

Hydrogeological modelling for future scenarios in the Far West Rand

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ABSTRACT

The Far West Rand (FWR) characterised by its karstified dolomites and dewatering related gold mining activities, extends along the Witwatersrand mining region. These mining activities have directly impacted the characteristics of the dolomites, altering the groundwater flow pathways, accelerating sinkhole development and causing the regional high yielding springs to dry up. Subsequently changing the natural flow regime and hydrogeological characteristics of the dolomitic aquifer.

This study implements a simplified water budget method to model future rewatering scenarios for the karstified dolomitic compartments of the FWR. The Saturated Volume Fluctuation (SVF) method was selected based on its adaptiveness for estimating present-day aquifer conditions as well as simulating recovery scenarios solely based on the average groundwater levels from monitoring boreholes, unlike similar water budget models. Due to the general lack in present-day monitoring data, the SVF's lumped parameter approach provided easy calibration based on simulated water level response. Six SVF implemented equations each representing a separate compartment was calibrated over an 11-month period. Aquifer parameters were adjusted to simulate groundwater level responses within the compartments to best approximate the observed values over the same time period based on the historical rainfall sequence.

To simulate recovery scenarios for each of the dewatered dolomitic compartments, SVF models were created based on the calibrated model parameters. The recovery scenarios were based on forecasted rainfall sequences created from historical data. Each of the compartments rewatered to near initial groundwater levels with inter-compartmental groundwater flow (IGF) having little effect on the overall rewatering. This resulted in spring reactivation with spring flow volumes lower than historical records. The model has demonstrated that a water budget model can simulate water level recovery in a post-mining scenario.

Keywords: Post-mining spring flow, Karstified dolomitic compartments, Water budget, Far West Rand

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LIST OF ABBREVIATIONS

ACRU	Agriculture Catchment Research Unit
APLIS	Altitude, Pendant or Slope, Lithology, Infiltration, Soil
CMB	Chloride Mass Balance
CWRR	Centre for Water Resources Research
DWA	Department of Water Affairs
DWAF	Department of Water Affairs and Forestry
EPANET	Environmental Protection Agency Network
ESE	East South East
FWR	Far West Rand
GM	Gold Mine
GIS	Geographical Information System
GRAII	Groundwater Resources Assessment Phase II
IGF	Inter-compartmental Groundwater Flow/flux
mamsl	Meters above mean sea level
mbgl	Meters below ground level
MLU	Multi-Layer Unsteady State
MODFLOW	Modular finite-difference groundwater flow model
NGI	National Geo-spatial Information
NNE	North North East
NGA	National Groundwater Archive
NSC	Nash-Sutcliffe Coefficient
RMSE	Root Mean Square Error
SAWS	South African Weather Services
SSW	South South West
SVF	Saturated Volume Fluctuation
SWMM	Storm Water Management Model
TSF	Tailings Storage Facility
WFS	Wonderfonteinspruit
WMA	Water Management Area
WNW	West North West
WR	West Rand
WR2012	Water Resources 2012
WRC	Water Research Commission
WWL	West Wits Line

NOMENCLATURE

km	Kilometre
km ²	Area in square kilometres
m	Metre
m ²	Square metres
m ³	Cubic metres
m/s	Meters per second
mamsl	Meters above mean sea level
mbgl	Meters below ground level
mg/m ² /a	Milligram per square meter per area
MI	Megalitre
MI/d	Megalitre per day
mm	Millimetre
mm/a	Millimetre per annum
Mm ³	Million cubic metre
Sq. mile	Square mile

CHAPTER 1:

INTRODUCTION

1.1 Introduction

In semi-arid South Africa, some of the largest aquifer systems are represented by karstified dolomites. The karstified dolomites of the Far West Rand (FWR) and West Rand (WR) are best known for hosting the largest groundwater resources in South Africa with their renowned high yielding karst springs. Numerous geological events shaped these dolomitic aquifers over millions of years with the addition of rich mineral deposits along the Witwatersrand. Among these deposits, gold was the largest commodity and subsequent mining thereof commenced in 1886. The gold mining era boosted the economy of South Africa, enabling vast growth for the country. The mining however came at a cost, due to the largest section of the gold-bearing reefs occurring beneath the karstified dolomitic compartments, flooding occurred regularly as the mining progressed. Since mining had great importance to the economic well-being of South Africa, the water resource and environmental issues surrounding mining were a lower priority and consequently the impacts of dewatering of the dolomitic aquifers (Schrader & Winde, 2014; Schrader, *et al.*, 2014b; Stoch & Winde, 2010).

The dolomitic springs on which many farmers depended for water in the early 1900's soon dried up as a result of dewatering caused by gold mining activities (Stoch & Winde, 2010). The mining-related impacts not only depleted spring flow due to compartmental dewatering of the aquifer system to enable deep-level mining but also affected the streamflow of the main river, the Wonderfontein spruit (WFS). This was mainly due to the diversion of its natural flow through an artificial pipeline in the western parts of the FWR leaving the river in a rather non-perennial state (De Klerk, 2019; Schrader & Winde, 2014; Schrader, *et al.*, 2014a; Schrader, *et al.*, 2014c). The dewatering activities lead to heightened chemical weathering of the dolomite resulting in the formation of sinkholes, enlarged fissures and fault-planes together with mined out voids within the rock body which in turn changes the aquifer characteristics (Enslin, 1967; Schrader, *et al.*, 2014c).

1.2 Problem statement

Ford & Williams (2007) estimated that 25% of the global population is most likely supplied by freshwater from karst aquifers. These aquifers account for 20% of the earth's ice-free land as surface and sub-surface soluble karstic rock. Karst Aquifers characterized by their dolomitic geology are some of the most abundant sources of groundwater resources dominated by sinkholes, caves and high yielding springs. These aquifer systems with their complex hydrogeology prove themselves difficult to manage, with methods to determine their characteristics scarce (Enslin, 1967; Schrader, *et al.*, 2014b)

Due to South Africa's growing irregular and spatially diverse rainfall patterns, stable sources of freshwater are an ever-increasing problem. Coupled with increased usage and pollution of the available surface water resources, access to fresh clean water will give rise to a national crisis (Basson & Rossouw, 2003). With mining slowly coming to an end as deep-level gold mining becomes less economical within the historical Witwatersrand region of South Africa, rewatering of these vast karstic aquifers will follow suit as more and more mines are abandoned. In 1977 the appointed Steering Committee, in response to the development of a post-mining master plan for the West Wits Line agreed:

"...that because the future development of the Far West Rand, particularly after mining in the area ceases, (would be) totally dependent on the amount and quality of the water available, and since the demands on the Vaal River System are already exceeding the abilities of that system with the result that the Far West Rand will have to rely on its own sources your Commission be requested to undertake an investigation to determine:

- *the amount of water which can be obtained on a sustained basis from the Dolomitic compartments in the Wonderfontein Spruit catchment area; and*
- *the suitability of the water for various industrial and other urban uses".*

As mentioned by The Steering Committee of 1983 (cited by Stoch & Winde, 2010:86).

This prompted various researchers across the years to research and model how and when these dolomitic aquifers will rewater and if the people within the catchment area can rely once more on the high yielding springs that once flowed within the FWR for water supply. Researchers have made use of several modelling approaches by either mathematically or numerically modelling the region and even combining approaches, with rewatering estimates ranging between 3 – 60 years' time, if they rewater at all (De Klerk, 2019; Schrader, *et al.*, 2014c; Stoch & Winde, 2010; Wolmarans, 1984).

The area has been delineated by numerous researchers across the years based on their overall research aim with no “one size fits all” delineation. Researchers such as Wolmarans (1984); Winde & Stoch (2010); Schrader, *et al.* (2014b); Swart, *et al.* (2003a) and Enslin & Kriel (1968) to name a few, have referred to the area as the Far West Rand based on its strategic location along the historical West Wits Line (WWL) – gold mining region and Carletonville goldfields. The FWR in part has also been referred to as the Wonderfonteinspruit (WFS) Catchment/ Valley in hydrological regards by the likes of De Kock (1964); Swart, *et al.* (2003b); Krige (2006) and De Klerk (2019).

In this study, the study area is referred to as the FWR based on a hydrogeological perspective delineated by the well-known Malmani dolomitic compartments. Consisting of the six dolomitic compartments: Zuurbekom, Gembokfontein, Venterspost, Bank, Oberholzer and Turffontein, the study area is similar to that depicted by Morgan & Brink (1984) in Figure 1:

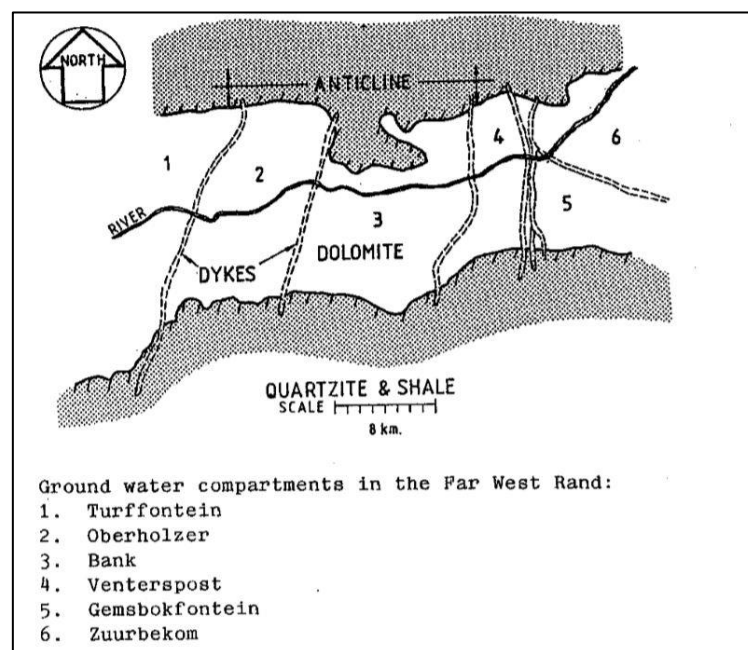


Figure 1: The FWR as depicted by Morgan & Brink (1984)

The study area’s boundaries are delineated by the presence of natural boundary conditions. These boundaries include impermeable dykes present throughout the catchment area with the Hartebeesfontein Anticline to the north, serving as a catchment divide. The study area not only includes the bulk aquifer system delineated by the dolomitic area but also the associated non-dolomitic runoff/ allogenic recharge areas.

The complex dolomitic systems of the FWR prove difficult to model as the aquifer properties have been significantly affected by the ongoing mining operations as well as dewatering. These effects have increased the complexity of how the system functions as a whole as well as on a smaller local scale. These factors include the state to which the fissures and fault-

planes within the compartments have been weathered as well as how other prominent karst features affect the system (Enslin, 1967; Swart, *et al.*, 2003b).

The complexity develops even more as the karstified dolomites of the FWR are divided into a range of dolomitic compartments through several syenite dykes that protrude the dolomitic body (Swart, *et al.*, 2003a). This causes each dolomitic compartment to function individually based on its degree of weathering, karst features, geomorphology and changes within that compartment. In addition, before gold mining started in the region each dolomitic compartment had its own groundwater level controlled by a spring that's discharge flowed directly onto the neighbouring downstream compartment (Swart, *et al.*, 2003a). Extensive gold mining not only generated voids below the dolomite but also caused perforations within the impermeable dykes causing the compartments to become hydraulically connected. Each compartment can therefore directly impact its downstream neighbour in the form of groundwater passing through mine voids within the compartment boundary dykes left by the mining activities. Changes in these characteristics have led to uncertainty as to future spring flow volumes once the groundwater levels recover in a post-mining environment (Schrader *et al.*, 2014b; Schrader *et al.*, 2014c).

With a growing demand for clean fresh water in South Africa, improved water management in the future is fundamental. The statement made by The Steering Committee as cited by Stoch & Winde (2010:86) highlighted the importance of understanding how mining operations have affected the FWR and how the dolomitic compartments of the FWR will react to rewatering during post-mining times. Determining the effects that dewatering activities have had on the various compartment areas' properties is crucial for understanding how each compartment's water level will recover. By successfully modelling these dolomitic compartments and calibrating each based data from current mining conditions, future post-mining scenarios could be modelled to simulate the aquifers' response when dewatering has ceased.

Although various studies have been conducted in estimating recovery scenarios for the FWR, few have taken an approach based on the water budget modelling approach where each dolomitic compartment is represented by its own budget and has an influencing effect on its neighbouring compartments.

1.3 Research question and objectives

The assessment of how dewatering activities have impacted the overall characteristic of the FWR and possible prediction of how these compartments would react in the future is of great importance as mentioned by Stoch & Winde (2010). In response to this statement as well as improving on previous research conducted in the area, the aim of this study is to simulate groundwater level recovery and spring reactivation through a water budget model in a post-mining rewatering scenario. To achieve this aim, the following research question is postulated:

Can a water budget model simulate groundwater level recovery during rewatering scenarios?

To successfully address the aim set out in this study and answering the research question, the following objectives have been compiled:

- Reviewing previous research conducted on the FWR for a better understanding of the mechanisms involved.
- To compare different methods in estimating recharge over the FWR.
- To create and calibrate a water budget for each dolomitic compartment based on current aquifer conditions.
- To simulate rewatering scenarios using a calibrated model for groundwater level recovery and spring flow estimation.

CHAPTER 2: DESCRIPTION OF STUDY AREA

2.1 Introduction

The area portrayed as the Far West Rand sits on the border of the North West and Gauteng provinces surrounded by major highway infrastructure. The study area begins in the headwaters of the Wonderfonteinspruit river near the city of Roodepoort and stretches along the Wonderfonteinspruit-Mooirivierloop river ending 40 km west of Carletonville. At this point, it merges with the Mooirivier river flowing from the northern Klerkskraal Dam and finally ending as one river within Boskop Dam (Figure 2). Consisting of the six dolomitic compartments as delineated by DWA (2019) the study area spans an approximate area of 2090km².

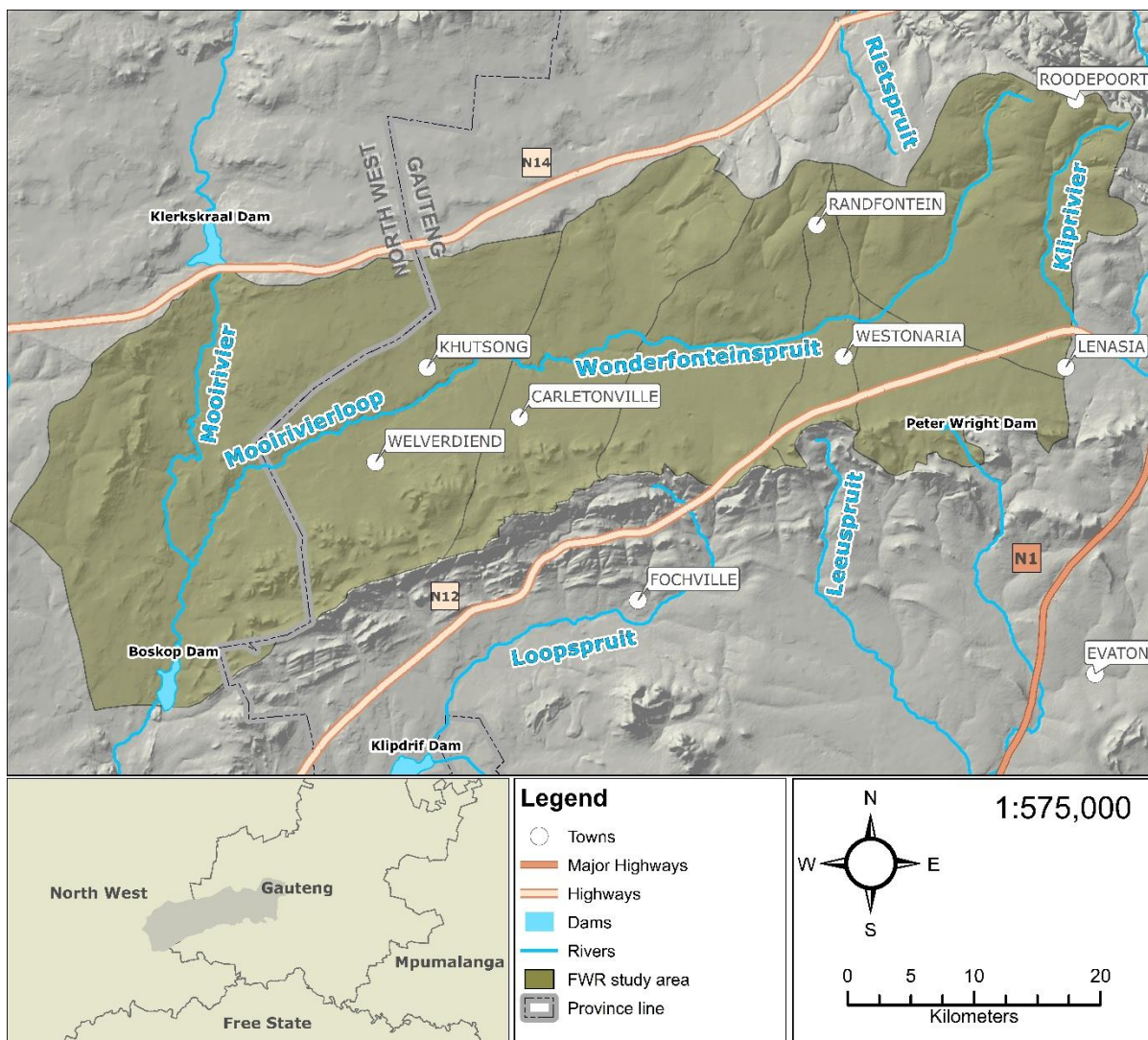


Figure 2: Locality map of the study area.

2.2 Historical overview of the study area

As early as 1871 human settlements described the area of the Wonderfonteinspruit Valley together with its characteristic karst environment consisting of caves, sinkholes and freshwater springs. During early development, the abundance of readily available freshwater was exploited through a thriving farming community that included agricultural commodities such as tobacco, fruit, corn and wheat as well as animal products including meat, skin and ostrich feathers (De Klerk, 2019; Swart, *et al.*, 2003b).

The mining of gold within the study area can be traced back to 1884, after two brothers Fred and Henry Struben established Sterkfontein Junction Mining Syndicate on several farms in the middle of the Highveld plateau. Situated along the quartzite ridge known as the Witwatersrand, these farms led to the rapid urbanisation of the West Rand (WR) and the development of one of Africa's biggest metropolitans, Johannesburg (Robb & Robb, 1998; Winde & Stoch, 2010).

Between 1899 and 1904 exploration drilling was conducted in the Carletonville goldfield, also known as the West Wits Line. This goldfield was the richest of all the Witwatersrand goldfields but the development of the area containing auriferous conglomerates was hampered by severe groundwater problems (Robb & Robb, 1998).

During 1911, attempts to mine the goldfields of the FWR south of the WR proved difficult and failed initially as waters from the dolomitic karst aquifers flooded the newly sunken shafts (Winde & Stoch, 2010). As technology developed in the field of cementation, the Venterspost Gold Mine succeeded in sinking a shaft in 1934, this was soon followed by other mines established in the FWR. Table 1 includes the registration dates of various mining companies registered in the West Wits Line.

Table 1: Registration dates of mine companies in the FWR (Winde & Stoch, 2010).

Date		Mining Company
05-Jun	1934	Ventespost Gold Mining Company Limited
04-Apr	1936	Libanon Gold Mining Company Limited
10-Jun	1937	Blyvooruitzicht Gold Mining Company Limited
07-Mar	1945	*West Driefontein Gold Mining Company Limited
25-Feb	1947	Doornfontein Gold Mining Company Limited
06-Aug	1957	Western Deep Levels Limited
16-Jun	1964	Kloof Gold Mining Company Limited
03-May	1968	*East Driefontein Gold Mining Company Limited
12-Mar	1974	Deelkraal Gold Mining Company Limited
27-Jun	1974	Elandsrand Gold Mining Company Limited
* These companies founded a merger, Driefontein Consolidated Limited, on 1 July 1981		

Over the course of the 19th century, the farming community was transformed into a mining community with many towns along the Witwatersrand established to provide housing for the growing mining workforce. The largest of these cities was the city of Johannesburg established in 1886 with the discovery of the main reef and other towns such as Roodepoort, Westonaria and Carletonville established as additional sections of the gold reef was discovered (De Klerk, 2019).

2.3 Geology and Hydrogeology of the Far West Rand

The gold-bearing reefs of the FWR are covered by thick karstified dolomites. These dolomites host an abundance of groundwater resources which are some of the largest in South Africa and include a range of high yielding karst springs (Schrader & Winde, 2014). A basic understanding of the geological units of the study area is important when the parameters are established for use in the water budget model. The geology governs which type of aquifers dominate the area as well as the overlaying stratigraphy that effects recharge into these aquifers. Mineral deposits, especially gold deposits within the study area, delineate mine activities along the FWR (Figure 3) and impact the movement of groundwater as well as infiltration thereof. Together with the characteristics of the stratigraphy, tectonic deformation as well as dyke intrusions provide mechanisms for groundwater movement. Furthermore, mine activity in this region has led to aquifer dewatering and the formation of sinkholes that further impact the aquifer parameters for the study area. A detailed Geological and Hydrogeological discussion follows in Chapter 4.

