and assigning no abstraction from the river cell, expressed as a percentage of the total flow volume.

The amount of water abstracted from this section of the Orange River is significant at approximately $226 \text{ Mm}^3/\text{a}$, which lead to a MCM run which assigns an abstraction volume to the D7H003 Cell of $18.8 \text{ Mm}^3/\text{month}$, in order to investigate the significance of omitting this volume initially. The MCM run, using scenario 1 weighting factors and assigning an abstraction volume to the river cell, estimated the groundwater contribution to baseflow from the groundwater sources defined by the borehole samples B166990 and B166744, located in the middle of the river cell, at $0.00 \text{ Mm}^3/\text{month}$ (Table 5-3-2). The groundwater contribution to baseflow from the groundwater sources located on the northern side of the river and defined by the borehole samples B166928, B166929, B166930 and B167049 were estimated at $0.00 \text{ Mm}^3/\text{month}$. The groundwater contribution to baseflow from the groundwater sources located on the southern side of the river and defined by the borehole samples B166931, B166932, B166933, B166934, B166935, B167031 and B167048 were all estimated at $0.00 \text{ Mm}^3/\text{month}$, while the groundwater source defined by the borehole sample B167032 was estimated at $0.64 \text{ Mm}^3/\text{month}$ which extrapolated to a yearly flow volume becomes $7.68 \text{ Mm}^3/\text{a}$. The inflow to the D7H003 Cell, defined with water quality data from flowstation D7H005, was determined to be $446.00 \text{ Mm}^3/\text{month}$ which extrapolated to a yearly flow volume becomes $5352.00 \text{ Mm}^3/\text{a}$, while the outflow from the D7H003 Cell to the fictitious o-D7H003 Cell was determined to be $486.24 \text{ Mm}^3/\text{month}$ which extrapolated to a yearly flow volume becomes $5834.88 \text{ Mm}^3/\text{a}$. The MCM was run for the same setup, using scenario 2 weighting factors, resulting in an estimation of the groundwater contribution to baseflow, from all the defined groundwater sources, at $0.00 \text{ Mm}^3/\text{a}$ (Table 5-3-2). The inflow to the D7H003 Cell, defined with water quality data from D7H005 flowstation, was determined to be $495.41 \text{ Mm}^3/\text{month}$ or $5944.92 \text{ Mm}^3/\text{a}$, while the outflow from the D7H003 Cell to the fictitious o-D7H003 Cell was determined to be $504.59 \text{ Mm}^3/\text{month}$ or $6055.08 \text{ Mm}^3/\text{a}$.

Table 5-3-2 Summary of MCM results for the quaternary D73F area model run for each scenario

Name of inflow	Rate of inflow (Mm ³ /month)			
	No abstraction volume		Abstraction volume	
	Scenario 1	Scenario 2	Scenario 1	Scenario 2
InflowD7H5	430.35	479.73	446.00	495.41
B166744	0.00	0.00	0.00	0.00
B167032	0.62	0.00	0.64	0.00
B167031	0.00	0.00	0.00	0.00
B166928	0.00	0.00	0.00	0.00
B166929	0.00	0.00	0.00	0.00
B166930	0.00	0.00	0.00	0.00
B166931	0.00	0.00	0.00	0.00
B166932	0.00	0.00	0.00	0.00
B166933	0.00	0.00	0.00	0.00
B166934	0.00	0.00	0.00	0.00
B166935	0.00	0.00	0.00	0.00
B166990	0.00	0.00	0.00	0.00
B167048	0.00	0.00	0.00	0.00
B167049	0.00	0.00	0.00	0.00

5.3.7. MCM Discussion and comparison

The water and chemical mass balance errors for each model run and weighting factor scenario are shown in Table 5-3-3. Once again it can be seen that changing the flow volume (abstraction volume from the river cell) has no substantial effect on the water and average chemical mass balance errors. A detailed, tracer-specific table of all the chemical mass balance errors can be seen in Appendix E Table E-2. The assignment of an abstraction volume from the river cell results in an increase of inflow estimates, for both the upstream river inflow and the groundwater baseflow estimated for borehole source B167032, in the MCM run using scenario 1 (Table 5-3-2). A large decrease in both the water and average chemical mass balance errors are seen when changing from scenario 1 to scenario 2 (Table 5-3-3). The groundwater contribution to the river from the one groundwater source defined by the borehole sample B167032 goes from 7.44Mm³/a to 0.00Mm³/a, when changing from scenario 1 to scenario 2, in both model runs. From these results, it can be seen that the unknown inflows estimated by the MCM are more sensitive to the changes in tracer concentrations and weighting factors, than to changes in assigned flow volumes. Considering the fact that no considerable amount of groundwater baseflow could be contributing to the river in this particular area, the use of scenario 2 weighting factors gives the more accurate estimation of the groundwater baseflow.

Table 5-3-3 Water and chemical mass balance percentage errors associated with each model run and scenario for quaternary catchment D73F MCM run

Balance	Percentage Error (%)			
	No abstraction volume		Abstraction volume	
	Scenario 1	Scenario 2	Scenario 1	Scenario 2
Water balance	16.64%	7.21%	16.64%	7.54%
Chemical mass balance	11.14%	2.27%	11.14%	2.63%

The tracers assigned a lower weighting factor in scenario 2, are NO₃_NO₂ (Nitrate/Nitrite), NH₄ (Ammonia), TAL (Total Alkalinity) and Na (Sodium). Extensive agricultural activities take place alongside the Orange River because it is a plentiful source of water for irrigation. The majority of the registered water use within the study area is for irrigation purposes. Nitrogen based pollutants such as nitrate and ammonia are commonly associated with agricultural practices and the fact that these tracers were causing the MCM to initially estimate a positive groundwater inflow to the river might suggest that nitrogen based pollutants are reaching the Orange River by means of runoff from farm areas. No surface water pollution sources to the river cell were assigned in the MCM run due to lack of data, but this omission might have been the reason for the model assigning a groundwater baseflow value when in fact it should have been an additional pollution source from runoff. The particular borehole sample that this supposed inaccurate inflow was assigned to shows no substantial difference in NO₃_NO₂ and NH₄ tracer concentrations when compared to the other borehole samples, but the sodium concentrations are much higher than in all the other boreholes used to define groundwater inflow (Figure 5-3-8). Figure 5-3-8 is an S.A.R diagram indicating the uniqueness of the borehole sample B167032 in terms of Sodium content and EC (Electrical conductivity). The water quality of this borehole indicates that the Sodium Absorption Ratio (SAR) is classed between a medium and high threat, while the rest of the samples fall within classes S1 and S2, low and medium threats. The high sodium concentrations of this particular borehole and the assignment of groundwater baseflow to this source alone suggests that a runoff pollution source of sodium was not accounted for in the MCM application and thus could be the reason for the MCM predicting groundwater baseflow when the area is assumed to have none.

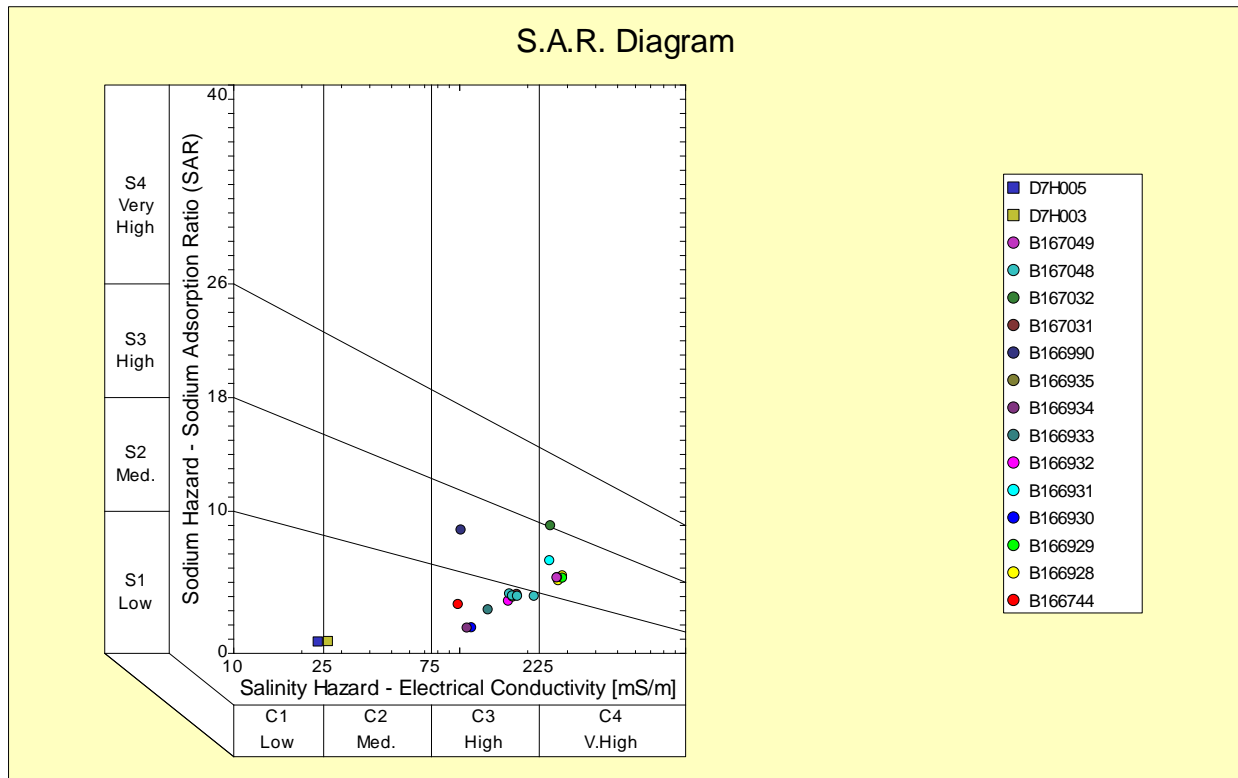


Figure 5-3-8 S.A.R diagram for the 14 borehole samples used for the defining of MCM groundwater sources within the quaternary catchment D73F study area

The MCM results from the model run assigning no abstraction volume to the river cell and making use of scenario 2 are used for comparison purposes. The total groundwater contribution to baseflow within the study area was estimated at 0.00Mm³/a by the MCM. Groundwater baseflow estimates from the Pitman, Hughes and Sami models for the quaternary catchment D73F as well as an estimate from another chemical method (Tracer method) using the same data as the MCM. The Pitman, Sami and Hughes model groundwater baseflow estimates for the quaternary catchment D73F were all 0.00Mm³/a (Table 5-3-4). The Tracer method estimate is shown for both the use of only electrical conductivity (EC) as well as an average value from using a number of tracers. The Tracer method groundwater baseflow estimate for the quaternary catchment D73F, using only EC was 93.49Mm³/a, while for an average from a number of tracers was 119.18Mm³/a.

Table 5-3-4 The groundwater baseflow estimates from the Pitman model, Sami model, Hughes model, tracer method and MCM for the quaternary catchment D73F.

Quat	Pitman Uncalibrated (Mm ³ /a)	Sami Uncalibrated (Mm ³ /a)	Hughes Uncalibrated (Mm ³ /a)	MCM (Mm ³ /a)	Tracer Method (Mm ³ /a)	
					EC	Average
D73F	0.00	0.00	0.00	0.00	93.49	119.18

The Tracer method estimate of groundwater baseflow is extremely unreasonable for this regionally-defined zero groundwater baseflow study area. The large flow volumes associated with the Orange River further highlight the Tracer method's short-coming, in that the method is principally dependent on the flow volumes assigned. The MCM is much less dependent on the flow volume assigned because only a slight difference in the groundwater baseflow estimate is seen when abstraction volumes from the river are included. The MCM is however more sensitive to changes in tracer concentration data and weighting factors assigned, as seen in the drastic change in the groundwater baseflow estimate between scenario 1 and 2 weighting factors. This sensitivity highlights the models shortcomings in terms of undefined sources to the system and non-compliance with the assumption of conservative tracers. However, when tracers showing high chemical mass balance percentage errors are assigned lower weighting factors, the MCM does estimate the groundwater baseflow accurately along with the Pitman, Hughes and Sami models.

Chapter 6 Collective Discussion

The application of the Mixing Cell Model to the quantification of groundwater – surface water is discussed in terms of: the groundwater baseflow estimates determined in each of the three study areas and their respective interpretations; the limiting factors and assumptions imposed on the MCM by the scope of this study; and lastly the data restrictions imposed on the application of the MCM in a South African setting.