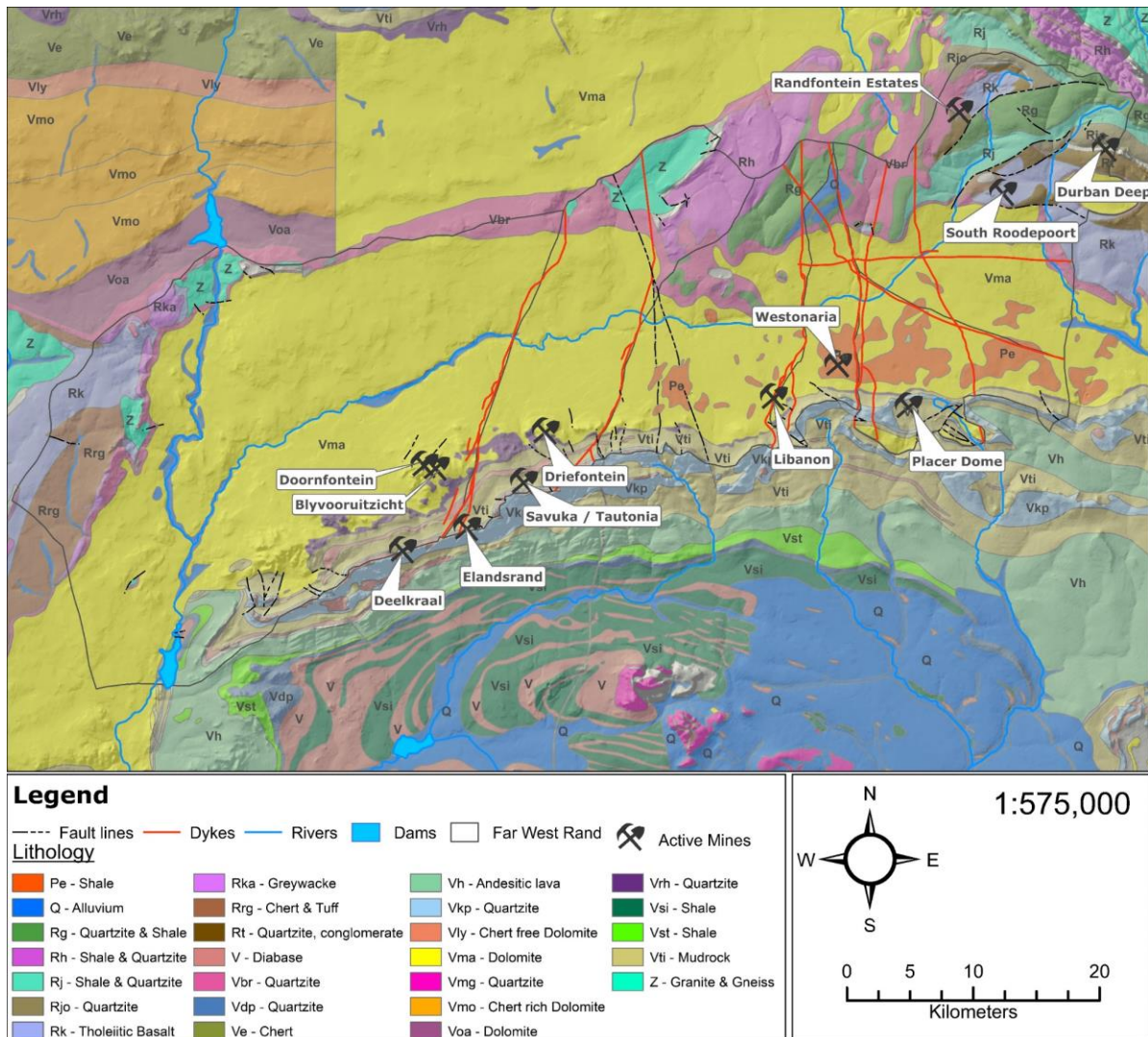


Figure 3: Mining activities along the study area.

2.4 Soil Conditions

The dolomitic lithology hosts an abundance of red to yellow weak-structured soils within the eastern bounds of the study area and minimal developed soils to the western side, where dolomitic rock outcrops are more prominent. The majority of these soils found above the dolomitic lithology have lower infiltration rate properties compared to those found in the surrounding non-dolomitic area. The soil's permeability and infiltration rate determine the soil's runoff response after a rainfall event. These properties have been characterised into four soil groups based on their overall runoff response as presented in Table 2 (De Klerk, 2019; Schulze, 2012). As presented in Figure 4 the FWR is mostly dominated by soil group C, indicating a moderately high runoff potential where soil infiltration rates are slow and soil permeability restricted with possible shallow soil depths.

Furthermore, soil group B can be seen in the eastern parts of the study area within the regions where the dolomitic rocks are absent, characterised by slightly restricted soil permeability and moderate infiltration.

Table 2: Soil runoff potential classes (Schulze, 2012).

Soil Group A	Soil Group B	Soil Group C	Soil Group D
High infiltration rate	Moderate infiltration rate	Slow infiltration rate	Very slow infiltration rate
High soil permeability	Slightly restricted permeability	Restricted permeability	Severely restricted permeability
Low runoff potential	Moderately low runoff potential	Moderately high runoff potential	High runoff potential

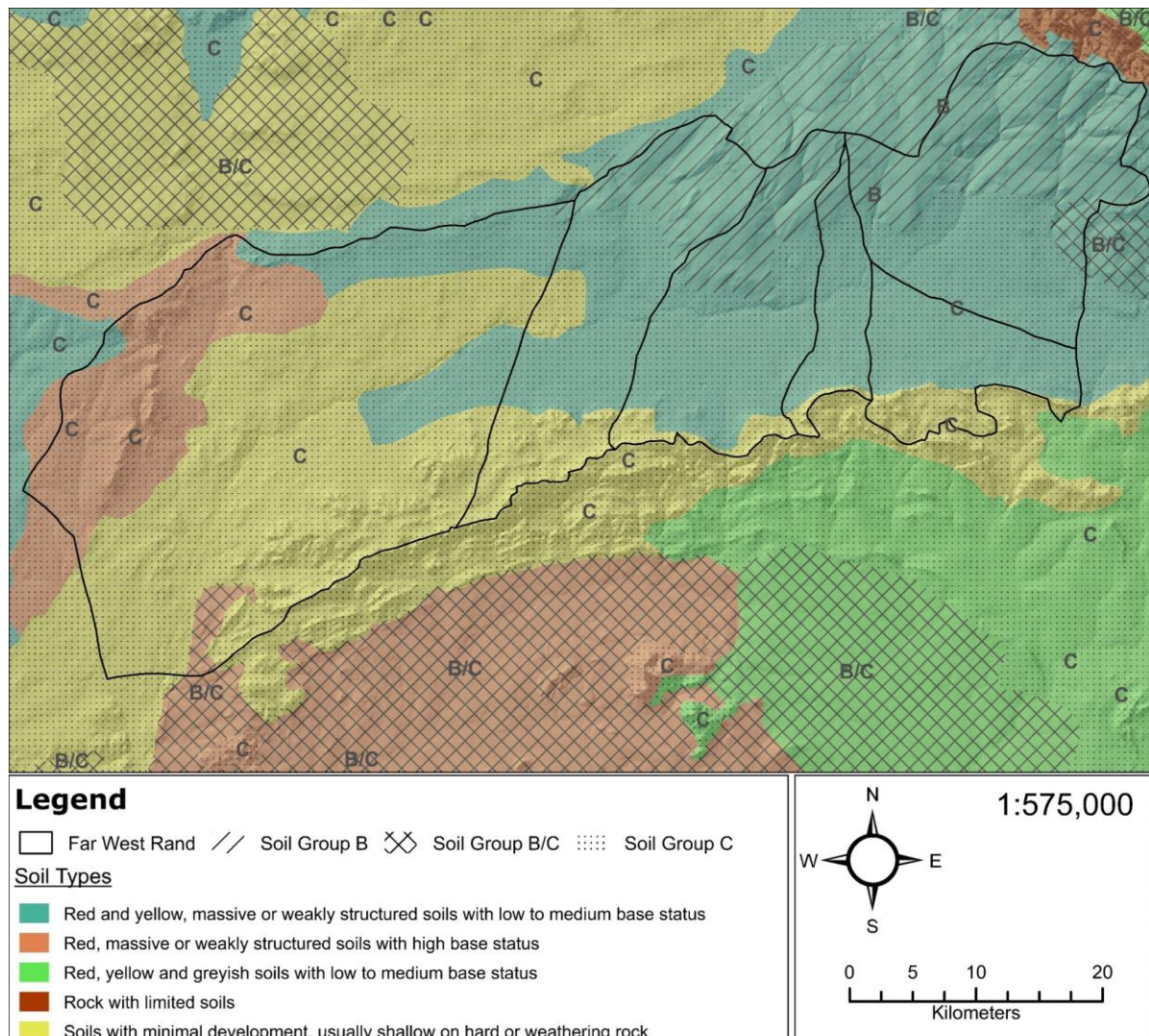


Figure 4: Soil characteristics of the study area.

2.5 Vegetation Cover

The FWR is dominated by grasslands, consisting of majority Carletonville Dolomite Grassland as well as the Soweto Grassland with parts Vaal-Vet Grassland, Tsakane Grassland and Rand Grassland. The vast grasslands spread across the study area with little woodland cover supports the presence of a rocky outcrop landscape, with shallow, weakly structured soil profiles. The southern mountainous region consists of the Gauteng Shale Mountain Bushveld, with several wetlands found across river paths Figure 5.

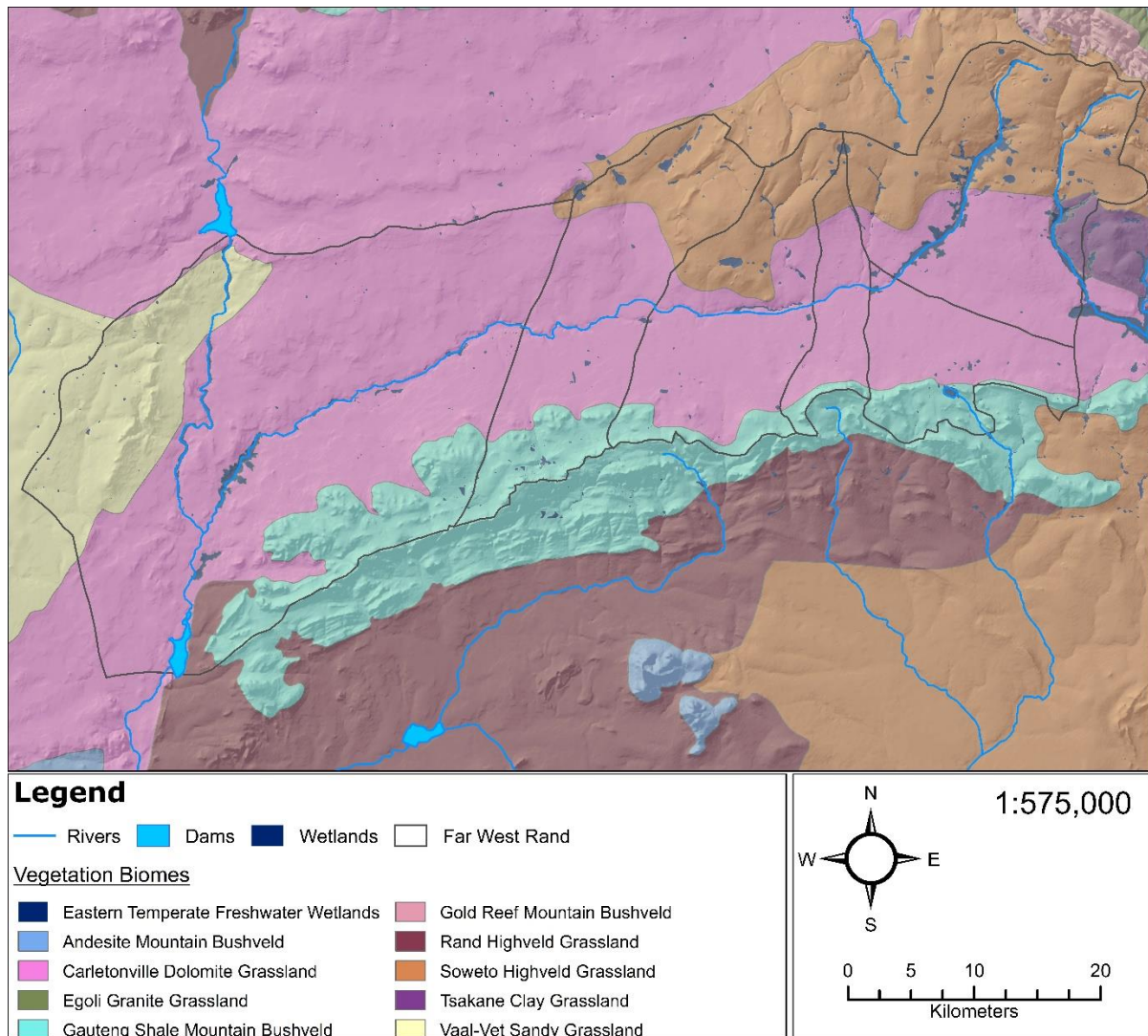


Figure 5: Regional vegetation cover.

2.6 Land Cover

Characterised by the developments brought on by the mining activities mentioned in Section 2.2, the FWR is mainly rural with most of the population and towns concentrated near to mining activities, as shown in Figure 6.

Mining activities have dominated as the main economical commodity over the past century. Some agricultural activities can still be seen across the study area as well as the natural landcover characterised by vast grasslands used for grazing (Table 3). The agricultural land use across the area not only provides the community with their basic produce but also serves as one of the highest employment sectors, second only to the mining sector (Swart, *et al.*, 2003b).

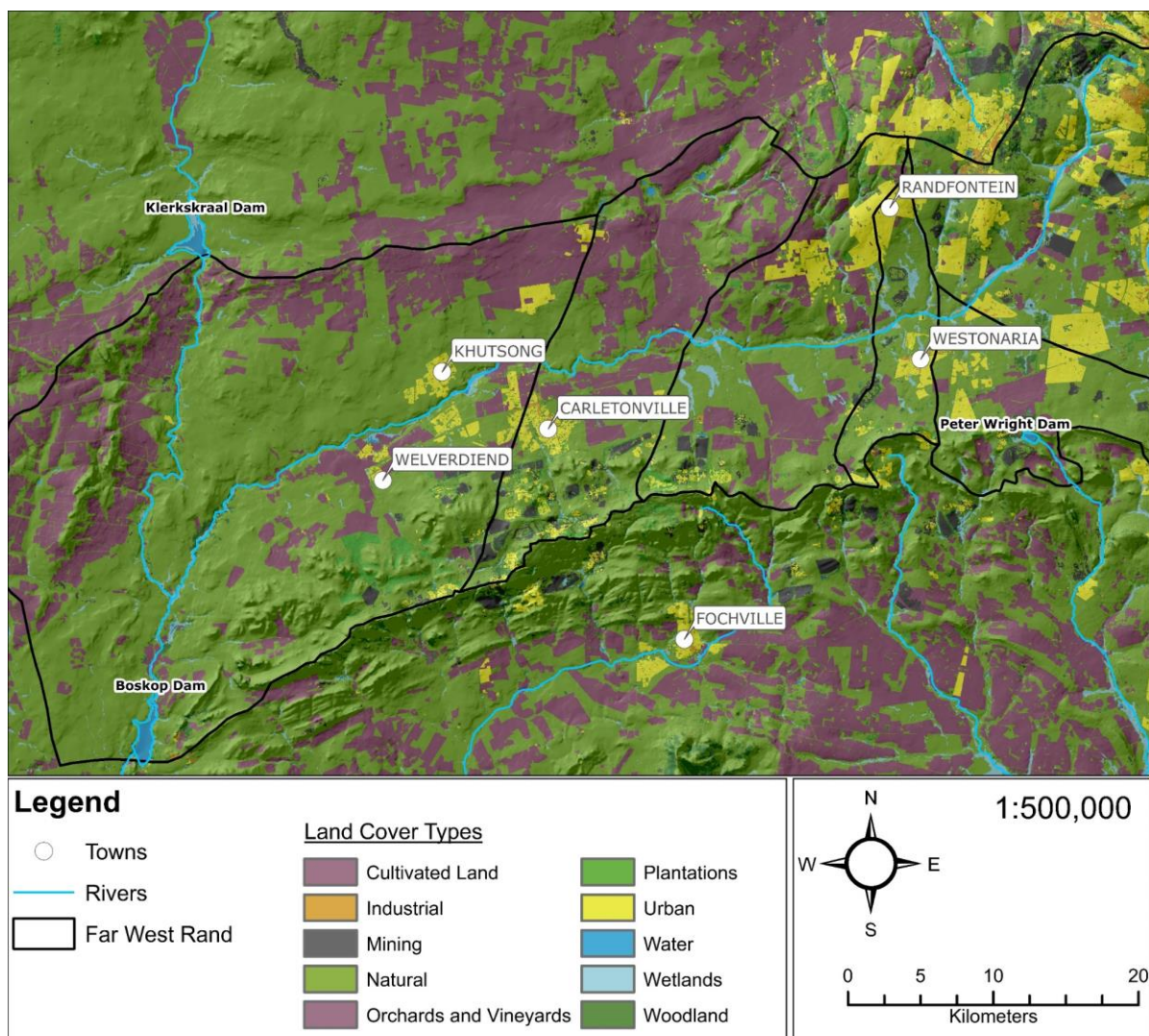


Figure 6: Land Cover for the study area (© GEOTERRAIMAGE 2015).

Table 3: 2013-2014 Land Cover statistics for the study area.

Land Cover Class	% of Total
Natural	58.6
Cultivated Land	21.7
Urban	9.3
Wetlands	2.9
Woodland	2.5
Mining	2.4
Plantations	1.6
Industrial	0.6
Water	0.3
Orchards and Vineyards	0.1

2.7 Climate of the Far West Rand

The climate of the FWR is characteristic of the Southern African Highveld. Hot summers extend from September until late April followed by warm winter days and cold evenings from June until late August (Van Biljon, 2017). Precipitation varies across the FWR with the headwaters of the Wonderfontein spruit in the east receiving more rainfall than the west. Figure 7 showcases the distribution of rainfall across the study area based on the metrological data obtained from the South African Weather Service based on various rainfall stations across the area (WR2012, 2012).

The majority of rainfall occurs during the summer months of October until March with little during the winter months as seen in Figure 8 (DWA, 2019). The semi-arid characteristics of the FWR are maintained by an evaporation rate that has exceeded the precipitation throughout the years, as shown in Figure 9. The precipitation data together with the evaporation data obtained from the South African Weather Service play a key role in determining the recharge input into the water budget model.

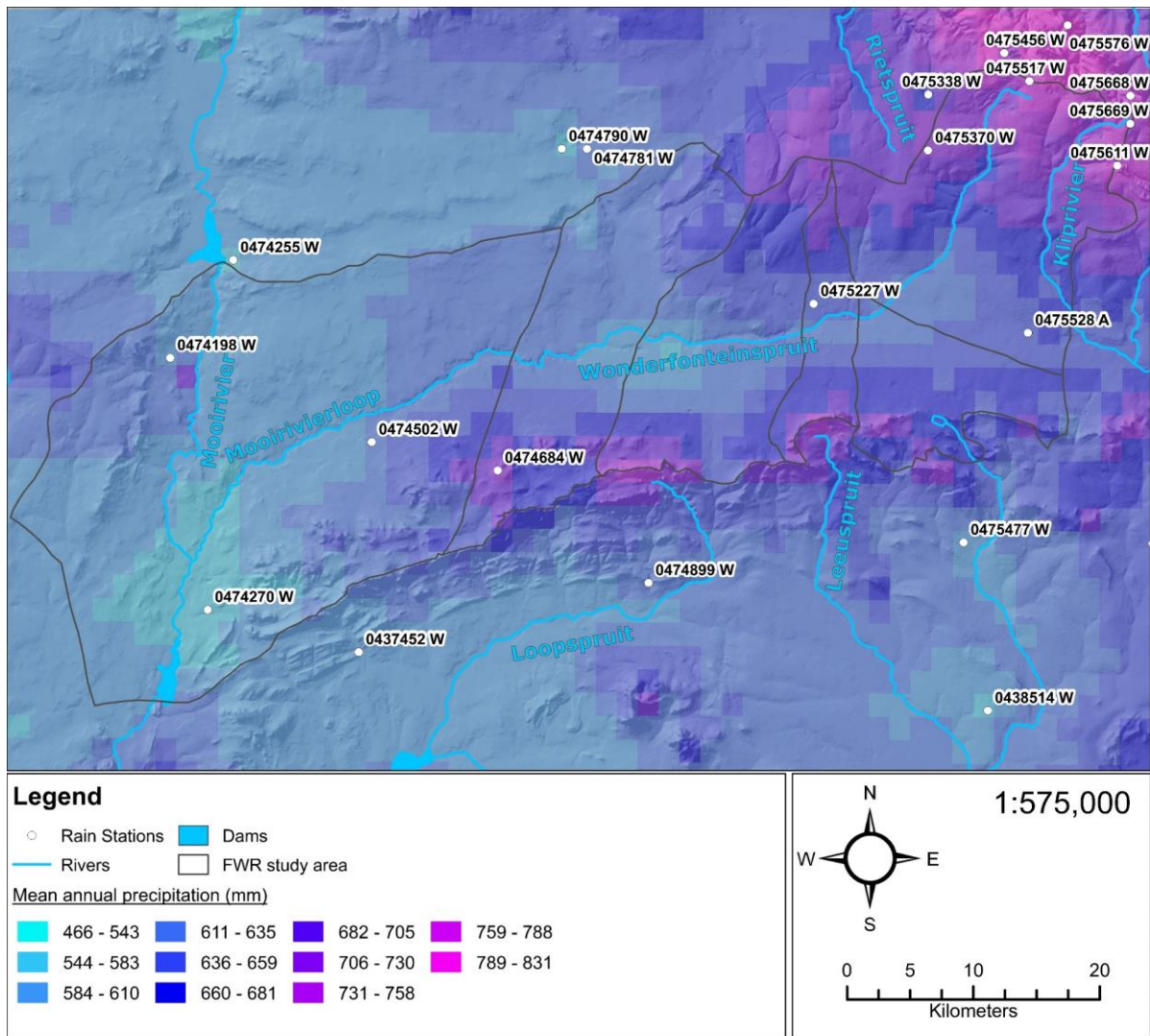


Figure 7: Rainfall map of the study area (WR2012, 2012).

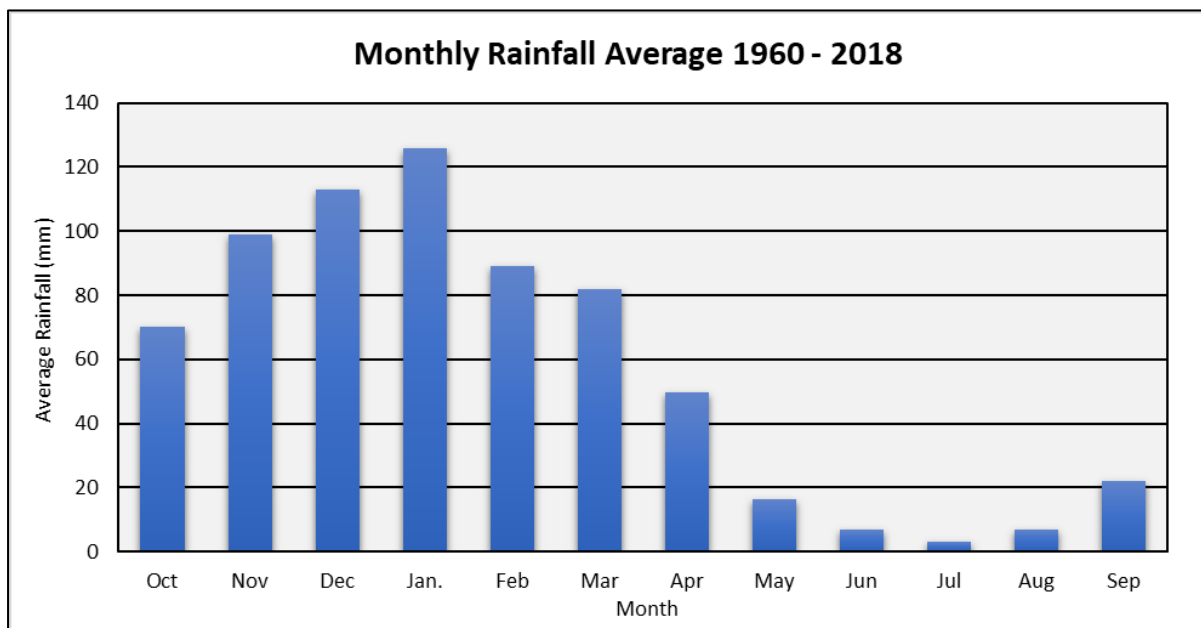


Figure 8: Monthly rainfall average for rainfall station C2E007 from 1960 - 2018 (DWA, 2019).

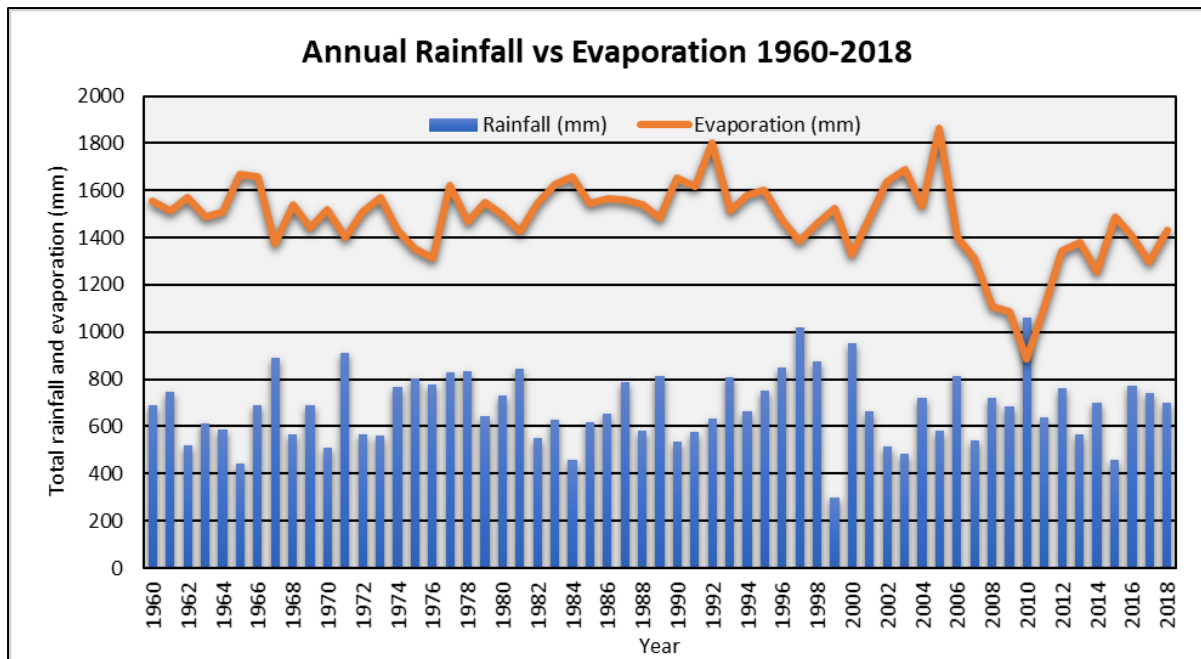


Figure 9: Annual rainfall and evaporation for rainfall station C2E007 from 1960 to 2018 (DWA, 2019).

2.8 Conclusion

In conclusion, the FWR is dominated by chert rich dolomitic strata, karstified over millions of years and characterised by karst features. The discovery of gold within the region has led to rapid urban developments within the last century and further altered the hydrogeological state of the associated aquifer system. This is mainly due to dewatering activities that enabled safe mining conditions within the shafts below this dolomitic aquifer system.

In the next Chapter, key terminology such as karst and associated karst features will be discussed in detail as well as how these karst features together with the soil and vegetation cover influences recharge within the study area.

CHAPTER 3:

LITERATURE OVERVIEW

3.1 Introduction

Modelling karst aquifer systems have a great deal of uncertainty. Contributing factors such as the overall recharge and storage of a system are affected by aspects such as the degree to which the dolomites have been weathered. To successfully model the hydrogeological conditions of the FWR an understanding of the basic mechanisms involved is required. This includes how these mechanisms are influenced by dewatering as well as how they influence each other (Enslin, 1967; Schrader, *et al.*, 2014b).

This chapter discusses the basic principles of karst landforms, karst aquifers and their commonly observed karst springs. In understanding how these systems function, further investigation will determine how the largest mechanisms involved can be estimated accurately as well as how a water budget model can demonstrate its usefulness by simulating these interactions in a bid to understand how a water budget model can be implemented for modelling possible recovery scenarios. In addition, a discussion as to how water budget model calibration can be assessed will conclude the chapter.

3.2 Karst landforms

“Karst” can most simply be defined as “a special style of landscape containing caves and extensive underground water systems that is developed on especially soluble rocks such as limestone, marble, and gypsum” (Ford & Williams, 2007:1). Karst landforms have distinctive hydrological and hydrogeological characteristics, complicated through karstification. The hydrological karst landforms, characterised by features such as large springs, caves, enclosed depressions, sinking streams and fluted rock outcrops, play a crucial role in recharge. Distinguishing a karst system from that of a well-developed fissure or fracture systems is often very difficult as even though karst morphological features can be absent on the surface of a landscape, a well-developed subterranean karst system may be present (Ford & Williams, 2007; Jeannin, 2014).

Karst terrain includes combination between high rock solubility and well-developed secondary porosity of fractures. These karst features decrease in frequency and dimension with the karst host rock depth. Highly soluble carbonate rocks that originate from organic-based depositions are prone to post-depositional alteration such as karstification. Carbonate rocks contain more than 50% carbonate minerals with two pure mineral end-members: limestone (CaCO_3), which

ranges between calcite and aragonite and dolostone or dolomite ($\text{CaMg}[\text{CO}_3]_2$), a calcium-magnesium composition. A general consensus is that an exposed carbonate rock can be assumed karstified to an extent even if the rock itself is covered in sediment unless proven otherwise (Bonacci, 1993; Ford & Williams, 2007; Jeannin, 2014).

The presence of large springs and a general lack in surface water network contributes to the fact that karst hydrological systems are highly permeable, with in most cases all precipitation lost to infiltration. The solubility of the rock mass itself is insufficient to produce karst landforms, therefore this process is accompanied by tectonic movements and well-linked fissure and fracture systems that vary in size (Bonacci, 1993; Eslamian, *et al.*, 2014; Ford & Williams, 2007).

3.2.1 Karst aquifers

Karst aquifers arise from karstified landforms, characterised by complex hydrogeological features such as fissures, fractures, conduits and altered pore structures. The most prominent of these features, conduits, have low-resistance pathways for groundwater flow, fast-tracking the granular permeability of the aquifer. A karst aquifer is seen as a dual hydrological system with water flow ranging between extremely fast and slow movement within the saturated and unsaturated zones, consisting of a network of embedded conduits that interact with less karstified rock (Ford & Williams, 2007; Fu, *et al.*, 2016; Goldscheider, *et al.*, 2007; White, 2002). Therefore, the permeability of karst aquifers primarily arises from enlarged fissures, fractures and joints (White, 2002).

A high heterogeneity is typical for karst aquifers with large differences in apertures within the four zones present in these systems. These four zones are the soil cover, the epikarst or subcutaneous zone, the vadose or unsaturated zone and the phreatic or saturated zone. Each differs in vertical extent, playing an important role in groundwater transport (Fu, *et al.*, 2016; Jukić & Denić-Jukić, 2015).

The soil cover and epikarst zones play important roles in overland flow generation and underground lateral flow that precede vertical flow to the vadose zone. The epikarst is characterized by an increased degree of dissolution, weathering and fractures that results in higher porosity and permeability (Jukić & Denić-Jukić, 2015; Poulain, *et al.*, 2015). Groundwater movement within the vadose zone can be divided into three components (Jukić & Denić-Jukić, 2015):

- Shaft flow; small volumes of water that flow along the walls of natural conduits.
- Vadose flow; predominant vertical flow down enlarged fractures and joints.

- Vadose seepage; slow movement through the smallest karst fissures and fractures.

These three components drive the phreatic zone input, where the flow is predominated by horizontal circulation in a network of channels that feed karst springs (Atkinson, 1977; Jukić & Denić-Jukić, 2015).

Bonacci (1993) explained that although the water retention capacity within these karst systems is poor, with previous analysis conducted by Bonacci (1987) indicating effective porosity figures averaging between 0.1% and 1%. Caves and transport conduits changes this characteristic, giving the karst apparent high storage capacities. On the contrary, sudden raising and lowering of the groundwater levels indicate poor storage capacity. This is mainly due to conduits contributing only a small percentage of the aquifer volume with the larger matrix volume of low porosity providing a larger storage contributing factor (Bonacci, 1993; Goldscheider, *et al.*, 2007). Therefore, karst systems are dominated by two types of flow: Diffuse flow, slow laminar flowing water through small karst features; and conduit flow, fast turbulent flow through large fissures, fractures and conduits larger than 10mm in diameter (Atkinson, 1977; Bonacci, 1993).

Figure 10 indicates the effects conduits have on the flow within karst aquifers. The hydraulic head indicates a rapid response to flow within the conduit flow path with little storage in the matrix, whereas flow within the matrix results in a dulled response. Flow within an area dominated by fractures would result in a combination of the two (Kresic, 2007).

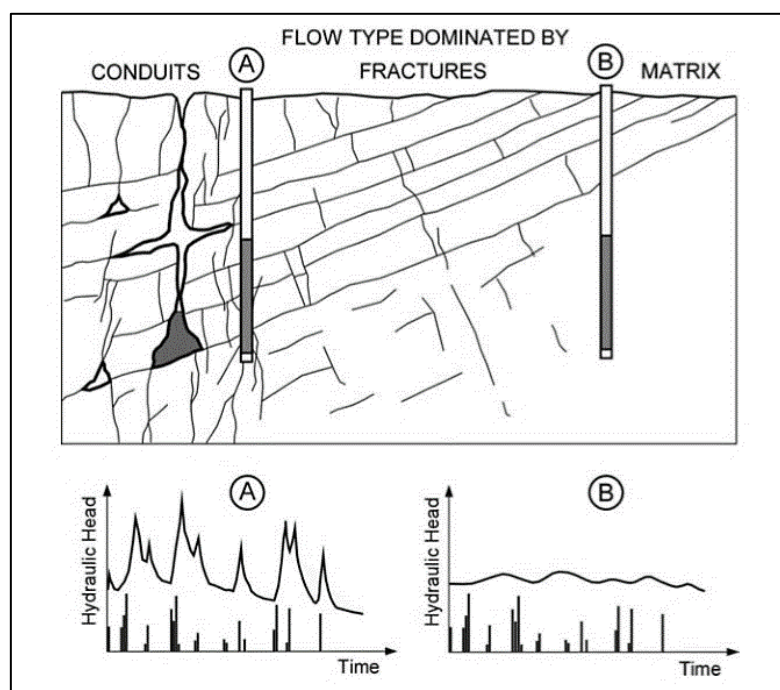


Figure 10: Hydraulic head response to different flow types within karst aquifers (Kresic, 2007).

3.2.2 Karst springs

A common occurrence within karst systems involves surface water that becomes groundwater when it sinks into streambeds or into karst features such as sinkholes. Whereas groundwater becomes surface water when it emerges from springs. These springs can yield high discharge volumes forming the headwaters of large surface streams or travertine dams in some cases (Ford & Williams, 2007; White, 2002).

Springs are usually classified though their hydrological function that mainly depends on the topographic and structural aspects and how they influence the aquifer (Bögli, 1980; Ford & Williams, 2007):

- Free draining springs, as seen in Figure 11(a) and (b). These springs are gravity-driven, where a karst rock slopes in the direction of the topographically lower adjacent valley. These systems are dominated by a large vadose zone with the development of phreatic zones in cases where a folded impermeable base rock causes subterranean ponding.
- Dammed springs, as seen in Figure 11(c) to (e). With these springs, temporary overflow is produced by high water tables caused by impoundment and obstruction of the groundwater flow path. This is mainly due to semi-/impermeable barriers including other lithologies, faulted or conformable contacts, valley aggradation by glaciofluvial deposits or saltwater barriers near the ocean. Dammed springs are the most common type of karst springs and include the springs situated within the FWR, each formed by a raised water table due to impoundment through impermeable dyke intrusions.
- Confined springs, as seen in Figure 11 (f) and (g). These springs are characterised by up-domed turbulent 'boil' prevail under hydrostatical pressure. These artesian conditions are brought on by impermeable rock layers that confine the karst rock, causing the water to discharge through fault planes or even areas where the caprock has eroded.

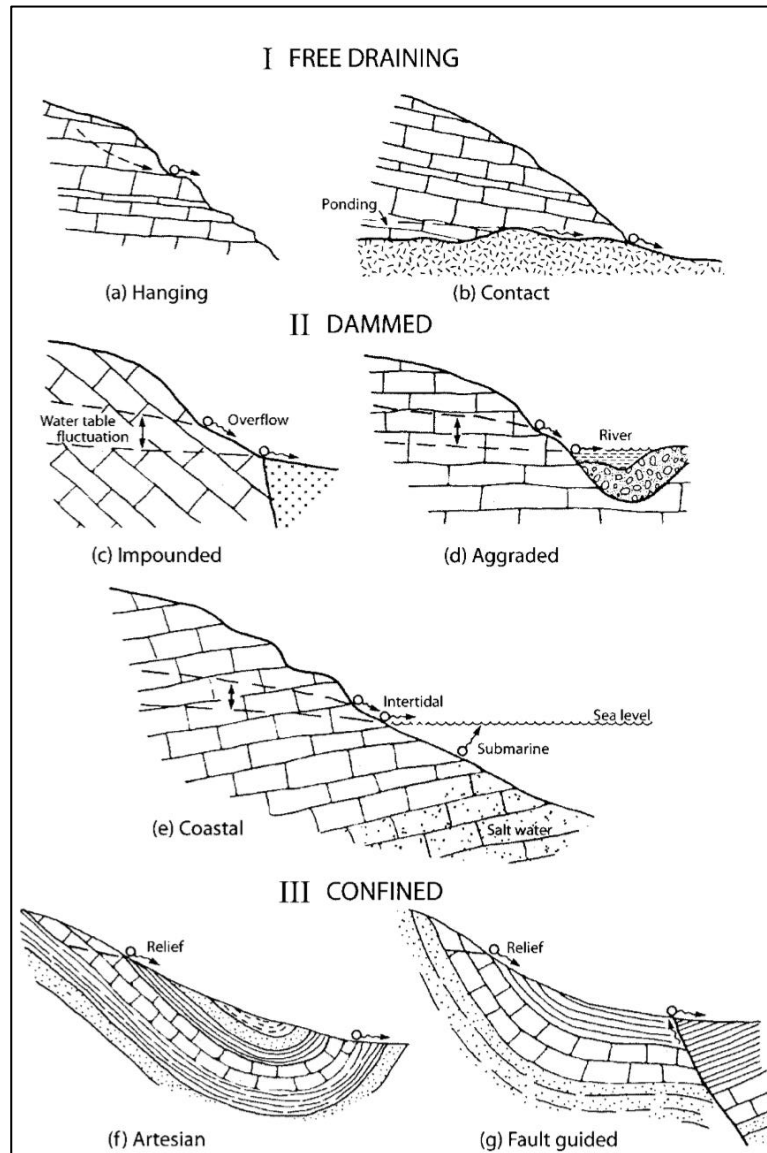


Figure 11: Karst springs classified through topographic and structural aspects (Ford & Williams, 2007).

“Most of the largest springs in the world are karst springs.”, a notion by Ford & Williams (2007:118). Springs control the elevation of their associated aquifers’ water table due to their vertical positions with the hydraulic conductance properties determining the drawdown slope associated with the output of upstream aquifer water levels (Ford & Williams, 2007).

Within the vadose zone, conduit networks integrated from recharge points extend like plumbing in the subsurface, draining the karst effectively through the systems associated springs (Ford & Williams, 2007). Deeper circulation within the phreatic zone is driven by the elevation difference between the spring and the upstream water table, with a larger head difference supplying more energy as depicted by Darcy’s Law (Ford & Williams, 2007).

Springs, being discharge conduits in layman's terms, can temporarily become inflow conduits during severe flooding events. Usually, these springs discharge at or beneath a major river, becoming deeply submerged during a flood event will cause the hydraulic head within the karst to reverse (Ford & Williams, 2007). The effects of these temporary sink forming 'estavelles' should be noted when addressing the recharge within the FWR as the dried-up springs may serve as inflow conduits.

Therefore, when addressing potential spring yields, it is important to define the contributing karst groundwater system and to consider that karst aquifers are often drained by several springs from overlapping catchments. The output of these springs is primarily influenced by recharge as well as different groundwater pathways (Drew & Goldscheider, 2007; Ford & Williams, 2007).

3.3 Groundwater recharge in karst systems

Recharge as a percentage of annual precipitation is more complex within karst aquifer systems than in other, inter-granular aquifers. The rate of recharge or effective infiltration is largely influenced by the variability in the porosity and permeability of these carbonate rocks. This variability determines whether the infiltration is diffuse and/or concentrated; how it moves through the unsaturated zone as matrix and/or fracture and conduit network flow; as well as whether the discharge is diffuse and/or concentrated. Aquifer recharge is either expressed as annual mean volume or percentage of precipitation. (Andreo, *et al.*, 2008; Bakalowicz, 2005; Guardiola-Albert, *et al.*, 2015b; Lerner, *et al.*, 1990; Martos-Rosillo, *et al.*, 2015).

Being spatially uniform, the recharge is distributed through a number of pores and fissures across the karstic outcrop. Areas of thick, less permeable soil cover regulate recharge whilst limiting recharge due to the infiltration capacity of the soil (Ford & Williams, 2007). Concentrated infiltration takes place by means of sinkholes and other conduit networks, allowing rapid surface runoff drainage as well as rapid transmission through the vadose zone to the saturated zone. Diffuse infiltration, being a slower form of drainage can take up to weeks or even months as percolation depends on the characteristics such as the thickness and hydraulic conductivity of the vadose zone (Fiorillo & Pagnozzi, 2015).

Recharge takes place in two ways (Figure 12), the first involves a simple context where the catchment only comprises of karst rocks with precipitation falling directly on them. This is known as autogenic recharge. The second type involves a more complex geological setting where runoff occurs from neighbouring or overlying non-karstic rock, draining onto the karst aquifer (Ford & Williams, 2007).

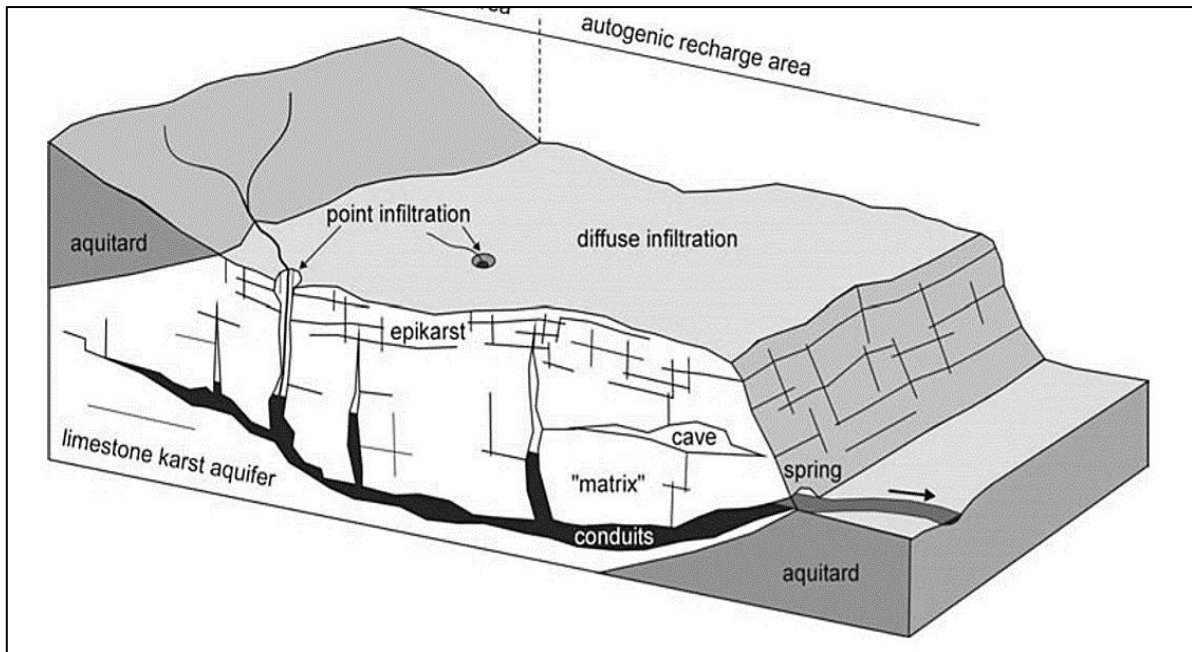


Figure 12: Recharge endmembers found in karstic systems (Goldscheider *et al.*, 2007).

Autogenic recharge is mainly diffuse and slow, infiltrating down fissures within the rock or even through overlying soils. Concentrated rapid infiltration does take place during point infiltration into shallow holes or dolines. Allogenic recharge occurs as concentrated point infiltration through sinking streams from adjacent non-karstic terrain (Ford & Williams, 2007; Goldscheider, *et al.*, 2007).

Autogenic recharge leads to the enhancement of pore and fissure porosity with little impact on conduit formation, whereas allogenic point recharge causes the formation and growth of these large stream passages (conduits). Not only does the recharge volume per unit area differ between the two types of recharge entities but it also affects the water chemistry causing a large distribution of the development of secondary permeability. Karstic systems are characterised by this duality effects of recharge, with other aquifer types singularity effected by autogenic recharge (Ford & Williams, 2007; Goldscheider, *et al.*, 2007).

The amount of precipitation that falls onto the surface, recharging the groundwater system depends not only on the lithological aspects discussed but also on vegetation cover as well as rainfall intensity and impermeable artificial landforms. Rainfall intercepted by vegetation subsequently evaporates from leaves and branches with some transpired by the plant themselves. Intense rainfall events can lead to inadequate infiltration due to the infiltration capacity being lower than the amount of water accumulating on the surface leading to increased runoff and evaporation, similar to the occurrence of impermeable artificial landforms

(Ford & Williams, 2007). To address the effects of evaporation Ford & Williams (2007) make use of Equation (3.1):

$$Q = (P - E)_{\text{autogenic}} + (P - E)_{\text{allogenic}} \pm \Delta S \quad (3.1)$$

Where discharge Q at the outflow of the basin depends on the different precipitation P and the evaporation E of both autogenic and allogenic runoff as well as the change in storage ΔS . These factors mentioned above not only influence the direct recharge of an area, the water added to the groundwater reservoir as precipitation but also indirect recharge, percolation to the water table through surface water bodies (DWAF, 2006). Furthermore, the recharge of the aquifer is directly impacted through groundwater pumping (Martos-Rosillo, *et al.*, 2013). These aspects are particularly important when calculating recharge in semi-arid South Africa as the evaporation rate far exceeds the precipitation. Figure 13 conceptualises all flows that take part within karst drainage systems and gives a basic idea of what could be expected within the karst system of the FWR.