The first pilot study area is located within a regionally-defined zero groundwater zone according to the Mean Annual contribution of Groundwater to Baseflow Map produced by the Department of Water Affairs (Department of Water Affairs, 2006b). The Pitman, Sami and Hughes models groundwater contribution to streamflow volumes were all zero for the quaternary catchment C52H, wherein the UFS groundwater – surface water interaction test site falls. The MCM estimated the average groundwater contribution to streamflow at $0.52\text{Mm}^3/\text{a}$ for the two time periods, while the Tracer method for EC-only estimates the average groundwater contribution at $0.59\text{Mm}^3/\text{a}$. The Mean Annual contribution of Groundwater to Baseflow map of South Africa indicates that the Pitman, Sami and Hughes models were correct in estimating the groundwater baseflow at zero, but the general nature of the map often results in inaccuracies on a quaternary catchment scale. Investigations on the UFS test site performed by Gomo (2011) found that groundwater was reaching the river within this area. Gomo (2011) suggested that the groundwater – surface water interaction occurring at this location is not taking place by traditionally-defined groundwater – surface water interfaces, based on the geology of the area and positive indications of groundwater contributions to streamflow. The concurrence between the MCM and Tracer method average groundwater baseflow volumes ($0.52\text{Mm}^3/\text{a}$; $0.59\text{Mm}^3/\text{a}$) is a good indication of the overall accuracy of these methods. The fact that both the MCM and the Tracer method indicated groundwater to be contributing to streamflow, while traditional methods did not, suggests that methods making use of water quality data could identify groundwater contributing to streamflow by unconventional pathways, where traditional methods could not. However, additional empirical studies and direct measurements are required to confirm this hypothesis.

The middle Modder River was divided into three sections based on the location of active flowstations in the investigated area. Section one is located within the quaternary catchment C52B. The Pitman model estimated the groundwater baseflow in quaternary catchment C52B at zero. The

Sami model has a low, but positive groundwater baseflow estimate of $0.03\text{Mm}^3/\text{a}$, while the Hughes model estimated the groundwater baseflow at $5.03\text{Mm}^3/\text{a}$. This quaternary catchment falls within a regionally-defined zero groundwater baseflow zone, yet the Sami and Hughes models have found that there is indeed groundwater contributing to the river. The uncalibrated estimates for these two methods (Sami and Hughes) are not in agreement, with a difference of $5\text{Mm}^3/\text{a}$. The MCM estimated the average groundwater baseflow at $0.09\text{Mm}^3/\text{a}$, while the Tracer method (EC) estimated an average groundwater baseflow of $1.31\text{Mm}^3/\text{a}$. The MCM estimate of groundwater baseflow is much lower than both the Tracer method and Hughes model estimates, but is in agreement with the estimate made by the Sami model of $0.03\text{Mm}^3/\text{a}$. This lower groundwater baseflow volume could be attributed to the fact that this section is the only section that includes a tributary source to the river mixing cell. The other sections also receive water from tributaries, but these tributaries weren't accessible for water samples to be taken. The inclusion of the tributary source improved the MCM estimate of groundwater baseflow when compared to the Sami estimate, but worsened when compared to the Tracer method. The quaternary catchment C52B still flows over Beaufort Group rocks of the Adelaide sub-group and the alternating layers of sandstone and shale could provide a reasonable supply of groundwater to the river. From these results and the underlying geology it seems that the MCM and Tracer method produce reasonable groundwater baseflow estimates, with the MCM being slightly more suitable due to the inconsistency of the Tracer method between the two time periods. The Tracer method is also variable between the use of only EC and multiple tracers, which makes the MCM better suited for the incorporation of multiple tracer concentration data.

Section two is located within the quaternary catchment C52E. The Pitman and Sami models determined that no groundwater was contributing to streamflow in C52E, while the Hughes model estimated the groundwater baseflow at $2.22\text{Mm}^3/\text{a}$. The MCM determined the average groundwater baseflow estimate at $1.30\text{Mm}^3/\text{a}$ and the Tracer method determined the average groundwater baseflow estimate at $0.78\text{Mm}^3/\text{a}$. Once again the Sami and Pitman models are correct based on the Mean Annual contribution of Groundwater to Baseflow map, but the Hughes, MCM and Tracer methods indicate that groundwater is indeed contributing to the stream in this area. The Modder River in Section two mainly flows over Beaufort Group rocks of the Adelaide sub-group that assumedly would be able to supply groundwater to the stream, if the two resources were connected. Determining whether the Sami and Pitman models indicating no groundwater baseflow or the Hughes, MCM and Tracer methods indicating groundwater baseflow were correct in this instance would require additional study of the area.

Section three is located within the quaternary catchment C52G. The Pitman and Sami models determined that no groundwater was contributing to streamflow in C52G, while the Hughes model estimated the groundwater baseflow at $5.35\text{Mm}^3/\text{a}$. The average groundwater baseflow estimate for the MCM is $2.37\text{Mm}^3/\text{a}$, and for the Tracer method $2.42\text{Mm}^3/\text{a}$.

From the results of the middle Modder River study area, it can be seen that the Sami and Hughes models are not always in agreement. The Hughes model estimate of the groundwater baseflow is on average $5\text{Mm}^3/\text{a}$ higher than the Sami model groundwater baseflow estimate. The MCM and Tracer method groundwater baseflow estimates tend to be in-between the two model estimates. For example, in section three the Sami, MCM, Tracer and Hughes models' groundwater baseflow estimates are $0.00\text{Mm}^3/\text{a}$, $2.37\text{Mm}^3/\text{a}$, $2.42\text{Mm}^3/\text{a}$ and $5.35\text{Mm}^3/\text{a}$, respectively. Again, further studies would need to be conducted on these areas in order to validate these findings.

The continuous MCM run for the entire middle Modder River section presented as a demonstration for the models applicability to ungauged catchments provided reasonable percentages of inflow from the defined groundwater sources. Further study would be required to confirm that the MCM would be the most appropriate method to apply as an initial estimate of groundwater baseflow to ungauged river systems.

The second pilot study area in the Limpopo falls within a groundwater zone of $6\,000 - 10\,000\text{m}^3/\text{km}^2/\text{a}$ zone according to the Mean Annual contribution of Groundwater to Baseflow map (Department of Water Affairs, 2006b). Using the area of each quaternary catchment, the groundwater baseflow zone equates to $3.5 - 6.0\text{Mm}^3/\text{a}$ for A42A, $3.1 - 5.2\text{Mm}^3/\text{a}$ for A42B and $4.2 - 7.0\text{Mm}^3/\text{a}$ for A42C. The uncalibrated Pitman, Sami and Hughes models groundwater contribution to streamflow estimates for the quaternary catchment A42A were $17.99\text{Mm}^3/\text{a}$, $8.70\text{Mm}^3/\text{a}$ and $15.62\text{Mm}^3/\text{a}$, respectively. The uncalibrated Pitman, Sami and Hughes models groundwater contribution to streamflow estimates for the quaternary catchment A42B were $16.86\text{Mm}^3/\text{a}$, $6.56\text{Mm}^3/\text{a}$ and $15.39\text{Mm}^3/\text{a}$, respectively. The uncalibrated Pitman, Sami and Hughes models groundwater contribution to streamflow estimates for the quaternary catchment A42C were $22.41\text{Mm}^3/\text{a}$, $8.83\text{Mm}^3/\text{a}$ and $20.26\text{Mm}^3/\text{a}$, respectively. The groundwater baseflow estimates from the three models for each quaternary catchment are highly variable, with a maximum difference of $11\text{Mm}^3/\text{a}$ (between the Sami and Hughes models). The calibration of the Sami model results in an increase in the groundwater baseflow estimate for quaternary catchment A42A ($8.70\text{Mm}^3/\text{a}$ to $8.99\text{Mm}^3/\text{a}$), for the quaternary catchment A42B ($6.56\text{Mm}^3/\text{a}$ to $8.50\text{Mm}^3/\text{a}$), and for the quaternary catchment A42C ($8.83\text{Mm}^3/\text{a}$ to $10.28_{(E)}\text{Mm}^3/\text{a}$). The calibration of the Hughes model

results in a large decrease in the groundwater baseflow estimate for the quaternary catchment A42A (15.62Mm³/a to 11.90Mm³/a), for the quaternary catchment A42B (15.39Mm³/a to 11.72Mm³/a), and for the quaternary catchment A42C (20.26Mm³/a to 15.43_(E)Mm³/a). From these trends in calibration, the Sami model is an under-estimation of the groundwater baseflow, while the Hughes model is an over-estimation of the groundwater baseflow. The area is extremely data scarce in terms of water quality data and this is reflected in the MCM results as the groundwater baseflow estimate for the considered time period is only 1.21Mm³/a, with a water balance error of 33% and an average chemical mass balance error of 10%. The MCM groundwater baseflow volume seems to be a substantial under-estimation when compared to the other methods. The Tracer method however does give a reasonable groundwater estimate of 9.83Mm³/a using EC (similar to the Sami model estimate). The Tracer method is designed to estimate the groundwater baseflow from an entire upstream catchment making use of only a single downstream sample point. In the investigation, the Tracer method produced accurate estimates of the groundwater baseflow in a data-scare area, where only a single river sample point was available.

The third pilot study area (quaternary catchment D73F) falls within a regionally-defined zero groundwater baseflow zone according to the Mean Annual contribution of Groundwater to Baseflow map (Department of Water Affairs, 2006b). The Pitman, Sami and Hughes models all estimated the groundwater contribution to streamflow as zero, for the quaternary catchment D73F. The MCM estimated the groundwater baseflow at 7.44Mm³/a when all tracers were assumed conservative, but when the tracers showing high chemical mass balance errors were given a lower weighting factor, the estimated groundwater baseflow becomes 0.00Mm³/a. The initially positive groundwater baseflow estimated by the MCM could be attributed to the omission of a pollution runoff source to the river which resulted in the runoff being erroneously assigned to a high sodium borehole sample. The MCM proved more sensitive to the changes in the chemical mass balance than changes in the water balance. The assignment of an abstraction volume from the River Cell had no significant effect on either the groundwater baseflow estimate or the associated balance errors. However, changing the weighting factor scenario had a drastic effect on both the estimated groundwater baseflow and the associated balance errors. The Tracer method proved extremely sensitive to the flow volume assigned. The Tracer method estimated a large groundwater baseflow volume (93.49Mm³/a) for the study area in response to the large outflow of the Orange River, in spite of the area not being able to supply groundwater baseflow to the river in such large quantities.

The MCM was not operating at its full capacity when applied to the pilot studies due to a number of factors associated with the limited scope of this study. The main limitation was the lack of isotope data. The MCM has the ability to incorporate isotope data into the chemical mass balance equations and the use of such data could have increased the accuracy of the groundwater baseflow estimates. The MCM also has the ability to determine the inflow to the river from both groundwater and interflow, but this capacity was not exploited within this study because water quality data required to define an interflow source is not readily available and difficult to measure. The omission of an interflow source to MCM river cell could also be a source of error in the estimated groundwater baseflow volumes seeing that the method is based on a water balance approach. The method has been investigated on the limiting assumption that the rivers on which the MCM was applied were gaining streams, thus determining positive groundwater flow from the aquifer to the river. The MCM application to the Northern Cape quaternary catchment located in a zero groundwater baseflow zone indicated that the model is able to accurately determine the groundwater baseflow even when there is none. The conceptual understanding of the area however leads to the hypothesis that the river is disconnected from the water table because the water table is far below ground level. The MCM can determine that there is no groundwater baseflow, but would not be able to simulate the flow from the river to the aquifer with the current mixing cell setup. A shift in the main focus of the mixing cells, from the river to the aquifer, is recommended for a losing river system investigation. This shift in focus would however require supplementary groundwater quality data close to the rivers to define alluvial channel aquifer mixing cells. The application of the MCM to connected/disconnected rivers requires further investigation.

The greatest limiting factor, as with most methods of groundwater – surface water interaction quantification in South Africa is data paucity. When the data available has been below perceived optimum levels for a realistic MCM application, the data has been referred to as insufficient. This rough description is used as there is no “required minimum number of data points per area” standard for a MCM application. In light of this short-coming a minimum standard description is attempted based on the findings of this study. It was found that the MCM did not produce optimum results when water quality data from a single river sample were used to define the river system. The MCM was able to produce reasonable estimates of the groundwater baseflow when small tributary inflows to a river segment are not accounted for, but was found to create large balance errors when larger tributary inputs were omitted. Water quality data from borehole samples relatively close to the river are required to define groundwater inflow sources, but terrestrial groundwater sources would be also beneficially. As many borehole samples are possible should be

made use of in a MCM application, which can be incorporated as single groundwater sources or can be averaged to represent the general groundwater quality within an area. Considering the MCM setups and results from this study, a rudimentary indication of the minimum data requirements for a small, single-cell MCM application as well as a multi-cell catchment scale application is given. For a single cell MCM application, water quality data from at least two river samples are required to define the river cell and the inflow into that mixing cell; at least one groundwater sample is required to define a groundwater source to the river; if possible tributary inflows should be defined; and lastly a outflow volume from the defined system is required. For a catchment scale MCM application, a river sample should define a river mixing cell at least every 5km, with a river sample every 1km being the optimum situation; at least one borehole sample in relative proximity should define the groundwater source to each river mixing cell; tributary inflows should be defined where possible; and lastly a outflow volume from the defined system is required.