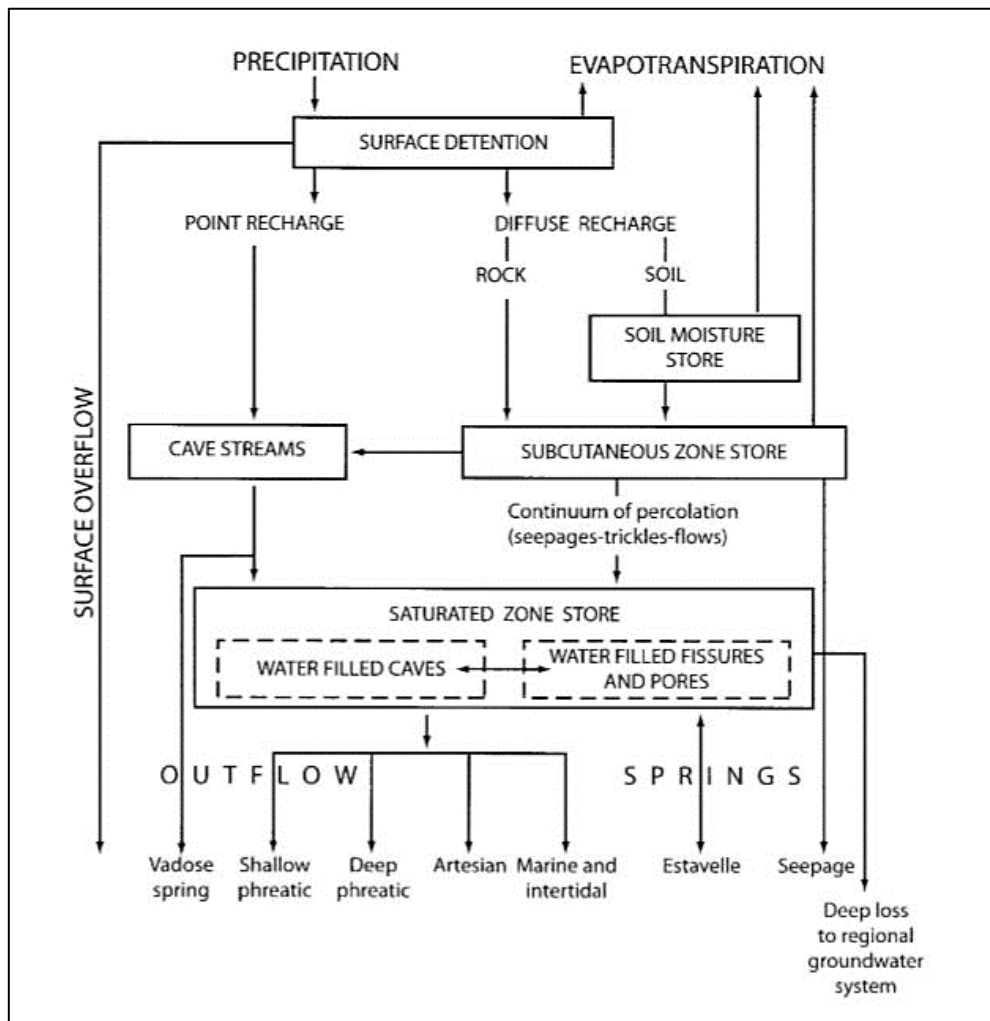


Figure 13: Karst drainage flow system (Ford & Williams, 2007).

In general, recharge in carbonate aquifers is estimated by means of methods adapted and derived from those used for detrital aquifers. However, carbonate aquifers have specific characteristics (forms of preferential infiltration including point recharge via swallow hole, flow concentration in the epikarst, thin soil, thicker unsaturated zone) and a hydrogeologic functioning that is typical of heterogeneous materials.

In most cases, the estimation methods for recharge within carbonate aquifer systems have been adapted and derived from those used in detrital aquifers. These methods are from a diverse collection, including direct determination through the use of lysimeters and seepage meters; Darcian (Darcy's law) approaches using field measurements and numerical methods; water budget methods, using empirical equations and chemical investigations including the chloride mass balance and isotope measurements (Andreo, *et al.*, 2008; Samper, 1997; Scanlon, *et al.*, 2002). Therefore, it's essential to compare results from different methods to minimize the degree of uncertainty (Martos-Rosillo, *et al.*, 2013).

Estimating groundwater recharge within the semi-arid areas including the FWR poses many challenges of uncertainty including the time variability of precipitation, the spatial variability of soil characteristics, topography, vegetation and land use (Beekman & Xu, 2003; Lerner *et al.*, 1990). Although there are many methods available for calculating recharge estimates, there are limitations surrounding the applicability and reliability of these methods (Beekman & Xu, 2003).

3.3.1 APLIS-recharge estimation

Due to countless limitations and a lack of access to accurate data as well as the availability of accurate data in general, conventional methods (natural, evapotranspiration, time-series precipitation calculation, chemical or synthetic isotope tracing) and numerical modelling methods for calculating recharge are problematic (Alem, *et al.*, 2017; Guardiola-Albert, *et al.*, 2015a; Radulovic, *et al.*, 2012).

With significant advances in GIS software in recent times, methods for calculating groundwater recharge has become easier, quicker and more concise. These GIS-based methods utilise parameter attributes such as altitude, geology, soil type, vegetation and precipitation for spatial-temporal recharge distributions (Alem, *et al.*, 2017; Andreo, *et al.*, 2008).

Andreo *et al.* (2008:913) had a fundamental aim to develop a method to:

- “*Estimate the rate of autogenic recharge into carbonate aquifers on the basis of a combination of geological, geographic, morphologic and ethnological variables.*”
- “*Determine the spatial distribution of this recharge into the aquifers.*”

The end result came in the form of an algorithm to calculate the recharge rate as an expression of the percentage of precipitation. The GIS-based method, APLIS (Altitude, Pendant or Slope, Lithology, Infiltration, Soil) not only allows for the mapping of the spatial distribution of spontaneous recharge, but also allows for annual groundwater recharge approximation in karst formations by accounting for direct as well as indirect recharge sources. Furthermore, APLIS does not require accurate nor periodic data, therefore, reducing water resource management costs (Alem, *et al.*, 2017; Andreo, *et al.*, 2008; Espinoza, *et al.*, 2015; Martos-Rosillo, *et al.*, 2015).

APLIS is a mathematical statement that superpositions all the main variables relating to infiltration. These variables do not have the same importance in recharge, therefore differ in weighted values according to the ranking system of APLIS. The ranking system presented in Table 4 has been derived from various multidisciplinary approaches including regression analysis, ideal spot analysis and linear weighted sum measurements (Alem, *et al.*, 2017; Andreo, *et al.*, 2008).

The mean recharge rate of the aquifer is calculated by conducting a weighted overlay in GIS, where each variable is accounted for in Equation (3.2) (Andreo, *et al.*, 2008):

$$R = [(A + P + 3L + 2I + S)/0.9] \cdot Fh \quad (3.2)$$

With R representing the mean recharge rate, A : altitude, P : Pendant or Slope, L : Lithology, I : Preferential infiltration and S as Soil type. As seen in Equation (3.2), the lithology variable is three times more important than that of altitude represented in meters above mean sea level (mamsl), slope and soil type with preferential infiltration twice as important. Because the variables are divided by a correction factor of 0.9, recharge values fall between 8.88% and 88.8%. This correction factor varies between 0.1 and 1 according to the permeability of the aquifer. Therefore, precipitation will never result in 100% recharge (Alem, *et al.*, 2017; Andreo, *et al.*, 2008).

Table 4: APLIS variable ranking system (Alem, et al., 2017).

A		P		L	
Altitude (m)	Rating	Slope (%)	Rating	Lithology	Rating
(<300)	10	(<3)	10	Limestones and dolostones	10 or 9
(300-600)	9	(3-5)	9	(karstified)	
(600-900)	8	(5-10)	8	Limestones and dolostones	8 or 7
(900-1200)	7	(10-15)	7	(fractured, slightly karstified)	
(1200-1500)	6	(15-20)	6	Limestones and dolostones	6 or 5
(1500-1800)	5	(20-30)	5	(fissured)	
(1800-2100)	4	(30-45)	4	Gravels and sands	4
(2100-2400)	3	(45-65)	3	Conglomerates	3
(2400-2700)	2	(65-100)	2	Plutonic and metamorphic rocks	2
(>2700)	1	(>100)	1	Shales, silts and clays	1
I		S			
Infiltration landforms	Rating	Soils		Rating	
Abundant infiltration landforms	10	Leptosols		10	
Moderate infiltration landforms	5	Arenosols and xerosols		9	
Scarce infiltration landforms	1	Calcareous regosols and fluvisols		8	
Fh		Euthricregosols and solonchaks		7	
		Cambisols		6	
Correction factor		Euthriccambisols		5	
Aquifer	1	Histosols and luvisols		4	
Non-aquifer	0.1	Chromicluvisols		3	
		Planosols		2	
		Vertisols		1	

Andreo *et al.* (2008) implemented and calibrated the APLIS method according to recharge values obtained by physical recharge estimation methods for eight aquifers in Southern Spain. The recharge values obtained by APLIS differed up to 5% from those obtained by physical methods, with APLIS underestimating recharge values.

APLIS was also implemented in the studies of Alem *et al.* (2017) and Martos-Rosillo *et al.* (2013) resulting in estimations close to that of physical measurements, prompting minimum recharge in low laying non-karstic zones and the highest recharge rates in minimal sloped karstic terrain. Zaree *et al.* (2019) implemented the APLIS method to estimate autogenic groundwater recharge, from this study it was discovered that the APLIS method has to be calibrated against another method/s due to rainfall shortages.

3.4 The water budget approach

Based on the law of the conservation of mass, the water budget principle accounts for all groundwater entering and leaving an unexploited aquifer during steady-state conditions over a period of time, Wright & Xu (2000) Equation (3.3):

$$\text{RECHARGE} - \text{DISCHARGE} = 0 \quad (3.3)$$

Recharge is represented by the water infiltrating into the subsurface from the host rock as well as runoff from adjacent geohydrological units, with effective recharge describing the water that reaches the groundwater table (Bredenkamp, *et al.*, 1995).

Discharge includes losses such as contribution to surface watercourses, subsurface inter-aquifer flows and direct evaporation and evapotranspiration losses. The exploitation of groundwater resources by man reduces the overall discharge of the aquifer together with the modification of recharge and changing of the aquifer storage, altering the water budget, Equation (3.4) (Wright & Xu, 2000).

$$(\text{ADJUSTED RECHARGE}) - (\text{REDUCED DISCHARGE}) - \text{PUMPING} + \Delta \text{STORAGE} = 0 \quad (3.4)$$

A sub-sequential lowered groundwater table can cause an increase in recharge as well as a decrease in discharge. The decreasing discharge is more profound than an increasing recharge, consequentially leading to a groundwater yield governed by the discharge component. Therefore, any unsustainable decline or misuse of the groundwater resource over a long period of time is visible as deteriorating groundwater levels, piezometric surface or groundwater quality. This approach can successfully be applied to estimate groundwater recharge (Wright & Xu, 2000).

3.4.1 The ACRU– agrohydrological model

The groundwater recharge map developed by Vegter (1995) was based on the agrohydrological model – ACRU originally developed for distributed catchment evapotranspiration studies within the Natal Drakensberg mountains in the early 1970s. The model was primarily developed as a teaching tool and used as an algorithm for the agrohydrological and argoclimatological atlas for Natal. Since then it has been continuously improved by several researchers and graduate students from the Department of Agricultural Engineering of the University of Natal in Pietermaritzburg for broader applicability (CWRR, 2019).

The model is made up of several components that can be multiplied within a basic tree diagram, enabling flexibility and configuration of the components. Figure 14 showcases the basic flow diagram used to implement the ACRU model.

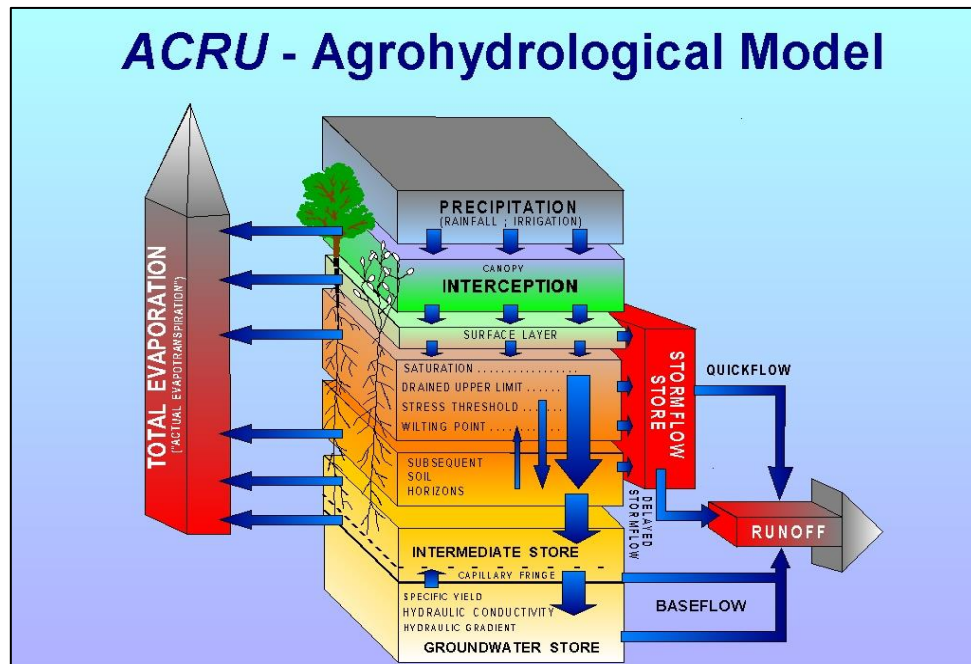


Figure 14: ACRU flow diagram (CWRR, 2019).

The complexity of direct and indirect recharge was addressed through the following factors (DWAF, 2006):

- The type, intensity, duration, amount and temporal distribution of rainfall and evaporation.
- Surface aspects including the slope, vegetation cover, stormwater runoff, interception and transpiration losses.
- The infiltration capacity of surface materials including the presence of macro-pores and fractures.
- The unsaturated zone's moisture retention capacity.
- The river bedrock deposit type as well as size.

Vegter (1995) ran this model for different homogeneous rainfall response zones for South Africa with point recharge results obtained from a number of studies. Figure 15 showcases the national recharge map developed by Vegter (1995) based on contoured mean values (DWAF, 2006).

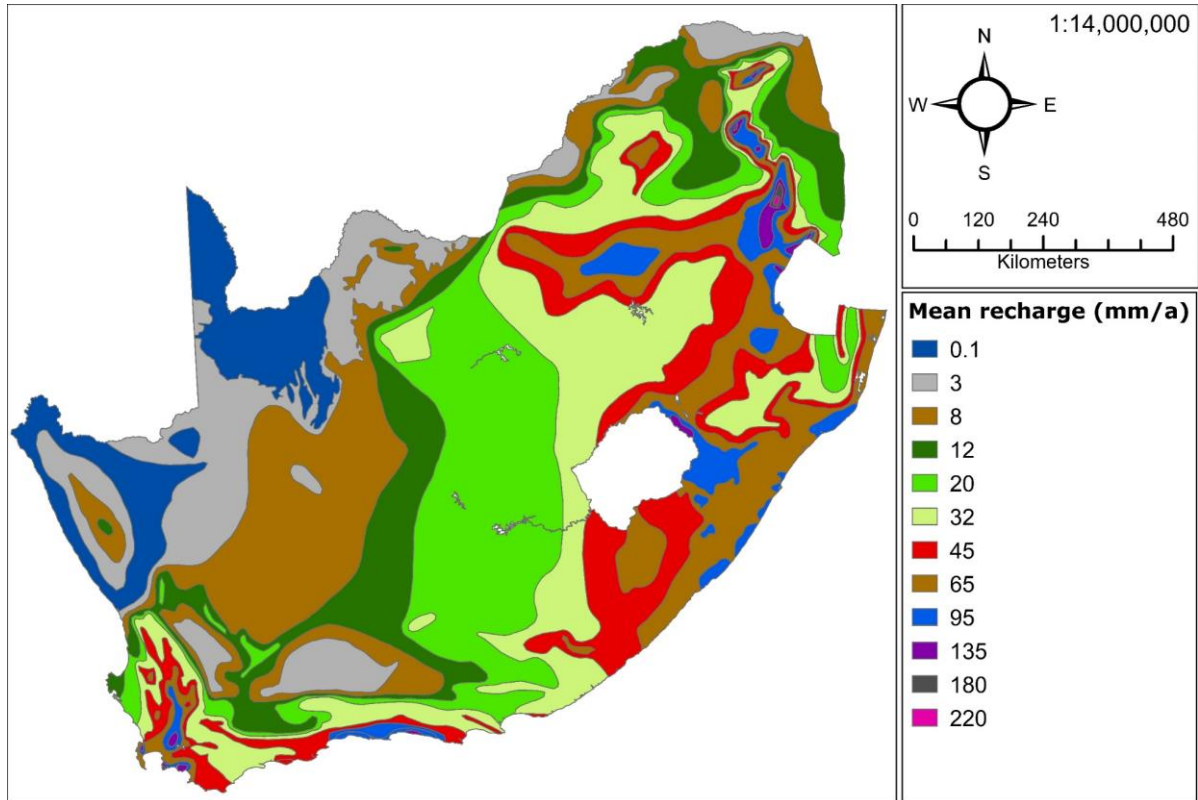


Figure 15: National recharge map developed by Vegter (1995).

3.4.2 The CMB method

The Chloride Mass Balance (CMB) assumes the conservation of mass between the inputs from the atmospheric chloride and the subsurface chloride flux when there is no other source of chloride other than rainfall. Not only does this method estimate the moisture flux within the unsaturated zone by means of diffuse flow profiling, it can also provide an insight into the recharge mechanisms involved (Andreo, *et al.*, 2008; Beekman & Xu, 2003).

Generally, for steady-state conditions the recharge is estimated between the chloride influx at the surface and that of the unsaturated/vadose zone, due to the constant nature of chloride and abundance in precipitation. The calculation is shown in Equation (3.5) (Beekman & Xu, 2003):

$$R = \frac{PCL_p + D}{Cl_{gw}} \quad (3.5)$$

With Precipitation as P , the dry chloride deposition ($\text{mg}/\text{m}^2/\text{a}$) as D , the chloride within the precipitation as Cl_p and the chloride in groundwater as Cl_{gw} . The groundwater chloride is expressed as the harmonic mean of multiple chloride values obtained due to fluctuations in the chloride concentrations, Equation (3.6):

$$Cl_{gw} = N \left(\sum_{i=1}^N \frac{1}{Cl_{igw}} \right)^{-1} \quad (3.6)$$

These fluctuations are due to increased chloride concentrations not only from fractured rock systems as a product of weathering but also from pollution sources. Overall the CMB method is useful in areas where the chloride concentration in the groundwater does not fluctuate greatly. Infrequent rainfall patterns across South Africa limits the use of the method throughout the year. The method also does not account for evapotranspiration losses (DWAF, 2006; Beekman & Xu, 2003; Van Tonder & Bean, 2003).

Furthermore, Werner (2014) concluded in his study to estimate recharge using the CMB method in karst aquifer conditions, that this method underestimated the recharge into the aquifer when compared to other methods. This according to Werner (2014) is due to the duality of recharge that occurs within karst systems.

3.4.3 The SVF-method

First introduced by Van Tonder (1989) as mentioned by Bredenkamp, *et al.* (1995) this method was later adapted to derive aquifer storativity. The SVF-method incorporates a lumped parameter approach where the status of the aquifer is assessed on the basis of waterlevel fluctuations within monitoring boreholes. The method analyse the water budget by lumping the abstraction and integrating the change in the aquifer storage. This is due to a saturated volume response of the aquifer over a period of time (Bredenkamp, *et al.*, 1995). Moreover, the method incorporates changes within the system brought on by imbalances between components. Variable rainfall together with delayed responses due to catchment linked features as well as drainage response differences due to soil type and geological characteristics can complicate the necessary links between the specific time orientated rainfall and the hydrological outputs (Adams, *et al.*, 2004).

The SVF-method not only determines the average annual rainfall and annual recharge variability by analysing the water level response of the aquifer but also derives the aquifer storativity from linear plot analysis of the SVF against the net discharge, Q (Adams *et al.*, 2004; Bredenkamp *et al.*, 1995).

A difference in storage is expressed as the change in groundwater level with all inflows and outflows rendered by the change in the head by using the aquifer area as well as the specific yield, Equation (3.7) (Dennis & Dennis, 2017) :

$$h_t = h_{t-1} + \frac{R_t}{S_y} + \frac{Q_{in}-Q_{out}}{AS_y} \quad (3.7)$$

Where,

Table 5: Equation (3.7) abbreviation list.

t	=	Current time step [T]	S_y	=	Specific Yield
h_t	=	Head in current time step [L]	A	=	Aquifer surface area [L ²]
h_{t-1}	=	Head in previous time step [L]	Q_{in}	=	Sum of all groundwater inflows [L ³]
R_t	=	Recharge in current time step [L]	Q_{out}	=	Sum of all groundwater outflows [L ³]

Unconfined aquifer conditions are assumed with the use of specific yield with head values expressed in meters above mean sea level (mams). The SVF method has specified inflows and outflows with inflows represented by inflows from neighbouring compartments and stream losses that will vary according to a change in head. A conductance term accounts for these head-dependent inflows formulated in terms of Darcy's law, Equations (3.8) & (3.9):

$$Q = kiA = k \frac{\Delta H}{L} A = C \Delta H \quad (3.8)$$

$$C = \frac{k}{L} A \quad (3.9)$$

Where,

Table 6: Equations (3.8) & (3.9) abbreviations list.

Q	=	Flow (L ³ /T)	L	=	Length of flow (L)
k	=	Hydraulic conductivity (L/T)	ΔH	=	Head loss (L)
i	=	Hydraulic gradient (L/L)	C	=	Conductance (L ² /T)
A	=	Cross sectional area of flow (L ²)			

Dennis & Dennis (2017) substituted the Equation (3.7) into Equation (3.8) to obtain Equation (3.10):

$$h_t = h_{t-1} + \frac{R_t}{S_y} + \frac{[\sum_{i=1}^n C_i \Delta H_{it} + Q_{in}] - [\sum_{j=1}^m C_j \Delta H_{jt} + Q_{out}]}{AS_y} \quad (3.10)$$

Where,

Table 7: Equation (3.10) abbreviations list.

t	=	Current time step index	C_i	=	Conductance of inflow term i
n	=	Total number of inflow terms	C_j	=	Conductance of outflow term j
i	=	Inflow term index	ΔH_i	=	Change in head of inflow term i
m	=	Total number of outflow terms	ΔH_j	=	Change in head of outflow term j
j	=	Outflow term index	Q_{in}	=	Sum of all non-head dependent inflows
k	=	Specific Yield layer index	Q_{out}	=	Sum of all non-head dependent outflows

Equation (3.10) can therefore be used to not only estimate recharge within a system when time series head data is available but can also be altered to estimate aquifer recovery by simulating an increase in the head with time considering all inflows and outflows.

3.5 Calibration Evaluation Techniques

The performance of water budget models is commonly evaluated by comparing the simulated outputs with the observed outputs. How well these simulated values fit the observed data gives an idea of how accurate the forecast is to the natural system (Hwang, *et al.*, 2012; Legates & McCabe, 1999).

Evaluation of a model's performance solely by a visual evaluation of the graphical output may result in bias or error. Therefore, the evaluation process should consist of both graphical and numerical analysis. Numerical analysis techniques usually evaluate the "goodness of fit" or error across the model's simulated and observed measurements (Hwang, *et al.*, 2012; Legates & McCabe, 1999). Numerical analysis typically consists of three categories:

- Scale-dependent error measures
- Relative measures
- Relative correlation coefficient

Root Mean Square Error (RMSE), is a scale-dependent error analysis where its generally accepted that better model performance results in lower RMSE values. RMSE can be calculated using Equation (3.11):

$$RMSE = \sqrt{\frac{\sum (p_i - o_i)^2}{n}} \quad (3.11)$$

Where,

Table 8: Equation (3.11) abbreviations list.

<i>RMSE</i>	=	<i>Root Mean Square Error</i>
<i>p_i</i>	=	<i>Predicted/ simulated value at point i</i>
<i>o_i</i>	=	<i>Observed value at point i</i>
<i>n</i>	=	<i>Number of reference points</i>

Nash-Sutcliffe Coefficient (NSC) measures the relationship between the simulated data and the observed data to a relevant reference. NSC (Equation 3.12) ranges between negative infinity and 1.

$$NSC = \frac{\sum_{i=1}^n (o_i - p_i)^2}{\sum_{i=1}^n (o_i - \bar{o})^2} \quad (3.12)$$

Where,

Table 9: Equation (3.12) abbreviations list.

<i>NSC</i>	=	<i>Nash-Sutcliffe Coefficient</i>
<i>P_i</i>	=	<i>Predicted/ simulated value</i>
<i>O_i</i>	=	<i>Observed value for the n observations</i>
<i>$\bar{O}$</i>	=	<i>Mean of observed values</i>

Table 10 includes the general performance ratings for NSC (Moriassi et al., 2007).

Table 10: General performance ratings for NSC.

Value	Performance Rating
0.75 – 1.0	Very Good
0.65 – 0.75	Good
0.50 – 0.65	Satisfactory

The Pearson correlation coefficient (r) determines the correlation between two variables, reflecting the linear relations of the two continues variables (Xu & Deng, 2018). The Pearson correlation coefficient is most commonly used for measuring the quality of the fit by means of Equation (3.13).

$$r = \frac{\sum(O_i - \bar{O})(P_i - \bar{P})}{\sqrt{\sum(O_i - \bar{O})^2 \sum(P_i - \bar{P})^2}} \quad (3.13)$$

Where,

Table 11: Equation (3.13) abbreviations list.

r	=	<i>Pearson correlation coefficient</i>
P_i	=	<i>Predicted/ simulated value</i>
O_i	=	<i>Observed value for the n observations</i>
$\bar{O}$	=	<i>Mean of observed values</i>

Results vary from -1 (negative linear relationship) up to 1 (positive linear relationship), with 0 indicating no relationship between the simulated and observed values. The degree of linear relationship can also be presented as a %-value with a positive 0.70 or 70% correlation indicating a strong correlation.

3.6 Conclusion

In this chapter the study area was identified as being largely a karst landscape with the dolomitic strata forming the karst aquifer. It was established that within these karst systems, recharge takes place in two different ways: slow and diffuse recharge through the overlaying soil and small fissures as well as rapid concentrated point recharge through karst features. The chapter also considered different means of calculating recharge, including the use of a GIS-based method specifically developed for karst environments as well as different water budget methods.

Chapter 4 will include an in-depth investigation into the characteristics of the Far West Rand and how the geological conditions together with the hydrological settings influenced the natural formation of the hydrogeological aquifer system. The chapter will also focus on how dewatering commenced within each of the relevant sub-compartments and the effects thereof on not only the aquifer system but also the landscape.

CHAPTER 4:

CHARACTERISTICS OF THE FAR WEST RAND

4.1 Introduction

The FWR can in large parts be identified as a karst landscape. The chert rich dolomite of the Malmani sub-group within the FWR experienced various changes that included tilting and fracturing processes together with the erosion of the carbonate rocks. These alterations led to the formation of the vast well-developed karst landscape found today.

Chapter 4 will discuss how the evolution of the geomorphological conditions within FWR led to its complex geological features and as a result the formation of its consequent hydrological and hydrogeological conditions. This will be followed by an investigation into the dewatering activities present within the study area as well as the succeeding sinkhole formation and other karst features.

4.2 General Geology

The regional geology of the study area is dominated by the dolomitic rocks of the Malmani Subgroup. The Malmani dolomites dip to the south underneath the Pretoria Supergroup as seen in Figure 16 and are on average between 900 to 1000 meters thick within the FWR (Fleisher, 1981). The Malmani subgroup is predominately comprised of alternating layers of dolomite, dolomitic limestone and chert-rich dolomite with interbedded chert layers as seen in Figure 17. The subgroup comprises of the Frisco, Eccles, Lyttleton, Monte Christo and Oaktree formations, each characterised by colour and presence of interbedded chert (Fleisher, 1981). The subgroup bottoms with the Black Reef Formation that's mainly comprised of carbonaceous shales which can be seen outcropping north of the dolomitic Malmani subgroup in Figure 18. Furthermore, a number of dykes span the FWR in a general north to south direction. These dykes rarely outcrop and have been delineated through geophysical surveys (Enslin & Kriel, 1968; Fleisher, 1981)

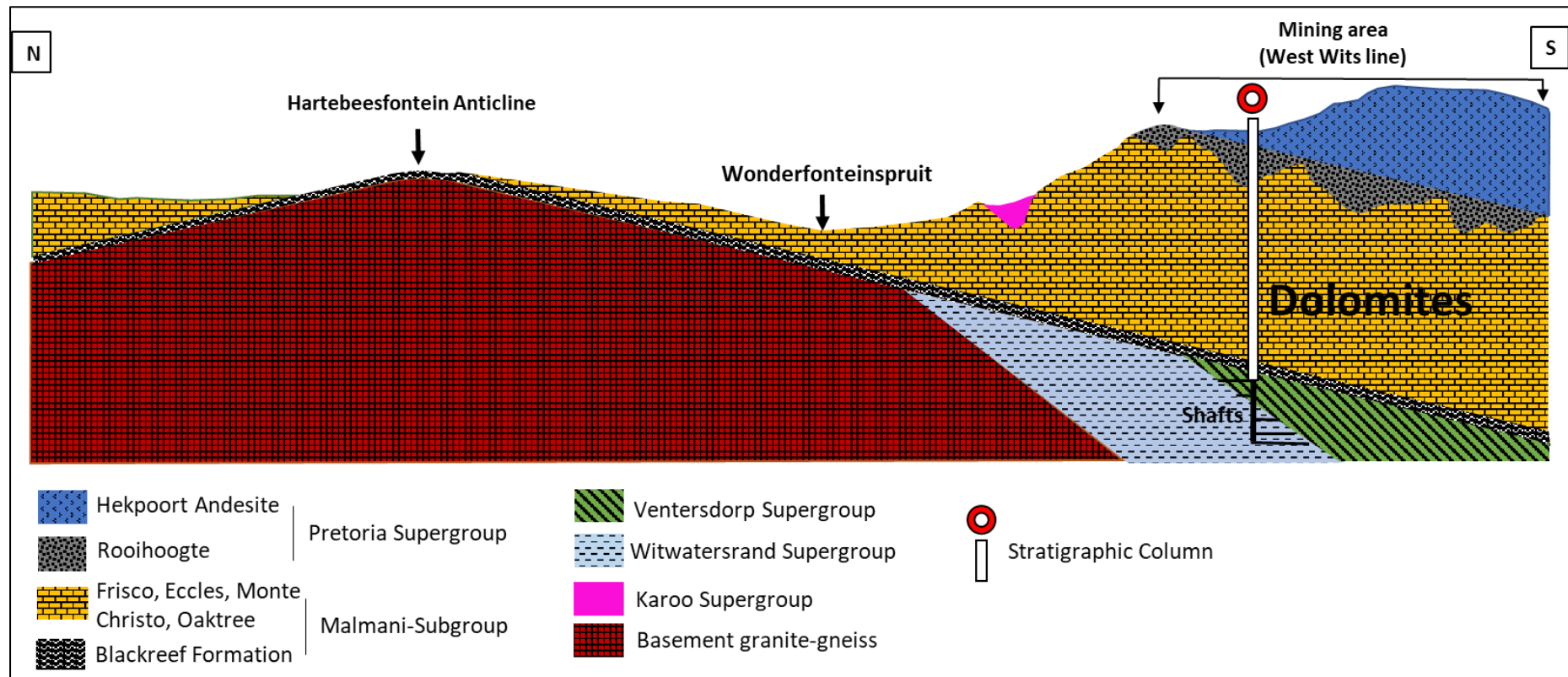


Figure 16: Geological cross-section of the FWR, modified after Morgan & Brink (1984).

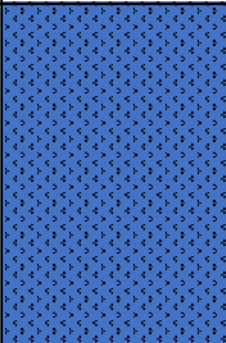
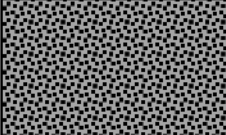
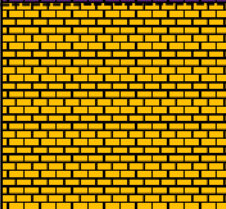
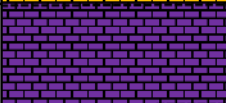
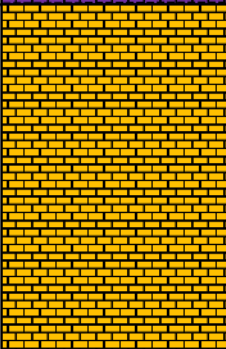
Thickness		Lithology	Formation	Sub-group	Group
200-500m		Andesitic lava and Tuff	Hekpoort Andesite		PRETORIA
50-200m		Quartzite, Shale	Timeball Hill		
		Diabase, Quartzite Chert Breccia	Rooihoogte		
30m		Chert-free Dolomite	Frisco	MALMANI	CHUNNIESPOORT
300m		Chert-rich Dolomite	Eccles		
150m		Dark Chert-free Dolomite	Lyttleton		
500m		Light Dolomite with abundant Chert	Monte Christo		
100m		Dark Dolomite	Oaktree		
10-20m		Carbonaceous Shale	Black Reef		
		Ventersdorp Supergroup (Lavas)			

Figure 17: Stratigraphic column of the FWR, modified after Fleisher (1981).

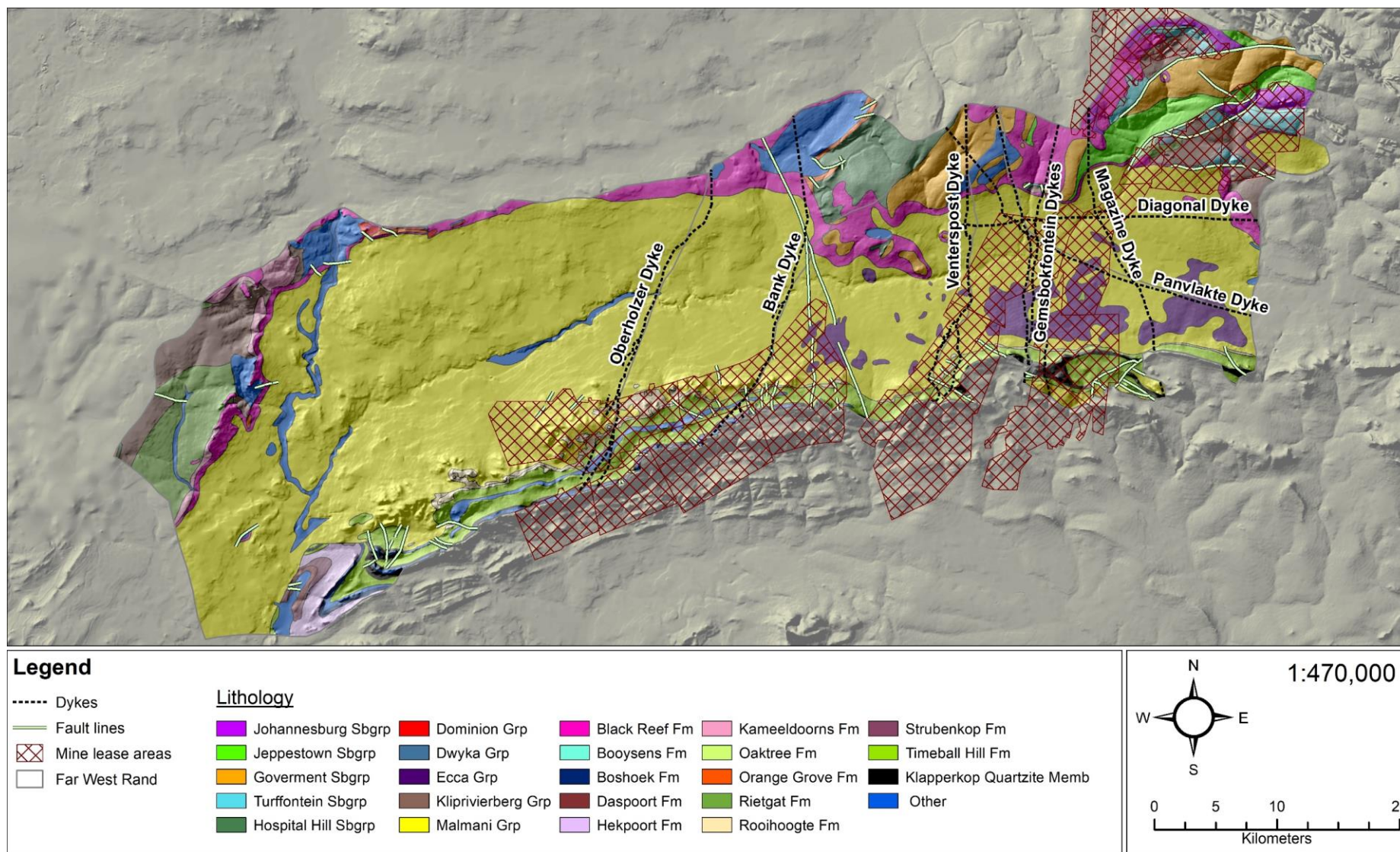


Figure 18: Lithology map of the study area.

The oldest group of rocks in the area are the ancient granite-gneiss Basement Complex which formed the floor of the Witwatersrand Basin. Sands, muds and gravels washed onto this basin forming the strata of the Witwatersrand Supergroup, thereafter tectonic movements tilted the Witwatersrand strata. After erosion smoothed the strata, the Ventersdorp lavas extruded onto them forming the Ventersdorp Supergroup. Uplift took place in the post-Ventersdorp and pre-Transvaal times, lifting the Johannesburg dome and Hartebeesfontein Anticline. This led to further localised tilting and fracturing. A long period of erosion followed, leading to the formation of the Transvaal basin where erosion debris laid down as the Black Reef Quartzite Formation 2000 million years ago. The waters occupying the basin at the time were rich in bicarbonates and silicates leached from the rocks of the Ventersdorp lavas and Basement granite-gneisses. The deposition of the dolomites resulted from chemical and organic precipitation with dolomite (calcium-magnesium-carbonate) deposited in alkaline and saline water conditions. Chert (silica) was deposited in mildly acidic conditions brought on by admixture of fresh river water from the intertidal zone (Figure 17). This resulted in the final interbedded layers of dolomite rich in chert, accounting for almost 20% in the top sequence of the Eccles Formation (Morgan & Brink, 1984).

In post-Malmani times, before the deposition of the Pretoria Group, a major uplift of the Johannesburg dome and Hartebeesfontein Anticline led to a long period of erosion in which a vast amount of dolomite was taken into solution. Later a brecciated chert zone was formed from the insoluble Chert and other substances creating the Rooihogte Formation on which the subsequent Pretoria Group was deposited. A final uplift of the Johannesburg dome and Hartebeesfontein Anticline together with the uplift of the Vredefort Dome ensured an erosion period exceeding 1600 million years. This resulted in the removal of most of the Transvaal Super-group, leaving the surface of outcropping dolomites as it is found today (Figure 18). Dwyka Tillite from the Karoo Supergroup was later deposited through glacial ice-scoured chasms in the dolomite somewhat 300 million years ago (Morgan & Brink, 1984).

4.3 Hydrology

The hydrological conditions of the study area have largely been shaped by its topography, geology and land cover characteristics driven by the climatic conditions in the FWR. The FWR falls mainly within quaternary catchments C23D, C23E and C23G that forms part of the larger Upper Vaal Water Management Area (WMA) (DWA, 2019).

In the hydrological sense, the FWR consists mainly of the Wonderfonteinspruit Valley drained by three major streams. For these, the main river: the Wonderfonteinspruit (WFS) consists of perennial and non-perennial channels along its runoff catchment. The WFS Valley has a total stream length of 416km with an overall drainage density, the ratio between the total stream length and the watershed area size consists of 0.25km/km² (De Klerk, 2019; Vafakhah, *et al.*, 2014). This low drainage density, according to De Klerk (2019) allows for a slower runoff time, resulting in more time for abstraction mechanisms such as infiltration to take place (Vafakhah, *et al.*, 2014).

With the predominate dolomitic sub-terrain, the Wonderfonteinspruit Valley contains characteristics of the typical karstic river system consisting of sparse soil cover with irregular surface drainage. This includes a lack of surface drainage as well as the brief disappearance of the Wonderfonteinspruit only for it to resurface down-stream. Although, karst geomorphological features including numerous caves, dolines, dry river valleys and landscape depressions/poljes occur within the study area. The semi-arid climate of the Highveld combined with the insoluble chert bedding found within the Malmani dolomites (Section 4.2) limits the frequency of surface karst features compared to typical karstic landscapes (De Klerk, 2019; Fleisher, 1981; Martini, 2006).

The Wonderfonteinspruit originates at Tudor dam near Krugersdorp, with the dam mainly used for storage by the Luipaardsvlei Gold Plant during the 1900's. Downstream of Tudor dam, Lancaster dam was also used as a storage dam but has since silted up. Downstream from Lancaster dam, mining activities dominate the area with several large tailings dams, and sand- and waste rock dumps in close proximity to the Wonderfonteinspruit. The river continues further downstream through a number of informal settlements and wetlands including a small dolomite inlier and over the small Witpoortjie Fault forming a possible ingress point into the Western basin mine void (De Klerk, 2019).

Onward the Wonderfonteinspruit flows into Luipaardsvlei dam, originally constructed to reduce stormwater. Since then problems within the Luipaardsvlei Dam outlet has limited the runoff (Van Biljon, 2017). From Luipaardsvlei Dam the river flows past agricultural and mining land use

areas into Donaldson dam. Consisting of two parts, Donaldson Dam has a combined storage capacity of approximately 1000ML (De Klerk, 2019). From Donaldson dam the water of the Wonderfonteinspruit is diverted into a 1m pipeline over the karstified terrain of the dewatered dolomitic compartments. Figure 19 shows the open channel/ parshall flume where the Wonderfonteinspruit is diverted into the 1m pipeline.



Figure 19: Open channel inlet for the Wonderfonteinspruit 1m diversion pipeline (Kriel, 2019).

The 32km pipeline (Figure 20) effectively replaces the lower-Wonderfonteinspruit, conveying overflow from Donaldson dam together with pumped volumes to the non-dewatered Boskop-Turffontein Compartment preventing water from returning to the aquifer system (Schrader, *et al.*, 2014b; Swart, *et al.*, 2003a; Winde & Stoch, 2010). However, due to the large amount of sinkholes present within this section, overflow from Donaldson Dam is lost due to infiltration into the aquifer system.



Figure 20: 1m diameter conveying pipeline (Kriel, 2019).

The pipeline ends nearly 3km northeast of Carletonville, merging with regional mine discharge canals and releasing water back into the Wonderfonteinspruit/ Mooirivierloop. The river makes its way further downstream passing through Harry's Dam followed by Abe Bailey Dam westwards to Welverdiend where its joined by water from mines along the western-mine lease areas. From there the river continues it's path through several wetlands followed by Coetzee Dam, Visser Dam and Turffontein Dam past the Turffontein springs finally merging with the Mooirivier (De Klerk, 2019), Figure 21.

In conclusion, the following remark was made by the Director of Water Affairs (1960) as quoted by Swart, *et al.* (2003a:757): "Prior to human settlement, the Lower Wonderfontein Spruit was a westerly draining perennial stream augmented by the springs". Subsequently, this all changed since the groundwater table was altered due to compartmental dewatering, leaving the Wonderfonteinspruit a non-perennial stream (De Klerk, 2019; Swart, *et al.*, 2003a).

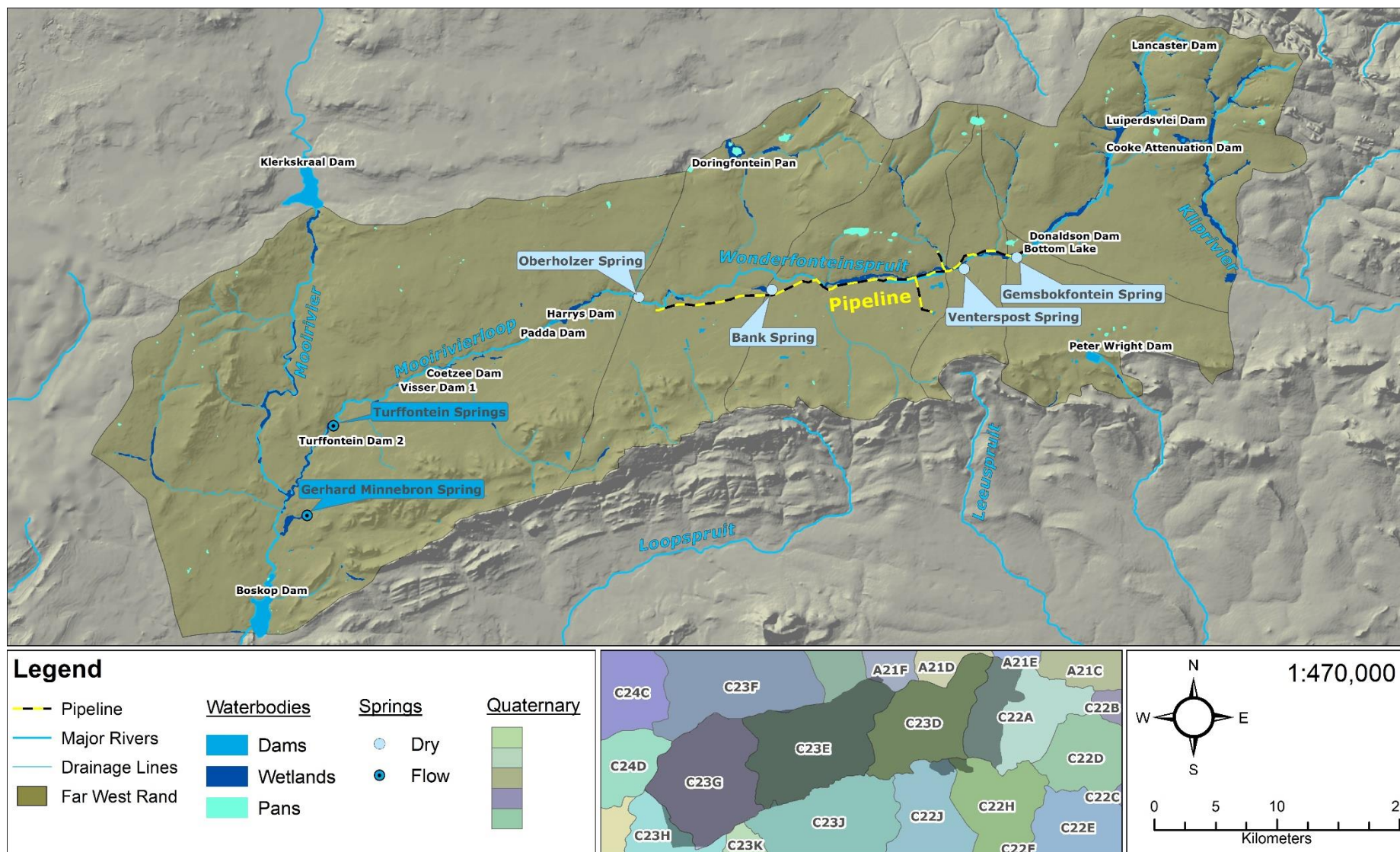


Figure 21: Hydrology map of the study area.