The MCM has been run for a maximum of two time periods for each study area, based on the available data. The water quality sample points and data records within most quaternary catchments are irregular and insufficient, which does not allow for a MCM to be run on a regular monthly or yearly basis. A continuous application of the MCM over an extended time period, which could then be averaged, would give a better estimate of the groundwater contribution to streamflow.

Chapter 7 Conclusions

The main findings of the study are:

- ➔ The traditionally used Pitman, Hughes and Sami models were found to give dissimilar groundwater baseflow volumes for most of the study areas. The Sami and Hughes models give better groundwater baseflow estimates than the well-known, hydrology-based Pitman model. The Sami model was found to under-estimate groundwater baseflow volumes, while the Hughes model over-estimated groundwater baseflow volumes, based on changes resulting from calibration. Based on the results of this study, the Sami model appears to produce more reasonable and stable groundwater baseflow estimates than the Hughes model.
- ➔ The Mixing Cell Model (MCM) application to groundwater – surface water interaction quantification has limitations in terms of its use in South Africa. The MCM requires both surface water and groundwater quality data for a successful model run. In data scarce areas of South Africa the water quality data required for a MCM might not always be available which will limit the use of this method. The use of insufficient water quality data in a MCM application has proven to result in severe under-estimation of groundwater baseflow. The chemical hydrograph separation method (Tracer method) however, proved to give reasonable groundwater baseflow volumes under these circumstances, but is highly dependent on the assigned outflow volume.
- ➔ The Tracer method and the MCM both require an outflow volume from the system to be defined in order to volumetrically determine the groundwater baseflow. The requirement of a volumetrically-defined outflow could prove problematic in ungauged catchments. The MCM can determine inflows to a system relatively as percentages of an assumed 100% outflow, in the absence of a known outflow volume, which could serve as a means of initially determining the groundwater baseflow in ungauged catchments.
- ➔ Both the water quality data based methods indicated groundwater baseflow to a system where the Pitman, Sami and Hughes models had indicated zero groundwater baseflow. The area was previously found to have untraditional groundwater – surface water connections, implying that the water quality data methods might detect groundwater baseflow that traditionally defined methods could not.

- The MCM is better suited for the incorporation of multiple tracer datasets than the Tracer method because the latter showed considerable inconsistencies. The incorporation of multiple tracer concentration data allows for the maximum amount of data to be used, ensuring the best possible groundwater baseflow estimate.
- The MCM was found to be more sensitive to the tracer concentrations assigned than the flows defined in the model, indicating that the assumption of conservative tracers is the limiting assumption. The assignment of lower weighting factors to non-conservative tracers does however allow the MCM to overcome this limitation to some degree.
- The MCM application to groundwater – surface water interaction quantification has advantages in physical-parameter data scarce and complex geological areas of South Africa because the model only requires water quality and flow data.
- The MCM is not limited to a specific scale of investigation as with the Sami model. The MCM can be applied to quaternary catchment scale, multiple catchments and local, site-specific investigation scales. The MCM does however perform better on smaller scales because the amount of data required for large scale applications is extensive and tends to result in greater errors.
- The algorithm-based MCM accurately determined the groundwater baseflow for a quaternary catchment in a zero groundwater baseflow zone, proving that the model can produce accurate results in spite of not directly using physical geohydrological principles.
- The MCM is able to report the associated water and chemical mass balance error of each groundwater baseflow volume determined which is advantageous in terms of expressing the level of confidence in the estimates.
- When there is sufficient data for a MCM run, the MCM groundwater baseflow volume tends to be in-between the Sami and Hughes model estimates. Considering that the Sami model was found to under estimate the groundwater baseflow and the Hughes model to over-estimate, the MCM is a good indication of the amount of groundwater contributing to a river.

In light of the findings of this study which found both advantages and disadvantages in using the Mixing Cell Model (MCM), the possible applications and scenarios where the use of the MCM would be beneficial are described. The MCM should be used to quantify the groundwater contribution to streamflow where:

- ✓ sufficient surface water and groundwater quality data is available or can be acquired,
- ✓ traditional methods of quantification are only able to produce low confidence estimates of groundwater baseflow (the MCM can be used to validate these estimates),
- ✓ traditional methods are not able to be utilised due to lack of physical-parameter data or complex geology, and/or
- ✓ a multiple-method approach is applied (the MCM can serve as an additional method incorporating chemical data).

The natural environment can hardly be described by linear, homogeneous expressions or with the assumption of conservative behaviour of solutes, especially in the geochemically very active hyporheic zone separating surface and groundwater. While the model allows for a partial compensation of such violations, the infringement of assuming conservative tracers is still likely to result in errors. We thus have to accept a certain level of inaccuracy in the quantification of natural phenomena such as the groundwater contribution to baseflow using the mixing cell or other methods. Similarly, data paucity and ungauged catchments limit the applicability of baseflow estimation models based on empirical data in large parts of the country. The number and maintenance of flowstations as valuable empirical data collection points should therefore be increased. The development of groundwater – surface water interaction test sites throughout the country is furthermore a major step towards improving the quality and quantity of data, with likely subsequent improvements in the different quantification tools once they can be tested against detailed data. Groundwater contribution to streamflow has proven to be a challenging property to determine, however the benefits of quantifying this flux greatly outweigh the difficulties.

As a final conclusion, this report and the proposed method should be a step in a continual process aimed at improving the quantification of groundwater baseflow. The significant role of this estimation has the potential to greatly influence our environment and how the country's water resources are managed and utilized.

Chapter 8 Recommendations

The Mixing Cell Model (MCM) is recommended as an additional tool used to quantify the groundwater baseflow volume in South Africa, and the following scenarios are suggested:

Validation:

The MCM uses a completely different dataset from the currently applied Sami and Hughes models for groundwater baseflow quantification. The MCM thus could serve as a means of confidently validating the groundwater baseflow volumes estimated by the traditional models.

Low confidence groundwater baseflow areas:

The GRA2 dataset which made use of the Sami model to quantify groundwater baseflow is known to have areas where the groundwater baseflow volumes have a low confidence. The MCM could be applied to these areas to increase the confidence of the estimated groundwater baseflow volume.

Physical-parameter data scarce and complex geology areas:

The MCM is able to estimate the groundwater baseflow without geohydrological parameters such as hydraulic conductivity (K) and the various parameters required for the Sami and Hughes models. Complex geology tends to limit the ability to determine such aquifer parameters which might limit the applicability of the traditional methods. The MCM could be applied to such areas to increase the confidence in the groundwater baseflow volume.

Ungauged quaternary catchments:

The MCM is able to determine the relative contribution of groundwater to a river system when there is no volumetrically known outflow from a flow system. The MCM could be used as an initial estimate of the groundwater baseflow in ungauged catchments.

Site-specific scale:

The MCM is not limited to a specific scale, allowing for both multiple catchment estimations and local scale applications. In highly impacted or high water use areas, the MCM can be applied on a small scale to more accurately determine the groundwater baseflow volume at these specific locations.

Multiple method approach:

The MCM could serve as a valuable additional method to a multiple-method approach of quantifying groundwater baseflow because the method incorporates chemical data (a different dataset to traditionally methods). The multiple method approach for the quantification of groundwater—surface water interaction has been recommended by numerous authors.

The following additional recommendations are made based on the findings of the study:

✓ Tracer method substitute:

When there is insufficient data for a MCM application, but a single river sample point at the outflow point of a catchment with water quality and flow data is available, the use of the Tracer method is recommended in place of the MCM. The Tracer method would allow for the incorporation of some chemical data when the MCM cannot be used.

✓ Improvement of water quality data and flowstation networks:

The above applications of the MCM are made based on the assumption that there is sufficient groundwater and surface water quality data for an accurate MCM run. It is therefore recommended that the water quality data monitoring system within South Africa be both maintained and expanded. The optimum situation for each quaternary catchment would be a water quality monitoring site at the inflow point, outflow and on each significant tributary to the main river system. Furthermore, in order to volumetrically determine the groundwater baseflow using the MCM, an outflow volume from the river system is required. It is thus also recommended that the flowstation network be maintained and expanded. The optimum situation would be an active flowstation at the base of each quaternary catchment.

✓ Multiple method approach:

The quantification of groundwater baseflow is a difficult component of the hydrological system to define because it is unseen. Recharge is a similarly difficult component to quantify and this has resulted in it becoming common practice to use an average recharge volume from a number of methods. It is recommended that a similar approach be applied to the quantification of the groundwater contribution to streamflow, in that an average volume from methods such as the Sami model, Hughes model and Mixing

Cell Model (MCM) are used instead of a groundwater baseflow volume from one individual method.

✓ Further investigation of the Mixing Cell Model (MCM):

The MCM requires further investigation in terms of a comprehensive sensitivity and statistical analysis, which was out of the scope of this study. A rudimentary indication of the minimum data requirements for a MCM application was given within the context of this study, but also requires further investigation and definition.

✓ Comprehensive groundwater – surface water interaction test sites:

In order to ultimately conclude which methods are able to accurately determine the groundwater baseflow, a perfectly-defined or theoretical test site should be developed to compare the various methods. The comparison would be conclusive as the actual groundwater baseflow would be known and all necessary data for each method would be available. The test site would be able to determine which method is best at optimum conditions (all necessary data is available) as well as which methods work best in data scarce environments as data could be systematically be omitted.

There is a large amount of assessment that is still required on the MCM, but it has the potential to greatly improve the existing estimates of the groundwater contribution to streamflow if the method is made use of correctly, implemented in the correct situations and ultimately used as an additional method to a multiple-method approach for the quantification of groundwater—surface water interaction.

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Appendix A:

A step-by-step guide to the use of MCMsf

The MCMsf (Mixing Cell Model – steady flow) is accompanied with three example datasets which are used for tutorial purposes. The examples include a single-cell model, a four-cell model and a twelve-cell model. The four-cell model dataset is made use of and applied as an example flow system to present a step-by-step guide to a MCMsf model run. The conceptual box model representation of the simulated flows for the four-cell example model is shown in Figure A-1. From the flow diagram (Figure A-1) it can be seen that water from Cell 1 flows into Cell 2 and Cell 3, Cell 2 flows to Cell 3 and Cell 4, and Cell 3 flows to Cell 4. The outflow from the last cell, Cell 4 is defined within the model at $100\text{m}^3/\text{year}$. Fourteen inflow sources are defined for this flow system. Sources (Flow 1-4) are assigned to Cell 1, sources (Flow 5-9) are assigned to Cell 2, sources (Flow 10-12) are assigned to Cell 3 and sources (Flow 13-14) are assigned to Cell 4. The abstraction rates are assigned at $0.5\text{m}^3/\text{year}$, $3.5\text{m}^3/\text{year}$, $15.5\text{m}^3/\text{year}$ and $5.0\text{m}^3/\text{year}$ for the Cells 1 – 4, respectively.

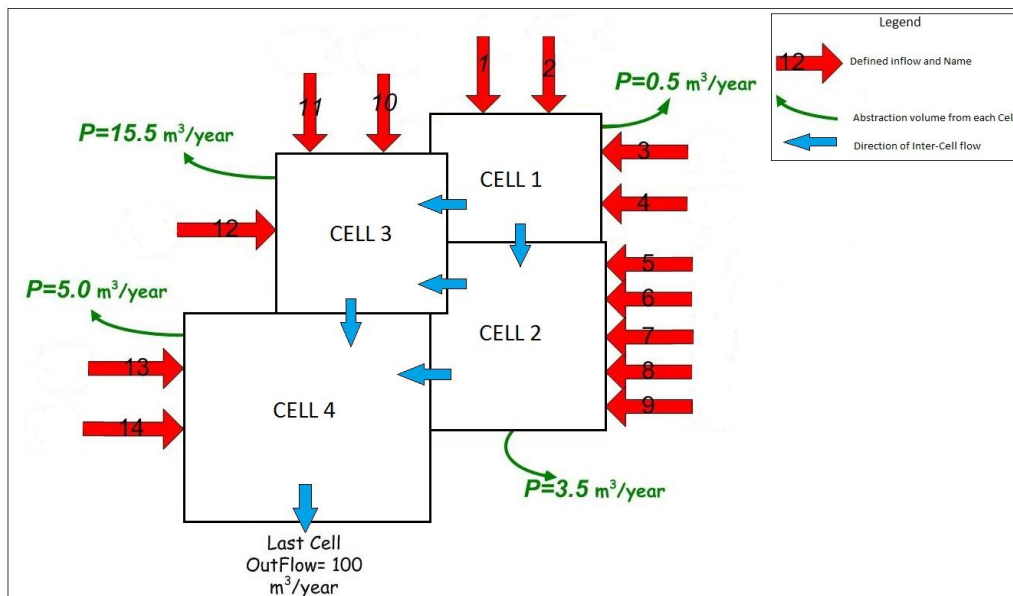


Figure A-1 Conceptualise box model flow diagram for the MCMsf four-cell example (Modified from Adar (2012)).

The MCMsf follows a specifically structured procedure to define the mixing cell model and create an input matrix for the FORTRAN© solver code based on the Wolf Algorithm. The structure of the process followed within the MCMsf programme is shown in Figure A-2. This guide to using the MCMsf programme will thus also follow these steps.

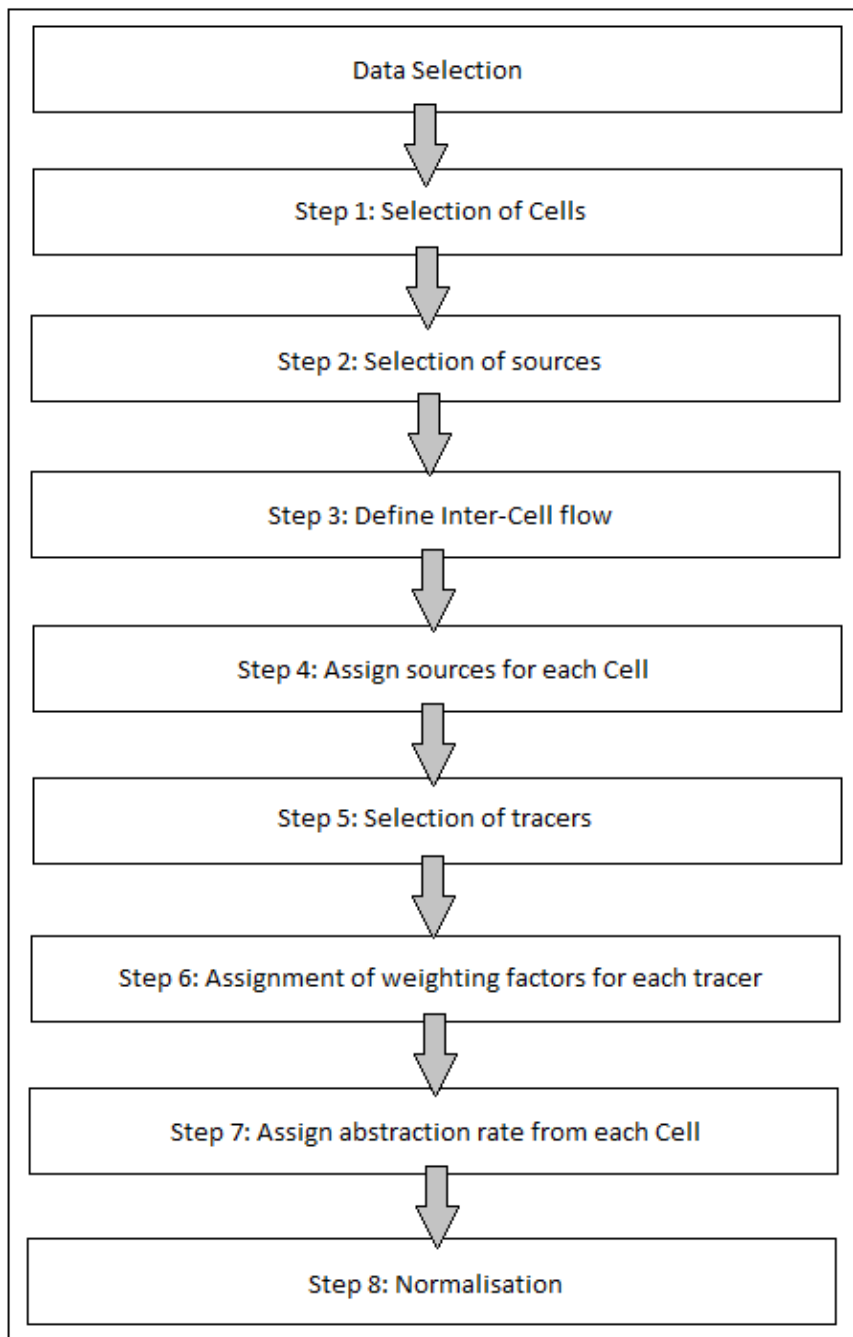


Figure A-2 The structured procedure followed in the MCMsf programme to define a mixing cell model (Modified from Adar (2012)).

Data Selection

The dataset from the MCMsf included four-cell example is made use of and shown in Table A-1. The water quality data forms the input into the programme is stored in an Excel format and requires a specific order of column labels to be read by the MCMsf programme (Table A-1). The data is first imported into the Microsoft© Access database and then is shown within the MCMsf programme under *List of Table Data* shown on the user-friendly interface as the MCMsf programme is started (Figure A-3). In the case of the example four-cell model, the data is

automatically available for selection within the MCMsf programme. The “tbl_data_4_14” is selected from the *List of data Tables*.

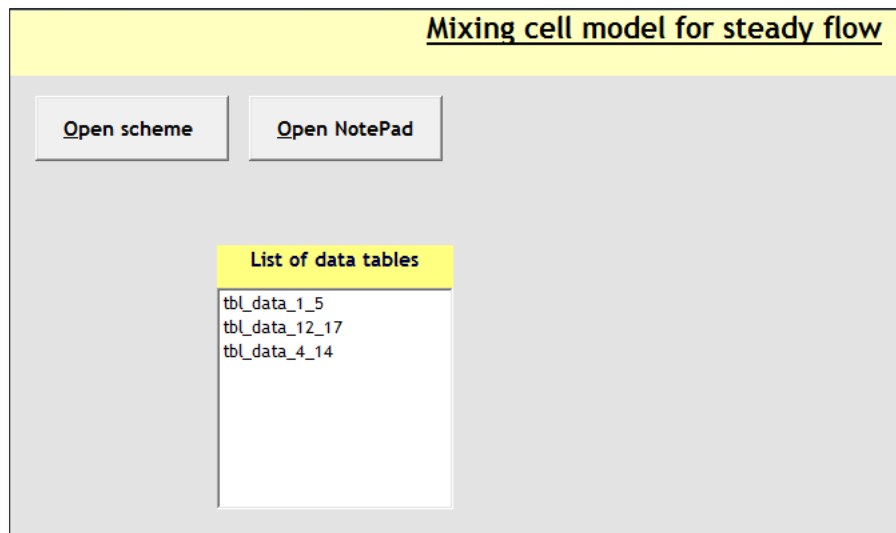


Figure A-3 MCMsf programme interface for the Data Selection step

Selection of Cells

The selection of Cells MCMsf interface is shown in Figure A-4. This interface allows the user to select the water quality analysis datasets which are used to define the mixing cell model Cells. For the four-cell example model, the datasets for Cell 1 – Cell 4 are selected from the *List of Cells* displayed on the interface (Figure A-4).

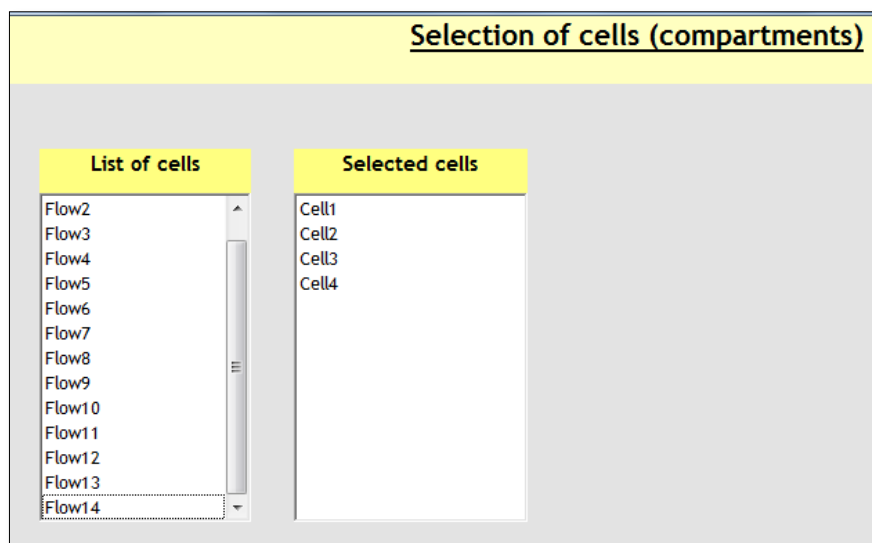


Figure A-4 MCMsf interface for Step 1 (Selection of Cells).

Table A-1 Water quality dataset for the MCMsf four-cell example model.

Location	Sample	Longitude	Latitude	GWL	TDS	Mg	Ca	Na	K	HCO3	Cl	NO3	SO4	F	Li	Si	D	O-18
Flow1					353.80	0.74	1.97	0.97	0.06	3.08	0.21	0.08	0.42	0.023	0.0026	18.01	-73.80	-9.73
Flow2					187.10	0.36	0.81	0.39	0.04	0.89	0.09	0.00	0.80	0.001	0.0003	18.18	-66.30	-8.56
Flow3					146.00	0.53	1.13	0.51	0.05	1.40	0.08	0.04	0.66	0.006	0.0012	17.53	-60.60	-8.55
Flow4					332.30	0.62	2.31	0.69	0.17	2.97	0.15	0.03	0.38	0.025	0.0013	16.83	-58.70	-7.41
Flow5					304.50	0.10	0.24	2.01	0.05	2.52	0.26	0.02	0.05	0.036	0.0022	14.00	-67.20	-8.70
Flow6					370.30	0.77	1.67	1.70	0.07	3.04	0.69	0.22	0.14	0.008	0.0032	8.80	-63.10	-6.52
Flow7					454.00	1.28	2.52	0.98	0.03	3.71	0.42	0.06	0.64	0.012	0.0016	16.35	-76.30	-10.18
Flow8					242.00	0.39	1.51	0.33	0.06	1.20	0.09	0.04	0.63	0.030	0.0020	17.70	-67.10	-8.98
Flow9					332.30	0.62	2.31	0.69	0.17	2.97	0.15	0.03	0.38	0.025	0.0013	16.83	-58.70	-7.41
Flow10					356.00	1.72	3.18	0.73	0.03	3.64	0.14	0.05	1.93	0.015	0.0016	14.80	-67.00	-8.43
Flow11					147.80	0.84	0.41	0.33	0.03	0.07	0.07	0.01	0.98	0.120	0.0020	7.80	-68.90	-9.36
Flow12					362.00	0.66	2.06	0.87	0.05	2.93	0.14	0.08	0.51	0.022	0.0023	17.10	-69.30	-9.66
Flow13					242.00	0.39	1.51	0.33	0.06	1.20	0.09	0.04	0.63	0.030	0.0020	17.70	-67.10	-8.98
Flow14					332.30	0.62	2.31	0.69	0.17	2.97	0.15	0.03	0.38	0.025	0.0013	16.83	-58.70	-7.41
Cell1					321.91	0.68	1.79	0.87	0.06	2.71	0.18	0.07	0.48	0.019	0.0022	17.98	-71.82	-9.46
Cell2					323.27	0.65	1.71	0.92	0.06	2.62	0.21	0.06	0.46	0.021	0.0021	17.23	-70.55	-9.24
Cell3					303.85	0.98	1.94	0.74	0.05	2.50	0.15	0.05	0.96	0.037	0.0020	15.13	-69.63	-9.16
Cell4					317.59	0.77	1.91	0.80	0.08	2.64	0.18	0.05	0.64	0.028	0.0019	16.35	-67.94	-8.86

Selection of Sources

The selection of sources MCMsf interface is shown in Figure A-5. This interface allows the user to select the water quality analysis datasets which are used to define the inflows to the MCM Cells. For the four-cell example model, the datasets for Flow 1-14 are selected from the *List of Sources* displayed on the interface (Figure A-5).

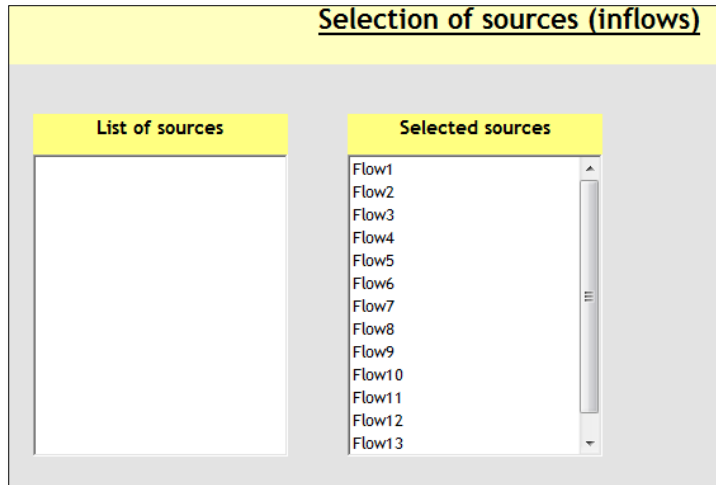


Figure A-5 MCMsf interface for Step 2 (Selection of Sources).