4.4 Hydrogeological Characteristics

The FWR is situated on top of the Malmani dolomites which consists of the majority chert interbedded formations found within the Chuniespoort Subgroup. According to Van Niekerk & Van Der Walt (2006) the Malmani dolomites can be up to 1450m thick within the FWR with the groundwater of the FWR sub-divided into two aquifer systems (Pulles, *et al.*, 2005):

- Perched aquifers, as the name suggests, consists of an aquifer system isolated on top of a saturated clay-like aquiclude that restricts groundwater interaction. In such case, this aquifer falls within the Karoo and Upper Transvaal strata isolated/perched by this clay and shale lithology. Groundwater mainly occurs within the upper 30m weathered zone with a deeper fractured aquifer up to 70m below ground level. This aquifer may be connected with the shallower weathered aquifer as well as the larger dolomitic aquifer (Freeze & Cherry, 1979; Pulles, *et al.*, 2005).
- Dolomite aquifers. These aquifers form the primary groundwater system of the FWR. These dolomites are in part highly karstified, forming an aquifer with the properties of storing large quantities of groundwater with increased porosity, storativity and conductivity (Pulles, *et al.*, 2005; Schrader, *et al.*, 2014b).

Dolomites are compact and impervious in nature with a porosity less than 3%, however, due to an abundance of highly developed fissures, joints and faults, rainwater can easily percolate through the rock mass (Morgan & Brink, 1984). Originally the dolomitic mass was effectively without any primary porosity but due to processes of dissolution, leaching and karstification the characteristics of the rock changed. These processes led to the development of the hydraulic conducting properties found in the carbonates resulting in shallow weathered zones throughout the FWR (De Klerk, 2019; Fleisher, 1981). The soluble carbonate rocks are susceptible to dissolution due to acidic waters that resulted from rainwater passing through the atmosphere and soil. In the vadose zone found above the water table, dissolution results in the widening of fractures and joints.

Chemical dissolution associated with the development of cavities and voids are however confined to the phreatic zone (Figure 22) below the water table. In the upper phreatic zone where the water is more acidic, large horizontal caverns are eroded into the rock. Deeper down where less acidic, slot corrosion continues to take place. The water within this zone is not static and therefore over long cycles of erosion the water table subsequently lowers, eroding deeper and leaving a karstified horizon in the vadose zone (Fleisher, 1981). With the bulk of the soluble dolomite dissolved, insoluble chert minerals leave the residual mass with low density and high void volume (Morgan & Brink, 1984).

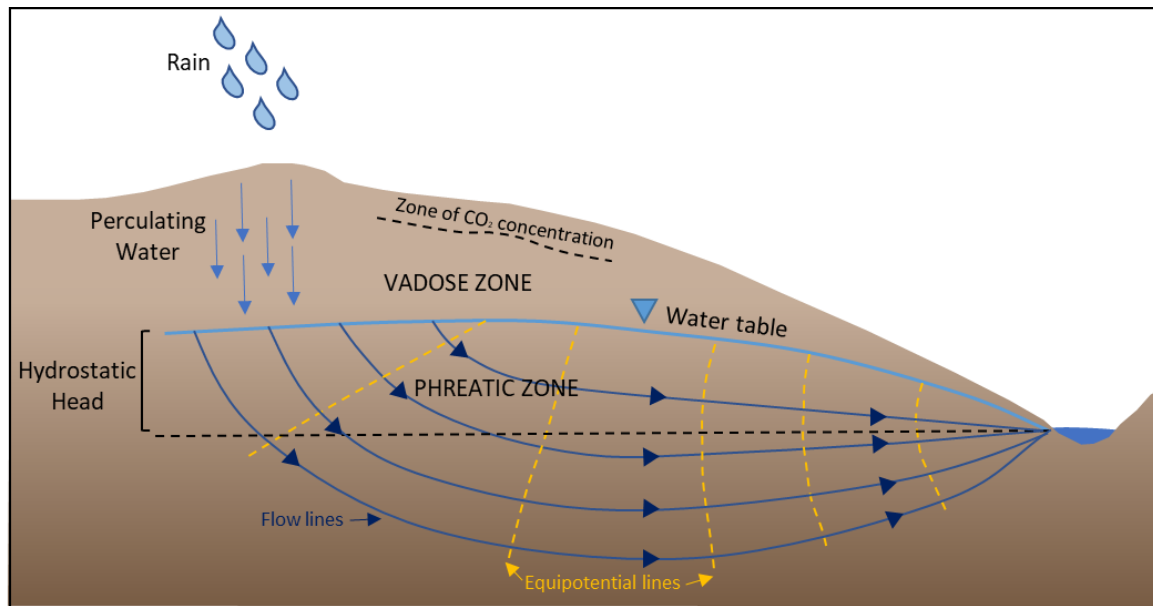


Figure 22: Water percolation in the Phreatic zone of the FWR dolomites, after Morgan & Brink (1984).

Therefore, the near-surface dolomites are highly weathered. It's estimated that nearly 75% of the groundwater is stored within this upper layer as indicated in Figure 23 (Schrader *et al.*, 2014b).

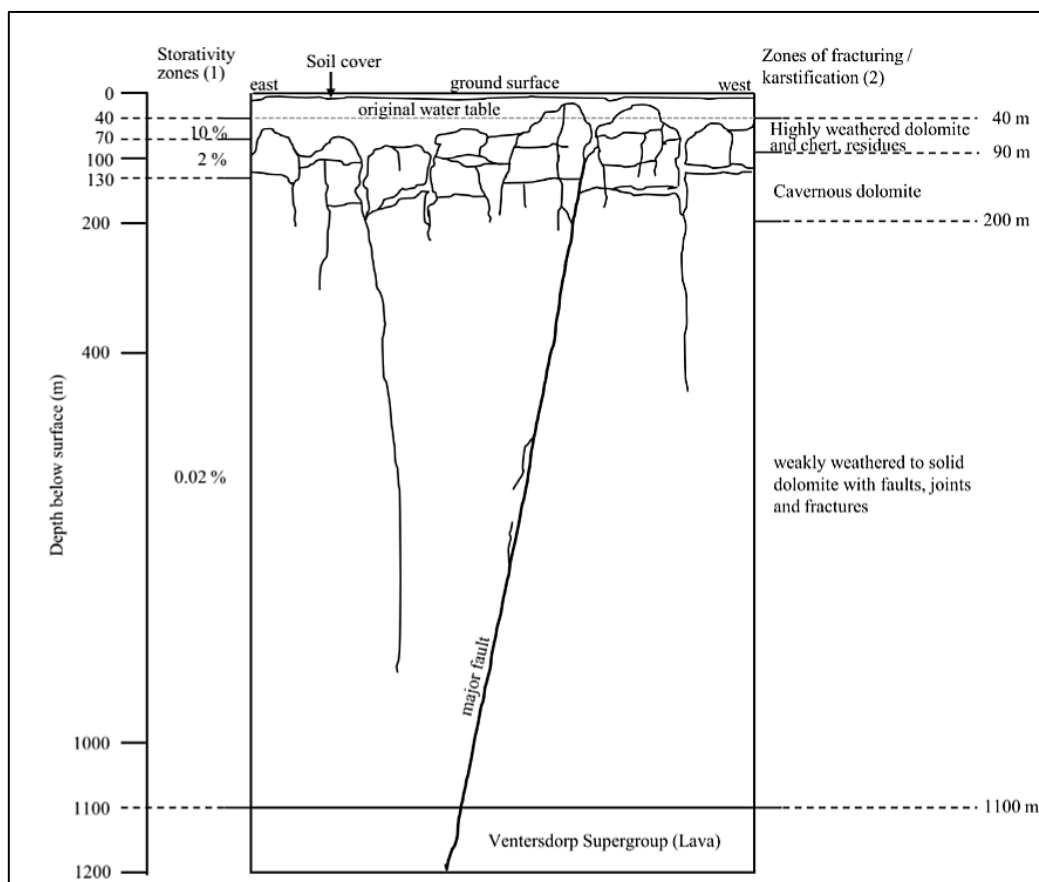


Figure 23: Zonation of the FWR dolomites due to karstification (Schrader *et al.*, 2014a).

The dolomitic aquifer system is delineated by the quartzites of the Witwatersrand Supergroup as well as the Ventersdorp lavas acting as a bottom boundary (De Kock, 1964; Schrader, *et al.*, 2014b). This boundary is created by the low storativity and transmissivity properties of these rocks. The granite-gneiss strata of the Hartebeesfontein Anticline from the northern aquifer boundary and the southern boundary is defined by unkarstified dolomites. This is caused by the Pretoria Group that overlays the dolomites with an increasing thickness to the south, shielding the dolomites from infiltrating water and therefore preventing karstification (Enslin & Kriel, 1968; Schrader, *et al.*, 2014b). The North-south trending dykes (Figure 18, Section 4.2) of which many are either syenite or diabase, cut through the dolomite. These 12- to 16m wide dykes act as aquicludes and subdivide the extensive dolomitic aquifer into smaller hydrogeological compartments (Fleisher, 1981; Swart, *et al.*, 2003a).

The study area includes six dolomitic compartments namely: Boskop-Turffontein, Oberholzer, Bank, Gemsbokfontein and Zuurbekom as seen in Figure 24. The elevation for each compartment's water table tends to be close to horizontal and controlled by its associated karst spring. Located upstream from each compartment's western dyke, these springs are commonly found in small depressions of micro-relief near the stream channel of the Wonderfonteinspruit (Schrader, *et al.*, 2014b). The springs of adjacent compartments differ in elevation by several tens of meters due to the dykes serving as groundwater barriers and forcing the water table to intersect with the surface (De Kock, 1964; Enslin & Kriel, 1968). Table 12 shows the volume of water from each spring before mining operations started. The Zuurbekom spring dried up due to pumping from the Rand Water Board with no record of original flow (Schrader *et al.*, 2014a).

Table 12: The dolomitic compartments' associated springs, after Usher & Scott (2001).

Compartment	Surface Area (m ²)	Spring	Average spring flow before mining (Mℓ/d)	Average of rain on compartment (Mℓ/d)	Proportion of rainfall that flows from eye %
Zuurbekom	100232000	Klip rivier	Dry*	190	
Gemsbokfontein	84951000	Gemsbokfontein	8.6	162	5.3%
Venterspos	54389000	Venterspos	20.9	104	20.1%
Bank	156690000	Wonderfontein	49	301	16.3%
Oberholzer	153845000	Oberholzer	54	295	18.3%
		Turffontein	18.4		
Boskop-Turffontein	704476000	Gerhardminnebron	52.7	134.9	5.6%
		Boskop	45		

In pre-dewatering times, the compartments were naturally recharged by means of rainfall-runoff as well as groundwater recharge through the streambeds of the Wonderfonteinspruit (Swart, *et al.*, 2003a). The springs or dolomitic eyes accounted for nearly all discharged water from the compartments and therefore reflected the amount of naturally recharged water received (Enslin & Kriel, 1968).

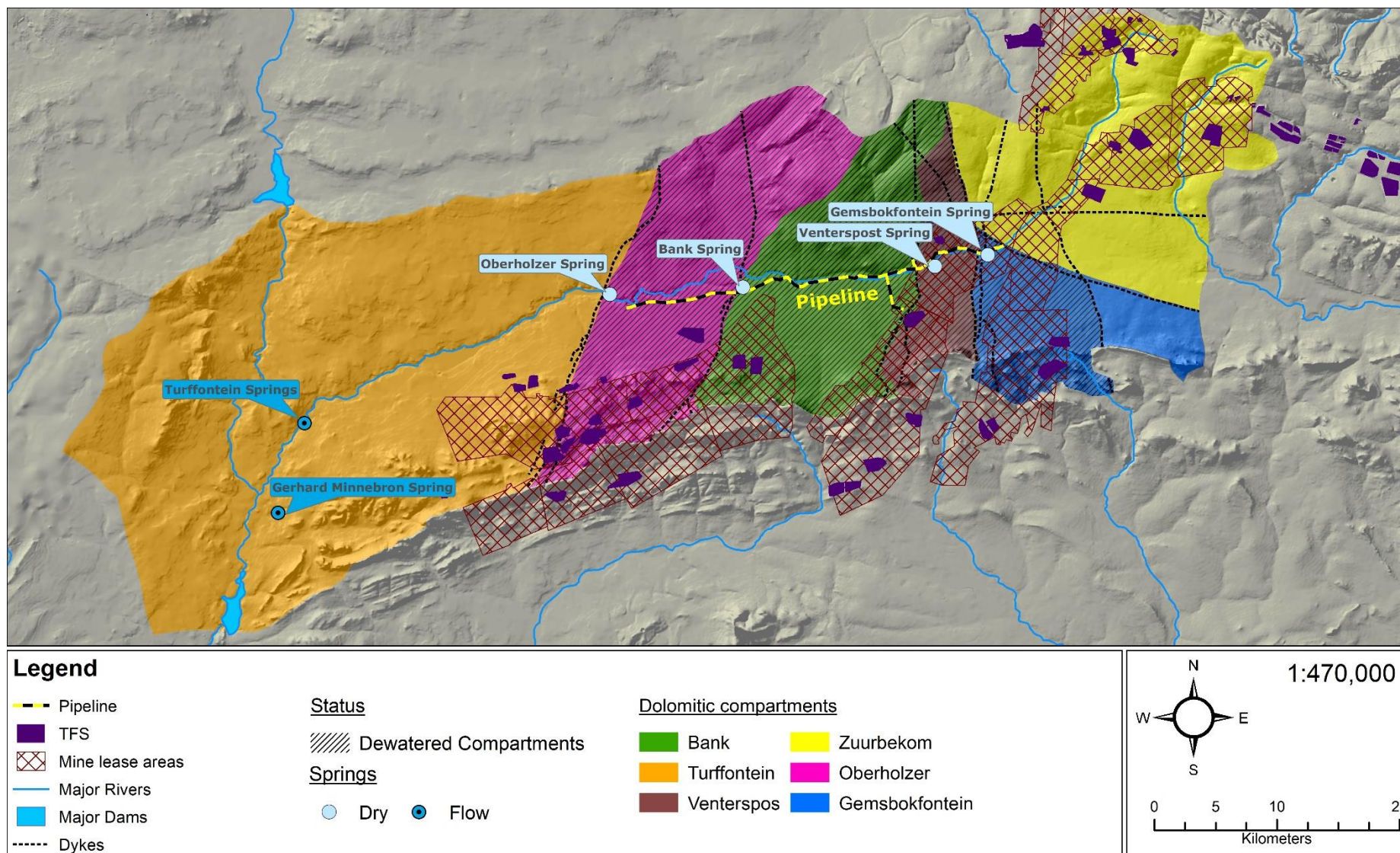


Figure 24: Hydrogeological map of the study area.

4.5 Mining and dewatering the FWR dolomites

As mentioned in Chapter 2 the discovery of the gold-bearing reefs created a gold rush event with early mining limited to reef outcrops. The development of technology during later stages enabled mining to go deeper, up to a point where shafts faced large quantities of water ingress. During the early mining stages underneath the unshielded dolomites of the Pretoria Supergroup, it was noted that water inflow increased as the mining progressed. A relationship between the areas mined and the inflow of water became clear as more water-bearing fractures were intersected (De Klerk, 2019; Swart, *et al.*, 2003a).

Mining within the FWR has not only produced voids underneath the dolomitic strata but also produced voids in the dykes that subdivided the dolomites into individual compartments causing inter-compartmental groundwater flow (IGF) (Figure 25):

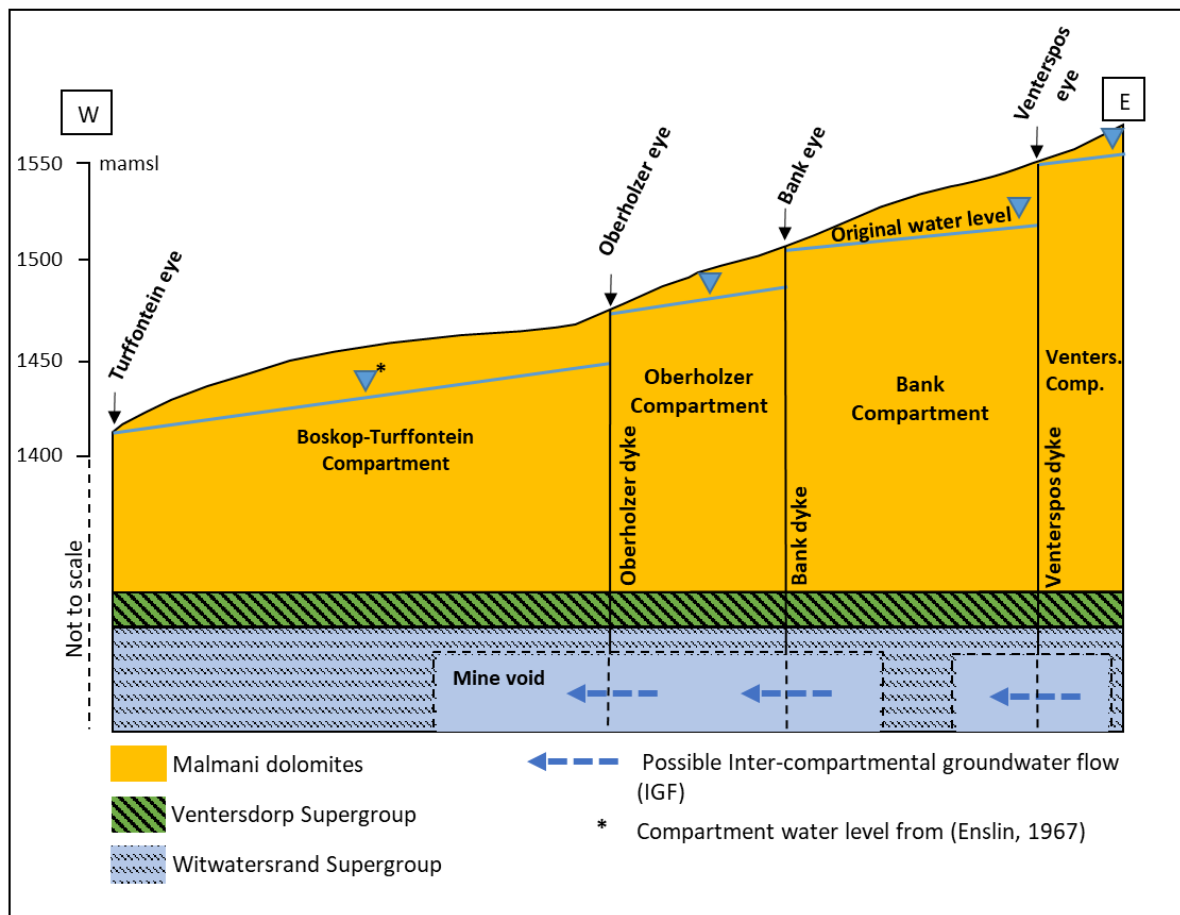


Figure 25: Conceptual model of the FWR mega-compartment, after Schrader *et al.* (2014b).

The mined-out voids underneath the dolomites will become an artificial aquifer with significant water storage capacity. This volume of mass rock hoisted from the mines was calculated at 600 Mm³ (600 Million m³) at the date of publishing by Swart, *et al.* (2003a) and needs to be taken into consideration as it will affect the overall rewatering process (Usher & Scott, 2001; Swart, *et al.*, 2003a).

Effective dewatering of the dolomites started in the 1940's to enable safe working environments for mine workers. Permission was given by the government that enabled mines to legally dewater the dolomitic compartments in which mining was actively ongoing. The mine dewatering reduces the inflow of groundwater from the dolomitic aquifers and comprises of removing the water that is in storage within the dolomites. It is important to note that although a compartment is dewatered, it still experiences inflow equal to natural recharge. Therefore, dewatering is achieved when the pumping rate = recharge rate (Usher & Scott, 2001). Table 13 indicates the inflow of groundwater experienced into mines in the different compartments.

Table 13: Groundwater mine inflow volumes, after Swart *et al.* (2003a).

Mine	Compartment	Influx from Dolomitic Aquifer (Mℓ/d)
Venterspost	Venterspost	28
Libanon	Bank	7
West Driefontein		34
West Driefontein	Oberholzer	20
Blyvooruitzicht		5
Doornfontein	Boskop-Turffontein	8

4.5.1 Abstraction in the Zuurbekom Compartment

The Rand Water Board has been the main abstractor of the Zuurbekom Compartments' groundwater since the turn of the 19th century at its Zuurbekom pumping station. The abstraction, mainly for domestic use, affected the natural discharge from the compartment causing the Klip River Eye to dry up. Although the eye occasionally flows during high rainfall seasons, there is no data supporting large flow volumes (Fleisher, 1981).

Conditions deteriorated as mining developed in the area during 1978. Water that infiltrated the underground working were abstracted together with water pumped from boreholes at shallow depth. Water was sourced through pumping within various shafts as well as disposed back into the aquifer or discharged through the Wondefonteinspruit, inadvertently affecting the water balance of the area. A total pumping figure of 10300MI (Megalitre or 1,000,000 litres) and returns of 2700MI into the dolomitic aquifer resulted in a net abstraction of 7600MI at the Cooke section of Randfontein Estates G.M. for the year 1980 (Fleisher, 1981).