Definition of Inter-Cell flow

The inter-cell flow is assigned by first selecting the Discharging Cell from the *Discharging cell* list on the MCMsf interface and then selecting a Receiving Cell from the *Receiving cell* list (Figure A-6). For the four-cell example model Cell 3 is selected as the discharging cell and Cell 2 and Cell 3 are then selected as receiving cells. Similarly, flow from Cell 2 to Cell 3 and Cell 4 as well as flow from Cell 3 to Cell 4 are assigned.

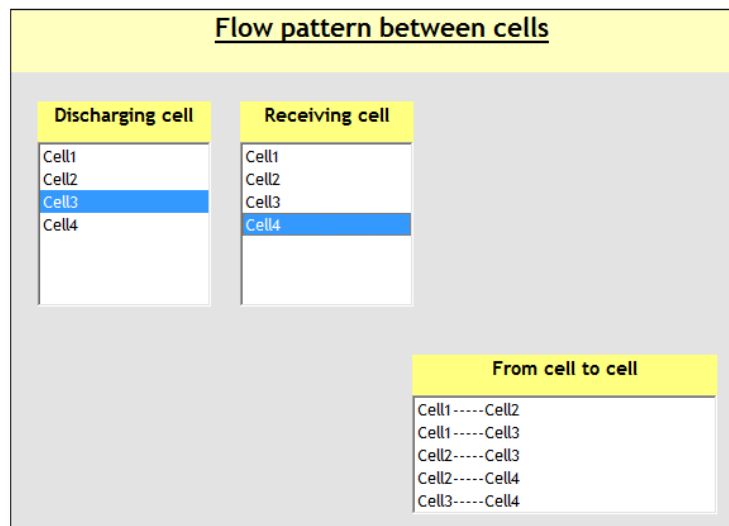


Figure A-6 MCMsf interface for Step 3 (Defining Inter-Cell flows).

Assignment of Sources to Cells

The MCMsf programme then requires the user to specify which of the chemically defined sources are have been identified as contributing to which Cells. For the MCMsf four-cell example, the defined sources (Flow 1-4) are assigned to Cell 1 by first selecting Cell 1 in the *Cells* list and then selecting each of the sources (Flow 1-4) in the *Sources* list. Similarly, sources (Flow 5-9) are assigned to Cell 2, sources (Flow 10-12) are assigned to Cell 3 and sources (Flow 13-14) are assigned to Cell 4 (Figure A-7).

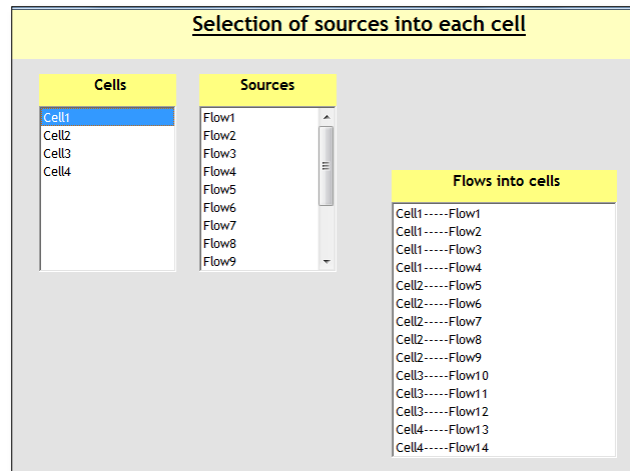


Figure A-7 MCMsf interface for Step 4 (Assign sources for each Cell).

Selection of Tracers

A list of tracers from the dataset loaded into the MCMsf is now shown in order to allow the user to selection which tracers are to be used in the chemical mass balance equation. For the four-cell example, all the available tracers are selected for use in the MCMsf programme, namely TDS, Mg, Ca, Na, K, HCO₃, Cl, NO₃, SO₄, F, Li, Si, D (Deuterium) and O-18 (Figure A-8).

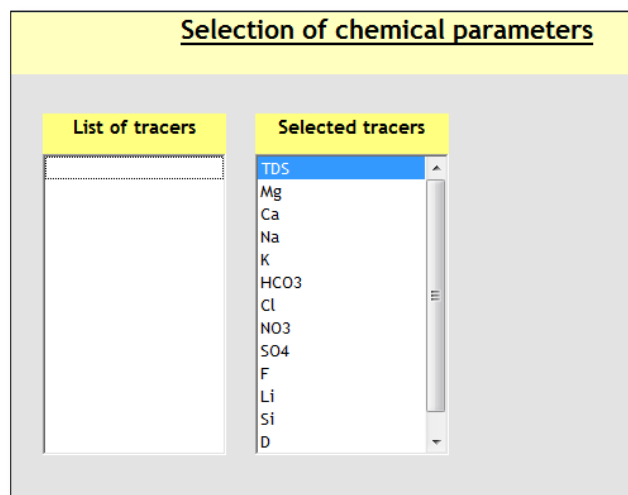


Figure A-8 MCMsf interface for Step 5 (Selection of Tracers).

Assignment of weighting factors for each tracer

The MCMsf programme allows weighting factors to be assigned to individual tracers in order to compensate for their lack of total conservation of that tracer. For the MCMsf four cell example all tracers are given a weighting factor of 1 which is the default setting (Figure A-9). In order to change the weighting factor the user simply selects the tracers to be changed and assigns a new weighting factor in a pop-up screen.

List of tracers	Weight (0.0 - 1.0)
TDS	1
Mg	1
Ca	1
Na	1
K	1
HCO3	1
Cl	1
NO3	1
SO4	1
F	1
Li	1
Si	1
D	1
O-18	1

Figure A-9 MCMsf interface for Step 6 (Assignment of weighting factors for each tracer).

Assignment of abstraction rates for each Cell

The MCMsf programme allows the abstraction volume from each cell to be defined. The abstraction volume in the four-cell example for Cell 1 is $0.5\text{m}^3/\text{year}$. To assign this volume in the MCMsf programme Cell 1 is first selected from the *Cell name* list and then the abstraction volume can be inserted into a pop-up screen. Once inserted, the abstraction volume assigned will be shown in the *Outflow (volume/time)* list (Figure A-10). Similarly, the abstraction volumes of Cell 2, Cell 3 and Cell 4 of $3.5\text{m}^3/\text{year}$, $15.5\text{m}^3/\text{year}$ and $5.0\text{m}^3/\text{year}$, respectively are assigned. The total outflow from the last cell is also required to be defined by the MCMsf programme which can be done in the *Outflow from the last cell* insertion block in the bottom right-hand corner of the interface screen. In the case of the four-cell example the outflow from the last cell is set to $100\text{m}^3/\text{year}$.

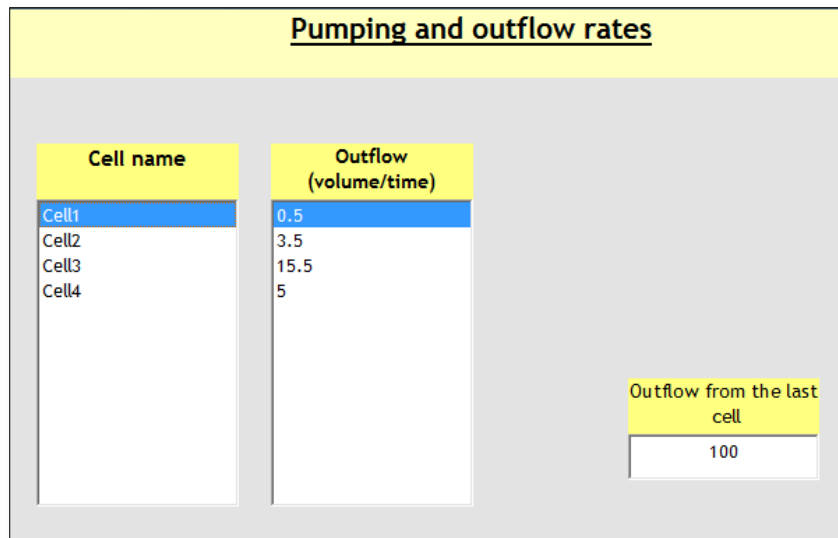


Figure A-10 MCMsf interface for Step 7 (Assign abstraction rate from each cell).

Normalisation

The MCMsf programme allows the user to select a set of tracer concentrations which are suitable to use a reference for normalisation. The MCMsf programme automatically indicates which datasets are suitable and shown under the *List of Cells* list. For the MCMsf four-cell example Cell 4 is used as a reference for normalisation and simply assigned by selecting Cell 4 from the list (Figure A-11). The MCMsf programme also allows for the option of selecting no dataset for normalisation by clicking on the button in the bottom right-hand corner of the screen indicating *Without Normalisation* (Figure A-11).

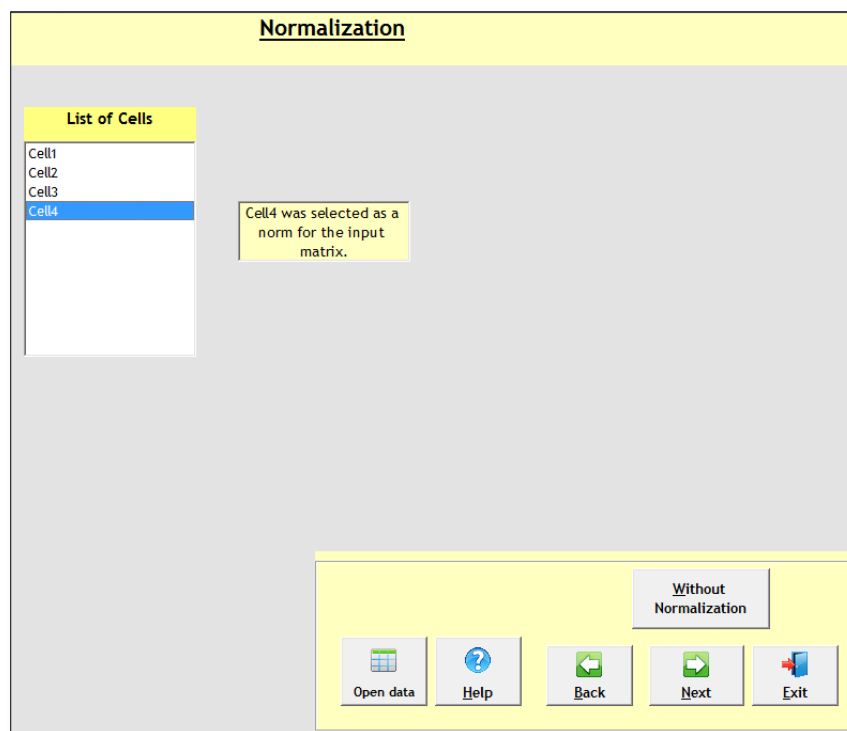


Figure A-11 MCMsf interface for Step 8 (Normalisation).

Input Matrix and FORTRAN© solver execution

The completion of Step 8 (Normalisation) is the final step in defining the Mixing Cell Model. One completion of this step, an additional interface will appear which then initiates the creation of the input matrix for the Wolf algorithm, save the input file and then allow the input file to be run in the Wolf Algorithm FORTRAN solver (Figure A-12). The input matrix has to be created first by selecting the *Build Input Matrix* button, secondly the input matrix has to be saved by selecting the *Save Input File* button and lastly the input matrix is used for a Wolf algorithm run by selecting the *Run Wolf Algorithm* button.

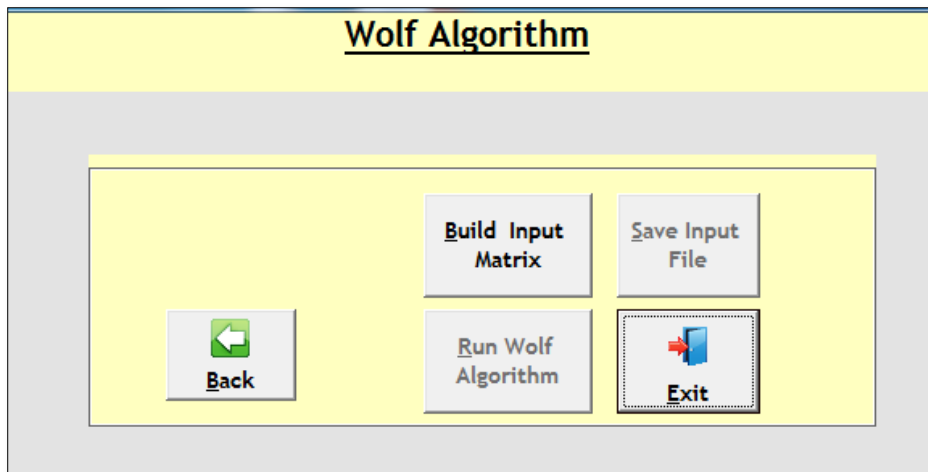


Figure A-12 MCMsf interface for the creation and application of the input matrix.

The FORTRAN solver has a basic interface shown in Figure A-13. The solver requires initially the input file name under which the created input matrix was saved. The user is then asked to assign a name to the output file, whether a printout of concentration data and the maximum number of iterations allowed for the algorithm. Once the number of iterations has been entered the solver will be executed and an output file created within the MCMsf Program File.

```

*****
Mixing Cell Model MCM (Double Precision )
*****
*** Type the name of Input File (data file)
name.txt
*** Type the name of Output File (results)
outname.txt

Do you want a printout of all concentration data
*** TYPE Yes = 1, No = 0
1
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
Input number of iterations for WOLF
100_

```

Figure A-13 The FORTRAN© solver interface

MCMsf Output

The resultant output file from the MCMsf four-cell example model run is:

```

*****
** Estimation of Flow into and between Cells and Transmissivities **
**           Program MCMsf-Wolf Algorithm           **
*****

Data_4_14
The date is: 03:20:2005
*****
*** Input File   In414.txt
*** Output File  Output414.
*****

There are 14 potential inputs.
There are 14 tracers to be considered.
There are 4 cells in this model.
There are 5 flows between the cells.

N = 14 NN = 18 NOC = 4 NOP = 14
Number of inflows to each cell:
  Num.cell 1 2 3 4
  Inflows  4 5 3 2
Number of internal flows from and into each cell:
  Num.cell 1 2 3 4
  Interfl. 2 3 3 2
The internal flows are:
  From cell 1 1 2 2 3
  To cell   2 3 3 4 4

```

The outflow out of the last cell is [m3/year]:

QOUT = 100.00

The rate of output (pumpage) and/or evapotranspiration from each cell is [m3/time]:

- PM(1) = .50
- PM(2) = 3.50
- PM(3) = 15.50
- PM(4) = 5.00
- Constant for Flows = 5.000
- PM(1) = .10
- PM(2) = .70
- PM(3) = 3.10
- PM(4) = 1.00

The weighting parameters are:

1.000 1.000 1.000 1.000 1.000 1.000 1.000 1.000 1.000 1.000 1.000 1.000 1.000 1.000
1.000

There are 60 mass balance equations.
Number of iterations: 28

The rest of the variables are equal to zero.

The unknown inflows are:

	Name of inflow	Rate of inflow	Perc. of tot. inflow	Perc. of cell inflow	flux
Cell 1					
1	Flow1	= 8.95	36.0 %	79.7 %	44.74
2	Flow2	= 1.48	6.0 %	13.2 %	7.42
3	Flow3	= .50	2.0 %	4.5 %	2.51
4	Flow4	= .30	1.2 %	2.6 %	1.48
Cell 2					
5	Flow5	= .80	3.2 %	22.5 %	3.98
6	Flow6	= .30	1.2 %	8.4 %	1.49
7	Flow7	= .83	3.3 %	23.4 %	4.14
8	Flow8	= 1.20	4.8 %	33.8 %	5.99
9	Flow9	= .42	1.7 %	11.9 %	2.11
Cell 3					
10	Flow10	= 3.00	12.1 %	49.2 %	15.01
11	Flow11	= 2.01	8.1 %	32.9 %	10.03
12	Flow12	= 1.09	4.4 %	17.9 %	5.45
Cell 4					
13	Flow13	= .00	.0 %	.0 %	.00
14	Flow14	= 4.01	16.1 %	100.0 %	20.05

Estimated fluxes for unknown inflows

The intermediate flows are:

From cell	To cell	Rate of flow	Real number
1	2	7.096	35.48
1	3	4.034	20.17
2	3	.981	4.91
2	4	8.973	44.86
3	4	8.012	40.06

} Inter-Cell flows

Total: 124.41 100.00 %
QOUT + PPP = 124.50

} Water balance error

Absolute diff.: .09
Percentage diff.: .07 %

**** Ion balance over the entire basin ****

Id	SUMIN	SUMOUT	Abs. error:	Perc. error:
TDS	24.770	24.780	-.01	.0%
Mg	25.649	25.636	.01	.1%
Ca	24.864	24.861	.00	.0%
Na	24.714	24.743	-.03	-.1%
K	23.534	23.517	.02	.1%
HCO3	24.723	24.731	-.01	.0%
Cl	24.587	24.581	.01	.0%
NO3	24.937	25.015	-.08	-.3%
SO4	26.264	26.265	.00	.0%
F	25.634	25.690	-.06	-.2%
Li	25.372	25.153	.22	.9%
Si	24.689	24.716	-.03	-.1%
D	24.987	25.010	-.02	-.1%
O-18	25.018	25.042	-.02	-.1%

} Individual tracer chemical mass balance errors

Total salt transport:
 Observed output (LB/DAY): 1887.5
 Estimated input (LB/DAY): 1887.6

} Overall chemical mass balance error

End of MCMsf Model.

From the output file of the four-cell example model the inflow from each of the defined sources into their respective Cells were calculated as well as the flows rates for defined inter-cell flows. The water balance percentage error is reported along with the chemical mass balance percentage errors for each individual tracer. An overall chemical mass balance error is also reported as total salt transport with observed versus estimated.

Appendix B

Table B-1 Model run data for UFS surface water – groundwater interaction test site application (24/01/2011). All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Name	pH	EC	Ca	Mg	Na	K	TAL	F	Cl	NO3	SO4
R 1	7.31	30.00	16.230	7.260	26.990	6.520	99.00	0.170	24.530	0.400	16.150
R 2	7.26	29.90	21.960	8.880	23.470	6.630	99.70	0.160	22.160	0.460	15.080
River Cell	7.29	29.95	19.095	8.070	25.230	6.575	99.35	0.165	23.345	0.430	15.615
o-River Cell	7.29	29.95	19.095	8.070	25.230	6.575	99.35	0.165	23.345	0.430	15.615
C5H039	8.11	23.33	18.182	6.975	15.267	4.795	79.35	0.299	15.737	0.489	10.991
BH1	7.45	96.90	39.320	40.650	113.100	5.100	411.00	0.560	57.000	0.370	25.650
BH2	7.36	90.10	39.420	40.090	93.290	8.230	381.00	0.490	54.000	0.820	21.870
BH3	7.57	109.00	36.310	50.060	157.340	5.590	589.00	0.670	70.000	0.180	31.270
BH4	7.48	89.10	39.950	43.300	100.120	5.420	381.00	0.510	53.000	0.370	21.320
BH5	7.46	105.00	35.540	56.600	123.660	5.580	462.00	0.620	61.000	0.000	26.310
BH6	7.45	99.30	41.660	48.910	114.490	5.580	426.00	0.550	60.000	0.230	25.760
BH7	7.43	100.00	34.350	49.500	111.730	5.700	430.00	0.570	60.000	0.080	26.050
BH8	7.68	105.00	35.650	47.350	124.830	5.600	436.00	0.550	69.000	0.250	32.020
BH9	7.55	104.00	37.550	45.180	121.540	5.310	424.00	0.570	70.000	0.320	34.050
BH10	7.52	93.40	37.290	42.370	103.350	5.260	394.00	0.540	57.000	0.340	24.820
BH11	7.61	94.90	34.620	41.430	108.990	5.500	398.00	0.530	59.000	0.330	26.410
BH12	7.63	93.60	36.320	41.590	106.680	5.400	395.00	0.490	57.000	0.310	24.870
BH13	7.63	94.50	35.470	42.210	108.560	5.410	396.00	0.510	59.000	0.370	26.560
BH14	7.46	88.30	40.940	42.280	86.500	4.750	380.00	0.410	51.440	0.510	20.560
BH15	7.49	86.50	40.450	41.530	84.140	4.850	373.00	0.360	48.840	0.510	20.020
SP	7.57	111.00	44.020	58.280	119.830	5.850	485.00	0.650	64.000	0.070	29.160

Table B-2 Model run data for UFS surface water – groundwater interaction test site application (05/08/2011). All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Name	pH	EC	Ca	Mg	Na	K	MAIk	F	Cl	SO4	Si
River 1	7.41	21.80	14.595	4.255	18.209	5.280	76.00	0.113	14.989	10.236	5.253
River 2	7.22	22.30	14.637	4.168	19.134	5.168	75.00	0.079	16.293	10.742	5.184
River Cell	7.32	22.05	14.616	4.211	18.672	5.224	75.50	0.096	15.641	10.489	5.219
o-River Cell	7.32	22.05	14.616	4.211	18.672	5.224	75.50	0.096	15.641	10.489	5.219
C5H039	7.99	22.91	23.559	7.557	16.787	4.431	84.15	0.418	15.594	11.632	6.431
BH1	7.53	107.00	44.471	48.253	122.343	5.557	443.00	0.550	86.176	37.235	20.435
BH2	7.72	84.80	34.481	41.926	89.049	5.052	376.00	0.516	54.000	21.319	21.721
BH3	7.50	108.00	36.251	51.339	132.011	5.615	477.00	0.375	71.772	28.159	23.648
BH4	7.46	87.70	36.664	41.400	94.401	5.313	392.00	0.554	56.000	22.060	21.791
BH5	7.25	168.00	53.622	107.197	172.241	7.310	613.00	0.512	206.021	82.338	23.225
BH6	7.30	102.00	45.229	53.007	114.453	6.102	449.00	0.401	70.726	28.009	23.684
BH7	7.31	109.00	43.493	63.526	123.158	6.590	470.00	0.246	84.179	31.530	23.932
BH8	7.54	124.00	39.770	59.230	143.891	6.038	509.00	0.555	97.549	53.305	23.031
BH9	7.36	165.00	56.389	82.750	209.587	6.612	539.00	0.734	204.496	105.043	20.254
BH10	7.33	91.90	41.908	47.327	108.947	6.111	407.00	0.618	59.000	25.806	22.192
BH11	7.37	93.70	35.749	42.731	108.666	5.684	410.00	0.501	63.000	25.578	21.483
BH12	7.51	93.60	36.871	44.146	106.863	5.610	408.00	0.456	65.000	25.819	21.756
BH13	7.36	92.50	35.968	41.626	103.871	5.527	411.00	0.428	59.000	24.291	21.079
BH14	7.40	86.60	45.459	45.852	86.420	4.997	388.00	0.351	53.000	20.639	21.641
BH15	7.44	84.80	41.044	41.705	82.906	5.008	383.00	0.328	51.028	22.154	21.274
Seep 1	7.52	135.00	54.446	74.174	139.166	5.737	533.00	0.265	106.088	74.832	21.846

Appendix C

Table C-1 Chemical tracer data for Section 1 (October 2012). All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Location	pH	EC	Ca	Mg	Na	K	MAIk	F	Cl	Br	SO4	TDS	Al	Fe	Mn	Si	Ba	Cu	Se	Zn
MR02	7.61	37.9	35.337	14.884	21.905	5.602	180	0.193	13.042	0.133	9.385	280.15	0.011	0.017	0.009	4.603	0.036	0.006	0.011	0.019
MR03	7.22	53.9	35.099	14.108	50.918	11.423	164	0.233	50.162	0.171	20.717	376.70	0.071	0.044	0.011	4.568	0.038	0.006	0.011	0.015
MR06-A	7.51	54.4	33.572	13.696	52.132	10.549	171	0.244	51.140	0.082	21.187	373.35	0.061	0.040	0.013	4.769	0.042	0.007	0.009	0.021
MR02 BH	7.48	50.3	49.713	22.522	27.097	0.930	245	0.215	9.673	0.017	20.290	377.73	0.009	0.018	0.009	12.621	0.007	0.003	0.013	0.048
MR03BH	7.94	46.2	24.210	2.838	71.667	0.538	203	1.597	9.515	0.075	28.584	342.24	0.010	0.012	0.053	7.984	0.003	0.004	0.010	0.072