4.5.2 Dewatering the Gemsbokfontein West Sub-Compartment

Up to 1966 groundwater abstraction in the Gemsbokfontein Compartment was around 40MI/month. The abstraction amount was drastically increased over the years to react to the inflow into the mine workings of Western Area. Within a 26-month period between January 1971 and February 1973, an accumulated total of 24320MI was pumped from the compartment. The result from all this pumping led to the considerable lowering of the water table and caused the development of sinkholes together with subsidence in areas with historical shallow water levels. The Gemsbokfontein Eye subsequently dried up in 1972 due to the lowered water table (Fleisher, 1981).

In a comparison of hydrographs during 1977-1978, it was found that boreholes on the eastern flank of a discovered divide react differently to those on the western side. This divide was witnessed as a central groundwater mound within the compartment and confirmed as a possible dyke structure. Therefore, the Gemsbokfontein Compartment can be sub-divided into a western and an eastern compartment (Fleisher, 1981). Active dewatering commenced in 1986 within the Gemsbokfontein West Sub-Compartment and up to 1986 the water pumped from underground was recirculated back into the compartment. A permit obtained in 1986 allowed for the water to be pumped from the compartment into the Leeuspruit and Kleinwes Rietspruit, not to recirculate within the compartment (Van Biljon, 2017).

The groundwater level ranges between 4m to 40m below ground level within the non-dewatered Gemsbokfontein East Sub-Compartment (Van Biljon, 2017). The dewatering of the Gemsbokfontein West Sub-Compartment resulted in 84MI/d from three shafts. The groundwater levels of the western compartment exceed 400m to manage water ingress (Van Biljon, 2017). The decreasing pump volumes in Figure 26 indicate that lower inflow volumes were encountered as the groundwater level in the Gemsbokfontein West Compartment entered a more pseudo steady-state condition.

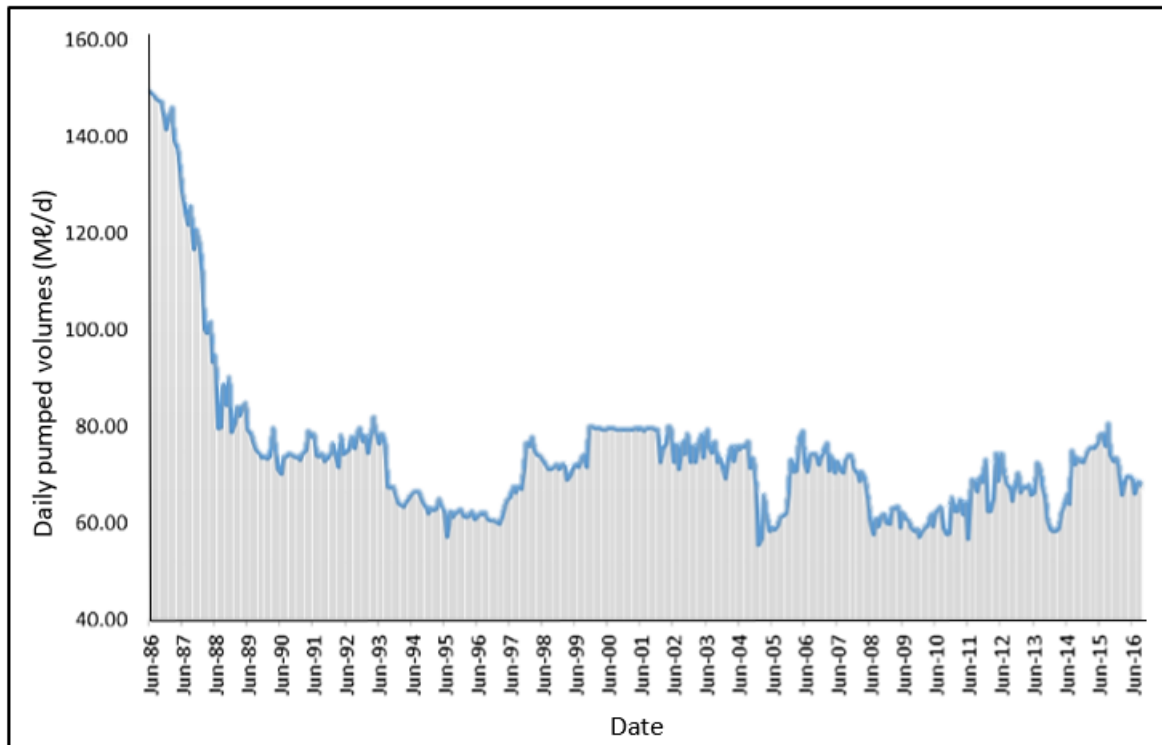


Figure 26: Gemsbokfontein West pumping volumes from Ezukwini Mine, from Van Biljon (2017).

4.5.3 Dewatering the Venterspost Compartment

As mining progressed in the FWR, the sagging of rocks resulted in the formation of ruptures which generated large volumes of ingressing groundwater from the dolomites. Venterspost Mine responded by sinking surface wells in a bid to pump large volumes of water and drop the hydraulic head, causing Venterspost to be the first compartment to be dewatered.

Farmers who depended on the Venterspost Eye in turn received the surplus water pumped from the mine in the late 1930's (Stoch & Winde, 2010). By 1945 it was clear that pumping had exceeded the recharge rate from rainwater (Wolmarans, 1984). The Venterspost Eye eventually ran dry in 1947, this was followed by the formation of sinkholes to such an extent that the area became known as the '*sinkhole farm*' (Stoch & Winde, 2010). By the mid-1950's the mining group embarked on a full-scale dewatering project as the recirculation of pumped water proved too expensive. The full-scale dewatering would also prevent further sudden and catastrophic underground floods (Swart, *et al.*, 2003b). During 1955 it became clear that cementing is no longer required (Wolmarans, 1984). In 1960 it was estimated that it would be a greater financial advantage to mine in a dewatered state than agricultural contributions within a non-dewatered state (Swart, *et al.*, 2003b). Figure 27 indicates the dewatering of the Venterspost Compartment between 1940 and 1956 by Venterspost Mine.

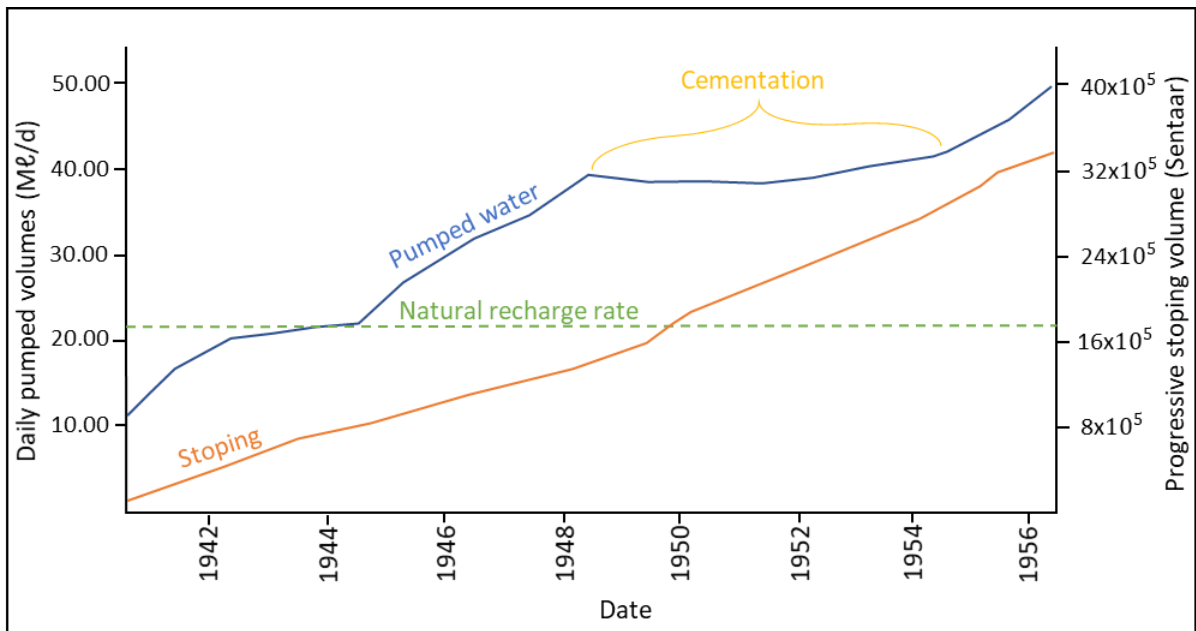


Figure 27: Dewatering of Venterspost Compartment from Venterspost Mine, modified after Wolmarans (1984).

The peak pumping rate was reached in 1961 when Venterspost Mine pumped at 57M/d. In 1977 the 1m-diameter pipeline was constructed and used as an aqueduct to channel pumped water from the Venterspost Mine across the Bank and Oberholzer Compartments.

4.5.4 Dewatering the Bank Compartment

A hazardous inrush in October 1969 occurred after stoping against the base of the non-dewatered dolomites of the Bank Compartment encountered a major fault, the so-called '*Big-Boy Fault*'. The event led to the collapse of the stoping area's roof as well as the flooding of West Driefontein. The Bank Compartment was set to be dewatered to provide a safe and economically viable environment (Schrader, *et al.*, 2014a; Swart, *et al.*, 2003b). In 1969 permission was given for the dewatering of the Bank Compartment. The gradual lowering of the water table reduced the pressure above the mine void, consequently reducing the ingress volume (Schrader *et al.*, 2014a). The dewatering of the Bank Compartment peaked in 1970 with 340M/d pumped from the Driefontein Consolidated Gold Mine (Swart *et al.*, 2003b).

Figure 28 indicates the monthly pumped volumes from the Bank Compartment. The dewatering activities up to 1974 created a steep cone of depression in the south-west corner of the compartment at the former inrush point (Schrader *et al.*, 2014a) as shown in Figure 29.

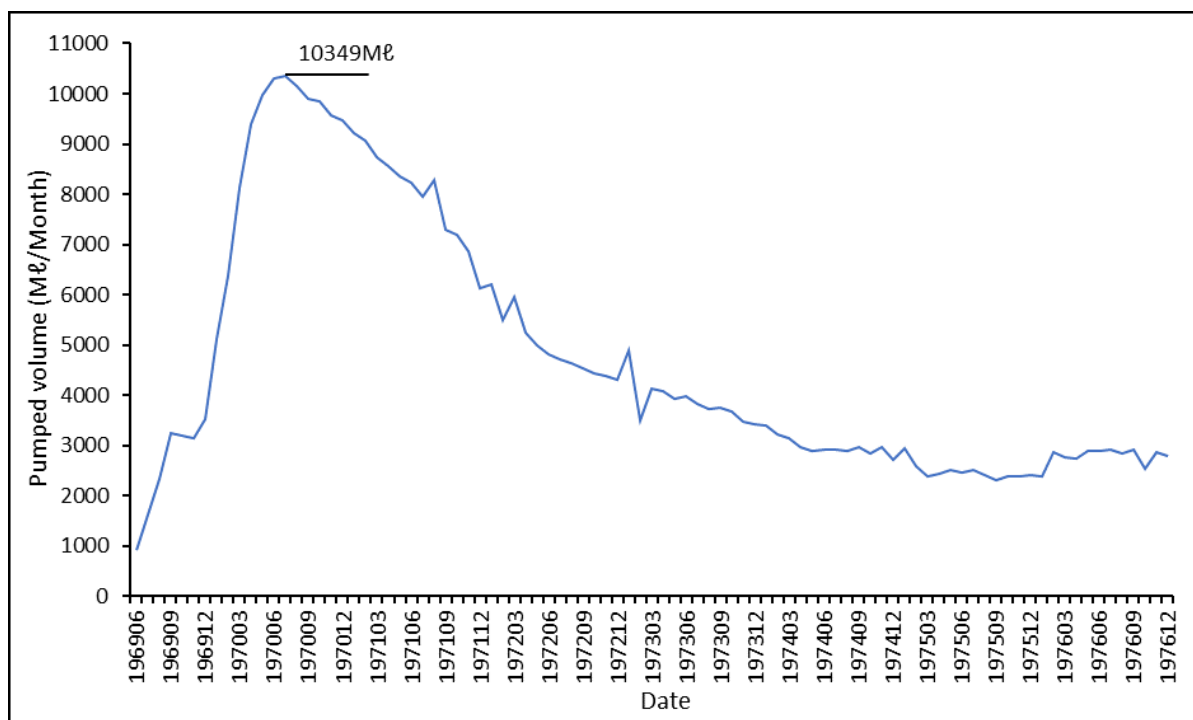


Figure 28: Monthly pumped volumes of the Bank Compartment, modified after Fleisher (1981).

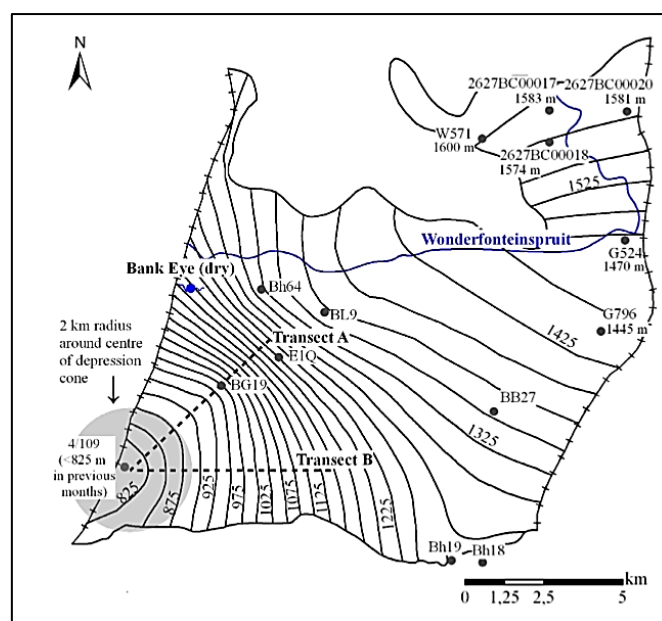


Figure 29: Cone of depression caused by the dewatering of the Bank Compartment, from Schrader *et al.* (2014a).

An estimated 90% of all water pumped from the Driefontein mines re-entered the mine voids as artificial recharge by means of the “Big-Boy fault”, this was characterized by a period of water loss as the mine voids and compartments recharged (Swart *et al.*, 2003a; Stoch & Winde, 2010).

4.5.5 Dewatering the Oberholzer Compartment

The effective dewatering of the Oberholzer Compartment started in the mid-1950's and resulted in a gradually lowering water table. As a result of this dewatering effort, the Oberholzer spring ran dry in September 1959 (Schrader, *et al.*, 2014b; Stoch & Winde, 2010). The legal approval to dewater the compartment followed after a 4-year investigation that argued in favour of dewatering based on the fact that a point of no return has been reached (Schrader & Winde, 2014; Stoch & Winde, 2010). The dewatering of the Oberholzer Compartment peaked around 1963 at 146ML/d pumped by Driefontein Consolidated (Swart *et al.*, 2003b) as shown in Figure 30.

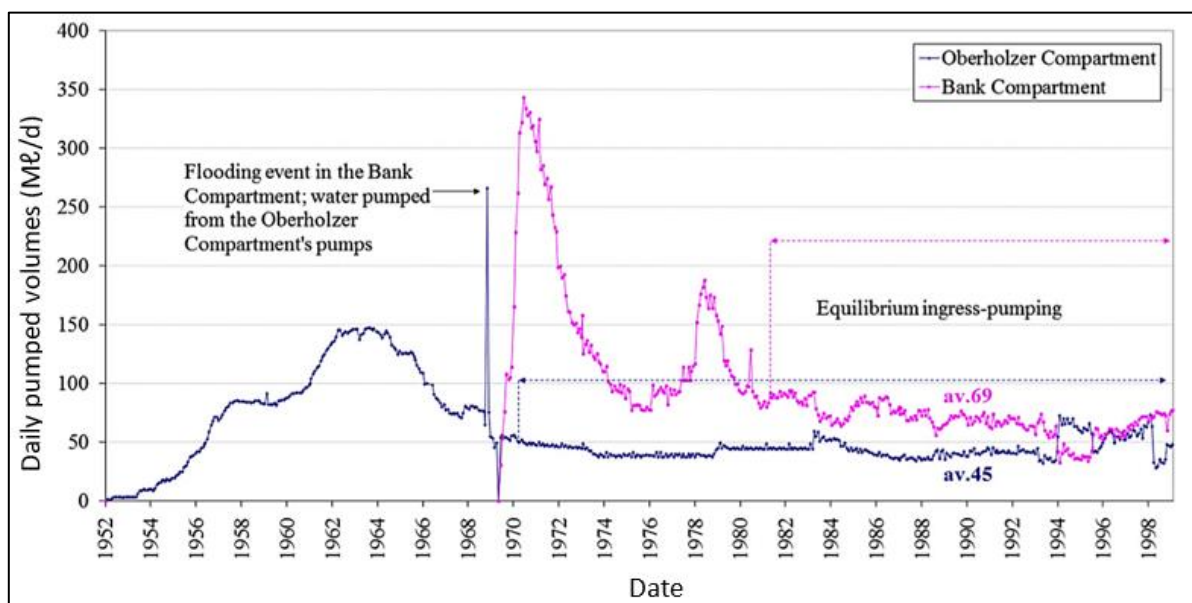


Figure 30: Daily pumped volumes of the Oberholzer and Bank Compartments (Schrader *et al.*, 2014b).

The pumping rate of the Oberholzer Compartment more or less reflects the compartments' natural recharge. The dewatering chart (Figure 30) with pumping rate against time is similar to a hydrograph with a rising and falling trend, that tapers into a constant low rate (Schrader *et al.*, 2014b).

The rising portion represents an increase in ingress caused by intersecting additional water-bearing fissures. A culmination point is reached where ingress and pumping volumes are at a maximum, at this point ingress increase from the void expansion is higher than ingress decrease from the lowered hydraulic head. The culmination point is surpassed with a reduction in ingress from reducing water levels. The final flattened portion represents a situation where naturally generated groundwater flows into the mine void. This observation assumes large connections between the cavernous zone and the mine void in the form of fractures, faults and fissures (Schrader *et al.*, 2014b).

4.5.6 The Boskop-Turffontein Compartment

Although there is a presence of gold mining within the compartment by Doornfontein Gold Mine, there is no need for any dewatering. The fractures connecting the mine void with the overlaying dolomites are compression faults sealed with mylonite. Therefore, the groundwater within the Boskop-Turffontein Compartment dolomites do not pose any risk (Swart *et al.*, 2003b). However, the pumping from Driefontein- and Blyvooruitzicht Mine has caused a groundwater depression on either side of the Oberholzer dyke affecting the groundwater level in the Boskop-Turffontein Compartment. Another, smaller groundwater depression is present at Doornfontein Mine due to groundwater abstraction (Fleisher, 1981).

The groundwater abstracted from the dolomitic aquifer by the gold mines in the FWR are diverted west of the Oberholzer dyke, into the Boskop-Turffontein Compartment. This water is distinguished from two groups of mines. Group one includes Venterspost Mine and the Driefontein Consolidated Mines. With group two including Blyvooruitzicht, Western Deeps and Doornfontein Mines. Discharged water from Donaldson dam and occasional flow from the Gemsbokfontein Eye contributed to the overall flow into the Boskop-Turffontein Compartment from mid-1977 (Fleisher, 1981; Swart *et al.*, 2003b).

The discharged water from the dewatering episode of the Oberholzer Compartment between 1954 and 1967 manifested itself with an increase in Turffontein spring discharge. Figure 31 shows how the artificial recharge from the dewatering episode in green increased the flow of this spring in red compared to the yellow Maloney's spring that was unaffected, being situated within the Mooirivier river basin in the Turffontein Compartment.

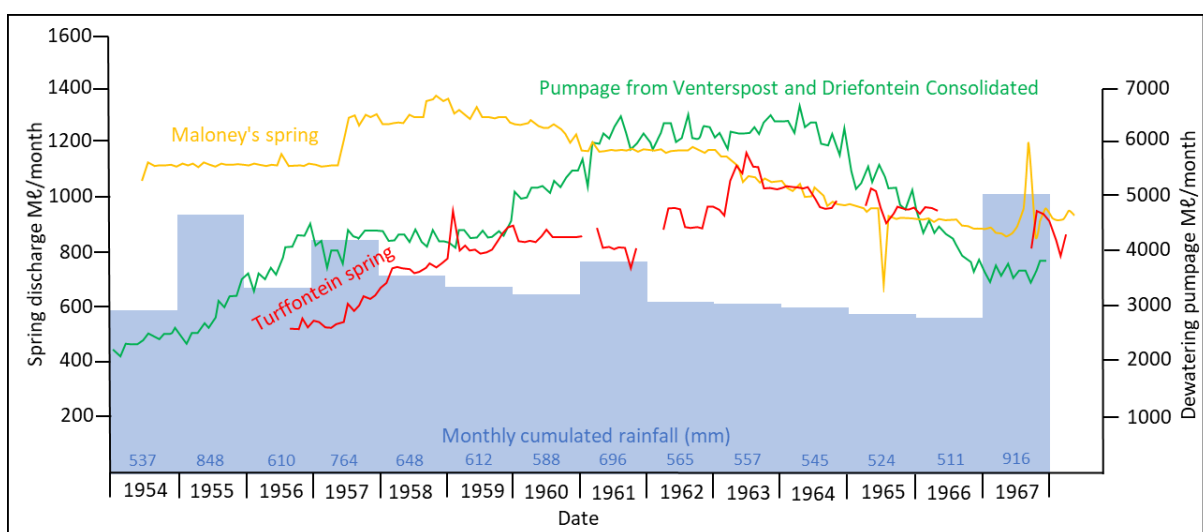


Figure 31: Spring flow response into artificial recharge, modified after Fleisher (1981).

4.6 Sinkhole formation in the FWR

Since the start of the compartmental dewatering in the FWR, 589 sinkholes formed up to 1984 within the Venterspost, Bank and Oberholzer Compartments. The majority of the Transvaal dolomites has an absence of naturally occurring dolines and sinkholes although sinkholes due to factors other than dewatering has been encountered, accounting for only 52 of the 589 sinkholes formed since 1957 (Fleisher, 1981; Wolmarans, 1984). It was an established fact that sinkholes would follow the drop in the water table brought on by dewatering (Stoch & Winde, 2010).

Sinkholes within the region are generally a roughly cylindrical or conical shape and vary in depth up to 50m, with a diameter between 1-100m. Dolines or substances are shallow depressions up to 5m deep and can vary up to 1km in length. Of the two, dolines develop slowly and compared to the rapid development of sinkholes are less dangerous. Sinkholes can occur without any warning and have catastrophic results (Swart, *et al.*, 2003b).