Table C-2 Chemical tracer data for Section 1 (January 2013). All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Location	pH	EC	Ca	Mg	Na	K	MAIk	Cl	SO4	TDS	Si	B	Al	Ba	Cu	Fe	Mn	Zn
MR03	7.42	31.9	23.283	10.156	22.393	5.799	118	21.022	12.639	223	5.524	0.126	0.101	0.063	0.019	0.084	0.018	0.021
MR06ave	7.64	30.2	19.290	8.681	24.329	5.575	105	20.784	13.539	205	5.273	0.120	0.104	0.052	0.017	0.088	0.022	0.018
MR02	7.43	36.1	29.683	13.150	19.658	4.963	163	12.952	9.468	254	5.284	0.142	0.060	0.069	0.012	0.065	0.016	0.019
MR02BH	7.42	50.9	47.263	21.585	23.878	0.871	243	9.788	20.590	371	13.018	0.133	0.012	0.014	0.012	0.026	0.014	0.111
MR03BH	7.68	49.1	32.714	9.738	60.622	0.779	223	8.640	25.801	364	8.856	0.512	0.019	0.007	0.012	0.025	0.066	0.610
MR06B	7.73	47.2	30.701	14.095	41.556	6.259	160	43.657	18.251	316	5.051	0.120	0.102	0.094	0.013	0.090	0.025	0.018
MR06CBH	8.11	76.1	46.022	19.921	94.119	6.228	349	22.453	23.791	573	9.371	0.173	0.016	0.067	0.010	0.013	0.011	0.018
MR07BH	8.05	73.7	61.678	19.884	66.545	1.735	244	59.300	61.300	518	6.842	0.130	0.017	0.067	0.012	0.098	0.062	0.148

Table C-3 Chemical tracer data for Section 2 (October 2012). All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Location	pH	EC	Ca	Mg	Na	K	MAIk	F	Cl	Br	SO4	TDS	Al	Fe	Mn	Si	Ba	Cu	Se	Zn
MR11	8.09	50.4	32.17	17.39	42.009	8.40	190	0.209	39.37	0.202	17.70	350	0.010	0.014	0.016	0.952	0.049	0.004	0.010	0.014
MR12	7.85	69.0	41.87	24.69	68.489	9.59	225	0.220	62.10	0.267	37.12	479	0.008	0.017	0.010	3.272	0.034	0.004	0.011	0.018
MR13	7.56	70.9	43.74	21.24	67.844	11.4	212	0.204	63.90	0.204	43.07	507	0.047	0.035	0.012	5.378	0.027	0.006	0.012	0.017
MR12BH	8.27	280	101.5	169.1	317.811	1.17	678	0.244	508.28	1.470	226.80	2054	0.008	0.011	0.011	20.672	0.033	0.006	0.020	0.015

Table C-4 Chemical tracer data for Section 2 (January 2013). All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Location	pH	EC	Ca	Mg	Na	K	MAIk	F	Cl	SO4	TDS	Si	B	Al	Ba	Cu	Fe	Mn	Zn
MR11ave	7.56	20.4	15.953	6.546	14.418	4.896	88	0.274	9.345	7.322	148	5.701	0.083	0.128	0.048	0.013	0.185	0.016	0.018
MR12	7.57	24.8	20.029	9.539	16.141	4.830	104	0.228	12.197	11.608	180	7.014	0.076	0.104	0.052	0.011	0.121	0.016	0.016
MR12BH	7.83	301.0	60.765	138.893	375.110	0.997	629	0.099	535.527	235.319	2032	21.444	0.174	0.007	0.034	0.010	0.013	0.016	0.012
MR13	7.97	41.8	28.743	13.177	33.050	6.472	139	0.165	33.134	23.350	282	4.987	0.090	0.104	0.060	0.016	0.108	0.019	0.020

Table C-5 Chemical tracer data for Section 3 (October 2012). All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Location	pH	EC	Ca	Mg	Na	K	MAIk	F	Cl	Si
MR18-20	8.31	67.67	40.09	22.65	65.38	9.87	208.33	0.29	69.50	0.14
C5H039	7.81	64.6	48.124	22.033	56.291	8.359	174	0.445	54.817	0.607
MR21ave	8.17	102.6	44.519	49.704	120.431	6.347	404	0.488	67.680	18.707

Table C-6 Chemical tracer data for Section 3 (January 2013). All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Location	pH	EC	Ca	Mg	Na	K	MAIk	F	Cl	SO4	TDS	Si	B	Al	Ba	Cu	Fe	Mn	Zn
MR15BH	8.46	412	81.976	0.667	692.3	9.712	17	5.640	1297	1.621	2104	8.416	2.159	0.045	0.216	0.040	0.033	0.025	0.039
MR16-17	7.63	24.5	17.384	7.101	18.79	4.902	91	0.239	18.08	10.94	168	4.886	0.096	0.125	0.050	0.013	0.158	0.017	0.019
MR18-20	8.24	38.9	22.468	12.77	34.34	7.241	128	0.243	35.18	21.01	260	0.774	0.085	0.112	0.070	0.014	0.138	0.021	0.020
MR21BHave	7.86	106	45.956	46.90	115.1	5.913	409	0.264	64.58	41.52	756	19.095	0.181	0.020	0.149	0.009	0.044	0.024	0.019

Table C-7 Water Quality data for the continuous, gauged model run for the middle Modder River (MR01-MR20). All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Location	pH	EC	Ca	Mg	Na	K	MAIk	Cl	SO4	TDS	Si	B	Al	Ba	Cu	Fe	Mn	Zn
Rust Dam	7.92	37.9	22.011	8.689	16.530	5.486	122	8.872	10.518	195	2.049	0.158	0.145	0.067	0.013	0.200	0.018	0.018
MR01	7.58	30.2	23.293	9.661	18.470	5.667	137	8.804	8.806	212	1.122	0.151	0.128	0.062	0.013	0.111	0.019	0.018
MR02	7.43	36.1	29.683	13.150	19.658	4.963	163	12.952	9.468	254	5.284	0.142	0.060	0.069	0.012	0.065	0.016	0.019
MR02BH	7.42	50.9	47.263	21.585	23.878	0.871	243	9.788	20.590	371	13.018	0.133	0.012	0.014	0.012	0.026	0.014	0.111
MR03	7.42	31.9	23.283	10.156	22.393	5.799	118	21.022	12.639	223	5.524	0.126	0.101	0.063	0.019	0.084	0.018	0.021
MR03BH	7.68	49.1	32.714	9.738	60.622	0.779	223	8.640	25.801	364	8.856	0.512	0.019	0.007	0.012	0.025	0.066	0.610
MR06ave	7.64	30.2	19.290	8.681	24.329	5.575	105	20.784	13.539	205	5.273	0.120	0.104	0.052	0.017	0.088	0.022	0.018
MR06B	7.73	47.2	30.701	14.095	41.556	6.259	160	43.657	18.251	316	5.051	0.120	0.102	0.094	0.013	0.090	0.025	0.018
MR06CBH	8.11	76.1	46.022	19.921	94.119	6.228	349	22.453	23.791	573	9.371	0.173	0.016	0.067	0.010	0.013	0.011	0.018
MR07Bridge	7.69	34.3	22.179	9.765	29.240	5.431	121	25.425	15.655	235	5.297	0.106	0.116	0.055	0.016	0.091	0.074	0.023
MR07BH	8.05	73.7	61.678	19.884	66.545	1.735	244	59.300	61.300	518	6.842	0.130	0.017	0.067	0.012	0.098	0.062	0.148
MR08	7.57	24.3	17.078	7.514	16.836	5.070	99	11.811	8.662	167	5.366	0.092	0.151	0.049	0.012	0.191	0.019	0.020
MR10	7.44	20.7	16.378	7.023	15.656	5.115	91	9.841	7.800	154	5.550	0.087	0.111	0.048	0.013	0.151	0.019	0.016
Koranaspruit	7.82	21.9	18.255	7.741	14.828	3.558	104	5.540	6.072	160	6.178	0.078	0.204	0.054	0.013	0.291	0.023	0.017
MR11ave	7.56	20.4	15.953	6.546	14.418	4.896	88	9.345	7.322	148	5.701	0.083	0.128	0.048	0.013	0.185	0.016	0.018
MR12	7.57	24.8	20.029	9.539	16.141	4.830	104	12.197	11.608	180	7.014	0.076	0.104	0.052	0.011	0.121	0.016	0.016

Application of the Mixing Cell Model to the quantification of groundwater – surface water interaction

MR12BH	7.83	301.0	60.765	138.893	375.110	0.997	629	535.527	235.319	2032	21.444	0.174	0.007	0.034	0.010	0.013	0.016	0.012
MR13	7.97	41.8	28.743	13.177	33.050	6.472	139	33.134	23.350	282	4.987	0.090	0.104	0.060	0.016	0.108	0.019	0.020
MR14	7.96	40.7	28.775	13.020	30.280	5.883	145	30.207	21.648	277	6.208	0.084	0.156	0.083	0.014	0.210	0.022	0.018
MR14BH	7.95	87.7	33.746	25.709	121.136	4.913	316	53.245	36.364	647	17.687	0.443	0.018	0.097	0.015	0.014	0.012	0.026
MR15	7.90	34.3	22.787	10.085	27.134	5.454	113	27.572	16.324	224	5.317	0.087	0.153	0.068	0.014	0.263	0.031	0.022
MR15BH	8.46	412.0	81.976	0.667	692.357	9.712	17	1297.000	1.621	2104	8.416	2.159	0.045	0.216	0.040	0.033	0.025	0.039
MR16-17	7.63	24.5	17.384	7.101	18.793	4.902	91	18.089	10.945	168	4.886	0.096	0.125	0.050	0.013	0.158	0.017	0.019
MR18-20	8.24	38.9	22.468	12.779	34.342	7.241	128	35.185	21.016	260	0.774	0.085	0.112	0.070	0.014	0.138	0.021	0.020
MR21BH1	7.85	108.0	48.055	47.889	112.505	5.706	414	66.224	42.319	768	19.688	0.183	0.024	0.154	0.008	0.048	0.019	0.019
MR21BH2	7.81	112.0	49.228	51.111	121.393	6.125	433	65.898	48.213	801	19.656	0.188	0.020	0.186	0.009	0.020	0.019	0.022
MR21BH3	7.82	103.0	43.626	45.336	113.723	5.869	389	65.658	39.084	734	19.684	0.180	0.024	0.128	0.009	0.024	0.020	0.019
MR21BH4	7.96	101.0	43.539	43.797	105.653	5.778	386	64.824	36.386	712	16.889	0.168	0.018	0.130	0.010	0.113	0.048	0.018
MR21BH5	7.84	106.0	45.332	46.381	122.339	6.086	421	60.323	41.611	765	19.557	0.187	0.015	0.146	0.009	0.016	0.015	0.018
MR21BHave	7.86	106.0	45.956	46.903	115.122	5.913	409	64.585	41.523	756	19.095	0.181	0.020	0.149	0.009	0.044	0.024	0.019
MR21BHP	8.39	179.0	37.749	63.382	235.315	20.374	506	285.182	8.917	1159	3.451	0.307	0.021	0.098	0.010	0.030	0.016	0.012
MR Weir	7.60	27.8	20.171	8.446	21.443	5.657	101	20.248	13.163	192	4.057	0.082	0.152	0.053	0.013	0.281	0.022	0.017
Renoster Weir	8.28	50.8	32.119	13.570	44.529	8.722	150	48.694	31.898	340	3.437	0.095	0.041	0.052	0.014	0.034	0.018	0.017

Table C-8 Chemical mass balance percentage error for each tracer in each section model run.

Chemical Mass Balance Errors (%)										
Tracer	Section 1		Section 2				Section 3			
	Jan '13		Oct '12		Jan '13		Oct '12		Jan '13	
	Scenario 1	Scenario 2	Scenario 1	Scenario 2	Scenario 1	Scenario 2	Scenario 1	Scenario 2	Scenario 1	Scenario 2
pH	-20.10%	-22.80%	-5.80%	4.40%	-16.90%	-2.40%	-4.20%	-2.80%	-78.10%	-48.20%
EC	9.70%	3.40%	-11.20%	-9.80%	-15.50%	-8.10%	-4.30%	-3.00%	-64.60%	-46.40%
Ca	27.40%	22.00%	-22.50%	-18.10%	-41.70%	-33.20%	-24.70%	-23.60%	-75.90%	-51.50%
Mg	26.40%	21.20%	30.00%	26.00%	6.10%	13.70%	-4.30%	-3.20%	-88.00%	-69.90%
Na	-3.70%	-13.00%	-12.50%	-13.40%	6.20%	12.90%	7.10%	8.40%	-46.60%	-33.10%
K	-21.30%	-24.80%	-40.50%	-32.50%	-37.60%	-26.00%	5.30%	7.00%	-82.70%	-61.00%
MAIk	25.20%	21.10%	-2.50%	2.20%	-20.40%	-10.70%	10.60%	12.00%	-84.50%	-61.40%
F			-8.80%	0.60%	38.50%	63.90%	-39.30%	-38.50%	-31.10%	-4.00%
Cl	-0.30%	-13.70%	13.80%	7.60%	23.70%	28.10%	14.20%	15.90%	-12.90%	-4.00%
B	-9.40%	-10.30%							-23.10%	8.50%
Br			37.50%	36.60%						
SO4	-19.10%	-28.00%	-24.40%	-28.50%	-11.70%	-6.80%			-88.70%	-71.80%
TDS	10.70%	4.80%	-12.10%	-11.10%	-12.60%	-4.50%			-69.40%	-50.10%
Al	-41.00%	-44.90%	-81.30%	-79.20%	0.90%	19.90%			-75.20%	-39.00%
Fe	-31.50%	-35.60%	-66.40%	-62.50%	39.70%	65.80%			-75.00%	-37.80%
Mn	-23.80%	-29.50%	19.40%	33.40%	-25.70%	-12.70%			-80.70%	-55.30%
Si	-20.10%	-22.30%	-54.60%	-59.00%	19.60%	37.20%	-2.30%	-8.90%	57.80%	260.80%
Ba	24.00%	17.20%	52.70%	70.70%	-32.50%	-20.40%			-78.20%	-55.50%
Cu	-41.60%	-43.70%	-37.80%	-31.80%	-29.90%	-17.30%			-73.90%	-43.70%
Zn	-3.20%	-3.80%	-25.60%	-17.70%	-20.30%	-6.00%			-75.20%	-44.10%
Se			-15.10%	-8.10%						
Average	-6.2%	-11.3%	-13.4%	-9.5%	-7.2%	5.2%	-4.2%	-3.7%	-60.3%	-26.7%

Appendix D

Table D-1 MCM input data for the A42 area in 2005. All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Location	Ca	Cl	EC	F	K	Mg	Na	NO3	pH	PO4	SO4	TAL
A4H2Cell	5.717	6.412	7.95	0.123	1.001	2.565	4.987	0.123	7.41	0.013	2.000	22.81
o-A4H2 Cell	5.717	6.412	7.95	0.123	1.001	2.565	4.987	0.123	7.41	0.013	2.000	22.81
A4H002c	6.923	13.770	10.57	0.050	0.940	3.773	7.409	0.040	7.74	0.018	2.000	20.43
B89783	17.569	16.267	40.50	0.438	1.041	6.327	60.595	2.739	8.12	0.012	9.788	154.34
B89784	21.489	5.632	26.00	0.142	1.336	12.297	9.740	0.212	8.47	0.014	2.000	110.96

Table D-2 The individual tracer chemical mass balance percentage error for the MCM run for A42 area.

Tracer	Mass balance error (%)					
	Outflow (Jan '05)		Outflow (Sum 2005)		Outflow (Sum 2006)	
	Scenario 1	Scenario 2	Scenario 1	Scenario 2	Scenario 1	Scenario 2
Ca	-6.40%	-7.30%	-6.40%	-7.30%	-6.40%	-7.30%
Cl	24.20%	40.60%	24.20%	40.60%	24.20%	40.60%
EC	2.40%	3.80%	2.40%	3.80%	2.40%	3.80%
F	-62.40%	-62.70%	-62.40%	-62.70%	-62.40%	-62.70%
K	-40.30%	-36.00%	-40.30%	-36.00%	-40.30%	-36.00%
Mg	11.70%	9.90%	11.70%	9.90%	11.70%	9.90%
Na	21.70%	27.80%	21.70%	27.80%	21.70%	27.80%
NO3	-13.20%	-17.90%	-13.20%	-17.90%	-13.20%	-17.90%
pH	-35.70%	-30.00%	-35.70%	-30.00%	-35.70%	-30.00%
PO4	-18.80%	-9.80%	-18.80%	-9.80%	-18.80%	-9.80%
SO4	-28.70%	-23.30%	-28.70%	-23.30%	-28.70%	-23.30%
TAL	-6.40%	-14.00%	-6.40%	-14.00%	-6.40%	-14.00%
Average	-12.7%	-9.9%	-12.7%	-9.9%	-12.7%	-9.9%

Appendix E

Table E-1 Water quality data used in the MCM application for the quaternary catchment D73F. All values expressed in mg/l, except EC expressed in mS/m and pH expressed in pH units.

Location	Ca	Cl	DMS	EC	F	K	Mg	Na	NH4	NO3_NO2	pH	PO4	Si	SO4	TAL
CellD7H3	22.500	17.900	195	26.5	0.180	1.800	9.000	17.000	0.080	0.300	7.85	0.024	8.040	16.200	89.3
InflowD7H5	21.500	20.500	177	23.9	0.250	1.910	8.600	16.000	0.060	0.580	8.00	0.023	7.780	17.900	72.0
B166744	66.800	77.100	795	99.5	3.740	1.050	25.600	128.000	0.020	0.020	7.73	0.015	23.370	121.700	304.2
167032	103.100	409.100	1854	255.0	6.600	16.000	40.400	421.700	0.040	0.020	7.55	0.016	15.240	313.500	445.9
B167031	109.100	280.000	1279	181.0	4.910	1.150	52.700	207.600	0.040	2.390	7.68	0.011	19.430	218.300	323.2
B166928	192.600	506.300	2013	282.2	4.410	4.630	77.700	339.550	0.040	6.090	7.70	0.028	17.760	410.150	369.8
B166929	199.800	500.000	2031	288.0	4.310	10.760	80.200	346.100	0.450	5.600	7.60	0.033	16.190	397.300	383.0
B166930	91.800	182.700	733	114.0	3.170	3.800	37.700	78.300	0.640	4.060	7.10	0.016	11.590	137.900	146.7
B166931	143.200	407.400	1874	252.8	5.080	3.810	57.700	362.000	0.080	4.960	7.80	0.031	16.690	414.400	375.8
B166932	109.800	161.300	1216	166.0	5.970	2.030	46.200	178.800	0.360	4.960	7.70	0.046	17.970	187.600	411.6
B166933	97.600	122.000	1020	135.0	6.280	4.290	39.900	139.300	0.070	3.380	7.60	0.041	22.930	106.900	400.4
B166934	96.200	33.200	892	109.0	6.410	3.690	47.700	82.300	0.070	1.550	7.40	0.073	27.630	202.600	338.6
B166935	118.100	192.400	1247	174.0	5.550	1.700	43.300	194.900	0.060	2.340	7.70	0.037	21.430	198.600	395.6
B166990	25.900	27.500	827	102.4	12.040	10.600	7.200	192.300	0.020	11.840	7.87	0.063	30.240	57.400	362.2
B167048	132.975	281.525	1387	184.8	4.745	1.220	53.250	215.350	0.088	2.405	7.85	0.027	17.985	221.725	381.7
B167049	183.050	503.300	1960	273.0	4.315	5.785	75.500	335.000	0.085	7.090	7.90	0.025	17.625	381.500	360.4

Table E-2 The individual tracer chemical mass balance percentage error for the MCM run for quaternary catchment D73F.

Tracer	Chemical mass balance error (%)			
	No abstraction volume		Abstraction volume	
	Scenario 1	Scenario 2	Scenario 1	Scenario 2
Ca	-19.90%	-11.30%	-19.90%	-11.60%
Cl	-1.90%	6.30%	-1.90%	5.90%
DMS	-23.30%	-15.80%	-23.30%	-16.10%
EC	-23.80%	-16.30%	-23.80%	-16.60%
F	20.00%	28.90%	20.00%	28.40%
K	-10.60%	-1.50%	-10.60%	-1.90%
Mg	-19.90%	-11.30%	-19.90%	-11.60%
Na	-18.70%	-12.70%	-18.70%	-13.00%
NH4	-37.50%	-30.40%	-37.50%	-30.70%
NO3_NO2	60.90%	79.40%	60.90%	78.80%
pH	-15.10%	-5.40%	-15.10%	-5.80%
PO4	-20.10%	-11.10%	-20.10%	-11.40%
Si	-19.20%	-10.20%	-19.20%	-10.50%
SO4	-5.70%	2.50%	-5.70%	2.20%
TAL	-32.30%	-25.20%	-32.30%	-25.50%
Average	-11.1%	-2.3%	-11.1%	-2.6%

Summary

The significance of a reliable groundwater resource assessment is of growing importance as water resources are stretched to accommodate the growing population. An essential component of a groundwater resource assessment is the quantification of surface water – groundwater interaction. The insufficient amount of data in South Africa and the apparent lack of accuracy of current estimates of the groundwater component of baseflow lead to the investigation of a new methodology. The applicability of the Mixing Cell Model (MCM) to quantify the groundwater contribution to baseflow is examined to determine whether the method would be of use in groundwater resource assessments. The MCM simultaneously solves water and solute mass balances to determine unknown inflows to a system, in this application the groundwater component of baseflow. The incorporation of water quality data into the estimation of the surface water – groundwater interaction increases the use of available data, and thus has the ability to decrease the uncertainty of the estimation process. The balance equations are equated to an error term which is used in the quadratic programming solution of minimizing the square error sums in order to determine the unknown inflows. The mixing cell model is applied to datasets from the surface water – groundwater interaction test site developed by the University of the Free State, in addition to data collected along the middle Modder River during a fieldwork survey. The MCM is subsequently applied to a set of quaternary catchments in the Limpopo Province for which there are available calibrated estimates of the groundwater component of baseflow for the Sami and Hughes models. The MCM is further applied to the quaternary catchment D73F, located in the semi-arid Northern Cape, to assess the applicability of the mathematically based MCM in terms of a flow system located within a regionally-defined zero groundwater baseflow zone. The MCM results for each study area are assessed in comparison to groundwater baseflow volumes determined by the Pitman, Sami and Hughes models. A chemical hydrograph separation method which also incorporates water quality data is additionally reported for the study areas to further validate the MCM. The results indicate that the mixing cell model can reliably estimate the groundwater component of baseflow to a river. This application of the mixing cell model could contribute to increase and evaluate the accuracy of current groundwater baseflow estimates in South Africa, which will in turn ensure the responsible and sustainable use of the countries water resources.

Keywords: surface water – groundwater interaction, mixing cell model, groundwater baseflow, quantification, groundwater resource assessment (GRAII).

Opsomming

Die noodsaaklikheid van die betroubare assessering van grondwaterhulpbronne is van toenemende belang aangesien hulpbronne gerek word om in die behoeftes van die groeiende bevolking te voorsien. 'n Wesenlike komponent van die assessering van grondwaterhulpbronne is die kwantifisering van oppervlakwater-grondwater-interaksie. Die ontoereikende hoeveelheid data in Suid-Afrika en die skynbare gebrek aan akkuraatheid van huidige ramings van die grondwaterkomponent van basisvloei het tot die ondersoek van 'n nuwe metodologie gelei. Die toepaslikheid van die Gemengde Selmodel (GSM) om die bydrae van grondwater tot basisvloei te kwantifiseer, is in hierdie studie ondersoek om te bepaal of die metode nuttig in toekomstige assessering van grondwaterhulpbron sal wees. Die GSM los water- en opgeloste massabalans gelyktydig op om onbekende invloei in 'n stelsel te bepaal, in hierdie geval die grondwaterkomponent van basisvloei. Die inkorporering van watergehaltedata in die raming van oppervlakwater-grondwater-interaksie gee geleentheid vir groter gebruik van beskikbare data en het dus die vermoë om die vertrouwe in die ramingsproses te versterk. Die balansvergelykings word gelykstel aan 'n foutterm wat gebruik word in die oplossing van kwadratiese programmering waar die kwadrate van die foutsomme geminimaliseer word ten einde die onbekende invloei te bepaal. Die GSM is toegepas op datastelle van die oppervlakwater-grondwater-interaksietoetsterrein wat by die Universiteit van die Vrystaat ontwikkel is, tesame met data wat gedurende 'n veldwerkopname langs die Middel-Modderivier versamel is. Die GSM is gevolglik toegepas op kwarternêre opvangsgebiede in die Limpopo-provinsie waarvoor daar gekalibreerde ramings van die grondwaterkomponent van basisvloei vir die Sami- en Hughes-modelle beskikbaar is. Die GSM is verder toegepas op die kwarternêre opvangsgebied D73F, geleë in die semi-ariëde Noord-Kaap, om die geskiktheid van die wiskundiggebaseerde GSM te assesser ten opsigte van 'n vloeistelsel wat binne 'n streeksgedefinieerde nul-grondwater-basisvloei-sone geleë is. Die GSM-resultate vir elke studiegebied is geassesseer in vergelyking met grondwater-basisvloei-volumes soos bepaal deur die Pitman-, Sami- en Hughes-modelle. Vir die studiegebiede is daar ook verslag gelewer oor 'n chemiese hidrograaf-skeidingsmetode, wat ook watergehaltedata inkorporeer, ten einde die GSM verder te bevestig. Die resultate dui dat die GSM op 'n betroubare wyse die grondwaterkomponent van basisvloei van 'n rivier kan raam. Hierdie toepassing van die GSM kan bydra tot die verbetering en evaluasie van die akkuraatheid van huidige grondwater-basisvloei-ramings in Suid-Afrika, wat op sy beurt die verantwoordelike en volhoubare gebruik van die land se waterhulpbronne sal verseker.