Sinkhole formation involves the mobilization of a blanketing layer, which consists of unconsolidated material such as chert rubble within the upper overburden overlying the dolomitic bedrock. The blanketing layer is mobilized into receptacles which present themselves as small openings, interconnected openings or larger openings such as caves within the dolomite. Furthermore, overburden is transported into these receptacles through a conduit within the sinkhole which enlarges through collapsing caused by a loss in shear strength brought on by wetting. The thickness of the blanketing layer determines the size and shape of a sinkhole. Mobilizing agents include excessive water seepage, groundwater level drawdown and seismic activities (Jennings, 1966; Swart, *et al.*, 2003b). Of these, the water-related agents are present in the form of compartmental dewatering and the artificial recharge from rewatering.

The impact of sinkholes, dolines and other cavernous karst structures need to be factored in when aquifer parameters are calculated (Stoch & Winde, 2010). Sinkholes that formed during the 1980s until present have mostly been left open, allowing direct recharge to occur from rainfall-runoff with runoff diverted purposely into sinkholes in some cases.

For a brief time period, mines affected by the sinkhole related ingress attempted to fill them with materials ranging from waste rock to slimes materials and covering them with topsoil to aid vegetation growth. This initiative was short lived as erosion from rainfall and runoff removed the fill materials reactivating some of the sinkholes (Dill, *et al.*, 2007; Swart, *et al.*, 2003b).

The aerial photograph in Figure 32 displays a typical sinkhole cluster found within the dewatered Venterspost Compartment. The sinkhole cluster displayed contains more than 70 sinkholes within an area of 0.67Km². Evident in the grey-overburden, it's clear that some attempts have been made to fill some of these sinkholes, but many opened up again.



Figure 32: Sinkhole cluster located in the Venterspost Compartment (NGI, 2015).

The karst features of the FWR extend beyond the sinkholes, springs and dolines with several cave systems that developed within the area. These cave systems include some of the largest in South Africa with most developed along tectonically induced faults and joints. The cave systems consist of mainly NNE-SSW and WNW-ESE passageways up to 12.4km long and 185m below the surface level (De Klerk, 2019; Foster, 1989; Swart, *et al.*, 2003b). The entrance into such a cave system within the FWR can be seen in Figure 33, indications of the weathered dolomitic rocks can be seen to either side of the cave entrance.



Figure 33: Entrance to Wonderfontein cave located within the FWR (Van Wyk, 2019).

Through the utilization of aerial photography, sinkholes were mapped within the FWR as seen in Figure 34. A total of 514 collapsed sinkholes were mapped together with 155 sagging doline structures, all within the Malmani dolomites. There is a clear correlation between sinkhole development and the dewatered and partially dewatered compartments of the FWR along the WFS river.

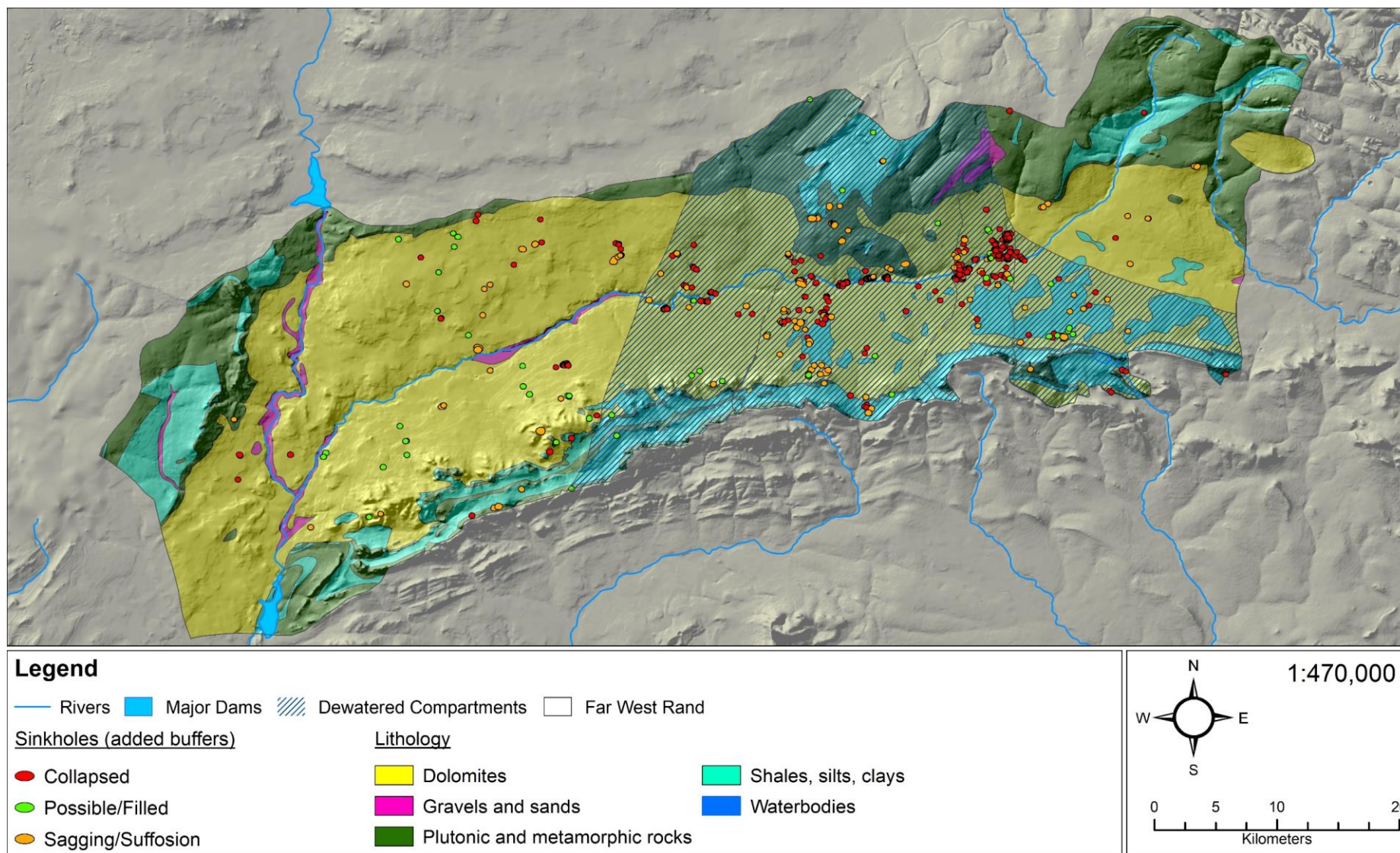


Figure 34: Karstification map of the study area.

CHAPTER 5:

DATA ANALYSIS

5.1 Introduction

From Chapter 4 it is clear that the FWR is a large, complex area that's hydrogeological characteristics depend on regional differences in the geomorphology and geology driven by the climatic conditions. As mentioned in Chapter 2 many researchers have conducted studies within the area. Each researcher implemented a different methodology for either the estimation of past and future recharge estimations; present and future hydrological conditions; the subsequent impact of dewatering on the geology and hydrogeological systems and most importantly the recovery of the groundwater levels when mining activities ceased. This chapter will include a brief look into these previous studies, the selected water budget model as well as the data requirements.

5.2 Previous studies on determining the FWR parameters

5.2.1 Enslin & Kriel (1968): "The Assessment and Possible Future Use of the Dolomitic Groundwater Resources of the Far West Rand, Transvaal, Republic of South Africa."

Enslin & Kriel (1968) found that groundwater contours can be used to calculate the permeability and specific yield of the dolomitic aquifer. Recharge was found to equal the mean flow of the spring or spring discharging within the compartment, neglecting losses through leakage, evapotranspiration and artificial abstraction (Table 14). The dewatering of the Venterspost Compartment helped assess the storage function of the aquifer as the combined pumpage within the Venterspost Compartment exceeded the natural recharge that's assumed equal to the normal pre-dewatering spring flow rate.

Due to water utilisation by the Rand Water Board that caused the associated spring in the Zuurbekom Compartment to dry up, it was assumed that the spring flow equalled that of the pumped volumes. The specific yield of the Venterspost Compartment was calculated using an analysis of the groundwater equation and the compilation of monthly and annual rainfall balances. The results from the study was compared with the specific yields derived from a study where the volume percentage of weathered rock penetrated by boreholes and shafts in Table 15.

Table 14: Percolation estimation, after Enslin & Kriel (1968).

Compartment	Dolomitic Catchment Area (Sq. mile)	Percolation to groundwater (Inches per Sq. mile)				Percolation direct from rainfall	
		Ave. spring flow	From streambed, canals and irrigated	From non-dolomitic areas	Directly from rainfall	% Rainfall	Mℓ/d
Zuurbekom	38.799	*3.64"	0.10"	0.10"	3.44"	13	20.63
Gemsbokfontein	32.8	1.86"	-	-	1.86"	7.5	7.19
Venterspost	21	6.93"	2.83"	1.82"	2.28"	8.5	5.68
Bank	60.5	4.49"	1.63"	1.40"	1.46"	5.8	13.25
Oberholzer	59.4	5.04"	2.84"	1.27"	0.93"	3.6	8.33

*pumpage by Rand Water Board

Table 15: Specific yield for the Venterspost Compartment, after Enslin & Kriel (1968).

Weighted depth below surface	Specific yield	
	Study of dewatering data, Venterspost compartment	Analysis of logs of shafts and boreholes
200' or 61m	9.10%	9.70%
250' or 76m	5.50%	6%
350' or 107m	2.60%	3%
415' or 126m	2.00%	2%
480' or 146m	1.30%	1.60%

5.2.2 Foster (1988): “The Groundwater Resources of the Far West Rand Dolomitic Compartments.”

Foster (1988) studied the pumping and water level data of the FWR in order to obtain realistic figures for effective aquifer porosity and recharge rates. After a detailed study of the data, it was clear that estimations of effective aquifer porosity would not be possible and only the recharge rates could be estimated. The pumpage and water level data from the Oberholzer and Venterspost Compartments appeared to be in a state of equilibrium, enabling for the approximation that the pumping rate at the time equaled that of the recharge rate. The recharge figures (Table 16), also include lateral groundwater inflow that originates from cross dyke groundwater flow.

Table 16: Recharge estimations, after Foster (1988).

Compartment	Area (x10 m ²)	Annual Pumpage (Mℓ)	Estimated Natural Recharge (m) per unit area	Recharge as % of Mean Annual Rainfall
Oberholzer	203.8	19000	0.09	12.9
Bank*	165	32000	0.19	27.3
Venterspost	57.8	22000	0.38	54.6
Combined Compartments	426.6	73000	0.17	24.6
* Hypothetical figure to determine maximum limit of recharge				

The Bank Compartment's groundwater level was in a declining state at the time of study and therefore the recharge rate was less than the pumping rate at that time. Consequently, the recharge is less than 27.3% of the annual rainfall. Foster (1988) explained that the high recharge value obtained for the Venterspost Compartment could be explained by additional volumes of water pumped from Libanon mine originating in the Bank Compartment.

To calculate the groundwater storage of these compartments Foster (1988) multiplied the effective porosity obtained from Table 17 (Enslin & Kriel, 1968) and Table 13 (Foster, 1988) with the applicable compartment area and the aquifer thickness.

Table 17: Effective porosity estimates, modified from Foster (1988).

Area/compartment	Aquifer thickness analysed (m)	Effective porosity %	Analytical method	Source
Bank	60	2.8	Planimetric/volumetric	(Fleisher, 1981)
	(a)	1.0	spring analysis	(Fleisher, 1981)
	(a)	1.55	Analysis of flooding and dewatering of West Driefontein	(Schwarz & Midgley, 1975)
Zuurbekom	48	3.0	Chemical mass balance	(Fleisher, 1981)
Gemsbokfontein	(a)	2.26	Chemical mass balance	(Fleisher, 1981)
Gemsbokfontein west	36	3.4	Planimetry of dewatering	(Foster, 1987)
(a) - Zone of groundwater circulation				

From these findings, the effective porosity profile for the aquifer included levels less than 250m and therefore: a depth of 0-100m was given an effective porosity of 6% and 100-250m of 2%.

5.2.3 Usher & Scott (2001): “Post Mining Impacts of Gold Mining on the West Rand and West Wits Line.”

Usher & Scott (2001) determined that in post-mining times the recharge of the mine openings and dewatered compartments would rewater at the same rate at which the springs had flowed in pre-dewatering times. This was based on the notion developed by Enslin *et al.* (1976). It was determined that the time to fill could be calculated from the mined volume and the steady-state spring flow rate. Uncertainties included the effects of IGF. As mining excavations have flooded, the dewatered storage would be recharged. This volume could be estimated from the difference between the amount of water pumped since dewatering and the volume that would have flowed from the springs over the same time period in a non-dewatered state (Usher & Scott, 2001). Table 18 indicates the storage in the dolomites as well as the time required to rewater the storage estimated from steady-state pumping tests.

Table 18: Dolomite storage and rewatering, after Usher & Scott (2001).

Compartment	Total volume removed		Dynamic volume from mine records		Dynamic volume from DWA records		Storage in compartments (Mℓ) -volume removed		Time to rewater dolomites (years)	
	Period	(Mℓ)	Constant dewatering rate (Mℓ/d)	(Mℓ)	Eye yields (Mℓ/d)	(Mℓ)	Mine records	Eye flows	Mine records	Eye flows
Bank	1969-1996	1000000	70	690000	48	470000	310000	530000	18.3	30
Oberholzer	1952-1996	1100000	50	800000	55	880000	300000	220000	15.2	11
Venterspost	1968-1996	410000	40	400000	21	210000	3240	190000	0.42	25
Gemsbokfontein	1986-1996	220000	?		8.9	42000		170000		54

Because the West Gemshokfontein Compartment’s pumped water was recharged back into the East Gemshokfontein Compartment, the water leaks through the dividing dyke causing rewatering within the West Gemshokfontein Compartment with no clear drawdown. Therefore, preventing any calculations for this compartment (Usher & Scott, 2001).

5.2.4 Swart *et al.* (2003a): “The future of the dolomitic springs after mine closure on the Far West Rand, Gauteng, RSA.”

Swart *et al.* (2003a) used historical surface- and groundwater flow data together with pumping rate data to comprise a geohydrological model for the FWR. Build on a theory that the water tables within each compartment will return to pre-mining elevations. Swart, *et al.* (2003a) calculated the inter-compartmental flow across the pierced dykes found in between each dolomitic compartment using Darcy’s law for saturated flow through porous media (Dill, *et al.*, 2007).

In order to create such a model Swart, *et al.* (2003a) assessed how the recharge mechanisms changed as mining progressed with a greater influence of artificial recharge during the mining period compared to the pre-mining period, Table 19.

Table 19: Importance of recharge mechanisms, after Swart *et al.* (2003a).

Recharge Mechanism	Pre-mining	Mining	Post Mining
Natural recharge form rainfall infiltration	Very important	Very important	Very important
Recharge from the Wonderfonteinpruit and dolomitic springs under normal rainfall events	Very important	Except for storm events, unimportant	Depends on future if pipeline
Recharge from the Wonderfonteinpruit and other drainage channels under extreme rainfall events	Unimportant	Significant but rare	Important
Leakage through dykes form adjacent compartments	Unimportant	Unimportant	Important
Recharge from mine dewatering operations	Nil	Boskop-Turffontein Eastern Gembokfontein	Nil

Storativity from different studies were compared based on different methods of estimation (Figure 35). The Director of Water Affairs (1960) as stated by Swart *et al.* (2003a) used geological sections through mine shafts and water budget studies to conclude that the storage of the dolomites decreases from a 15% estimate to a 1.5% estimate after the first 30m of the pre-mining water level.

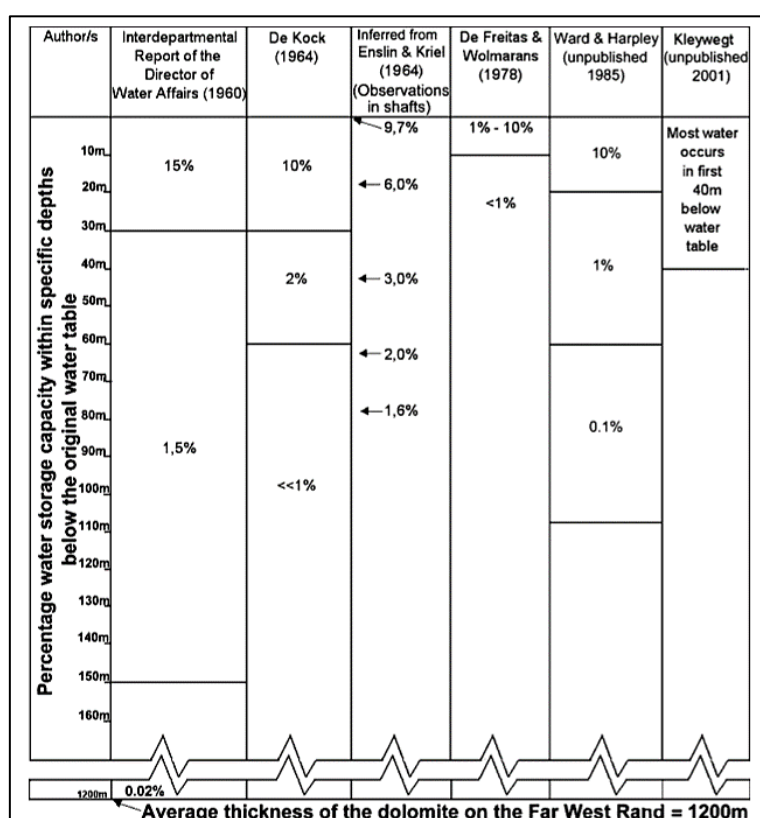


Figure 35: Estimations of the storage capacity of the FWR dolomite (Swart, *et al.*, 2003a).

Based on the following set of assumptions, Swart, *et al.* (2003a) comprise the geohydrological model (Dill, *et al.*, 2007):

- For simplicity, all pumping activities would cease simultaneously.
- Groundwater levels will recover to pre-mining levels.
- The cavernous zone within the dolomitic aquifer will play a significant role in storage once the water levels had recovered.
- The eyes / dolomitic springs will flow once again.

The model developed by Swart, *et al.* (2003a) represented each compartment's storage capacity as a "tank" with the series of tanks separated by the dykes representing the entire FWR system (Figure 36).

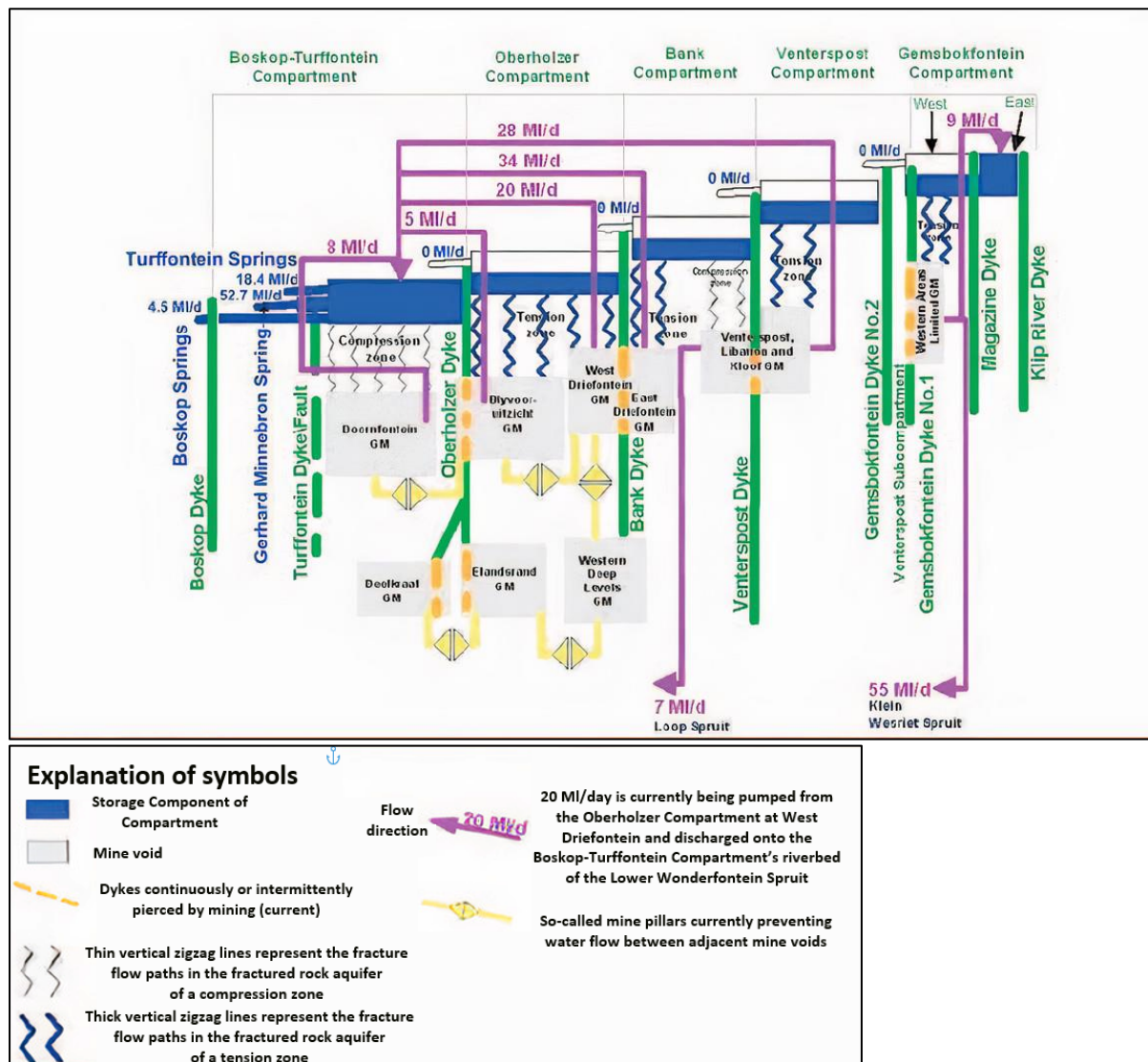


Figure 36: Schematic representation of the current geohydrology (Swart, *et al.*, 2003a).

The blue & white tanks represented dewatered compartments and sole blue coloured tanks representing non-dewatered compartments. A series of zig-zag lines represented the fractured dolomitic rock aquifer below the compartments that impart connected the dolomitic aquifer to the mine voids.

The IGF was calculated using Equation (5.1) for the downward leg through the fractured dolomite into the mine voids; the horizontal flow by means of mine voids through pierced dykes; and lastly the upward leg trough the fractured dolomite (Figure 37).

$$R_1 \gg Q = \frac{\Delta h}{\sum_{i=1}^n \frac{l_i}{k_i A_i}} = \frac{\Delta h}{\frac{l_1}{k_1 A_1} + \frac{l_2}{k_2 A_2} + \frac{l_3}{k_3 A_3}} \quad (5.1)$$

Where,

Table 20: Equation (5.1) abbreviations list.

Q	=	Total flux entering the dolomite aquifer from the mine void (Ml/day)
l_i	=	Flow path length across the porous media
A_i	=	Area of flow, (depends on mining area, area of dolomite aquifer affected by the mine void, etc.)
k_i	=	Permeability (m/s) of the fractured rock aquifer
R_1	=	Total Recharge to the upstream compartment

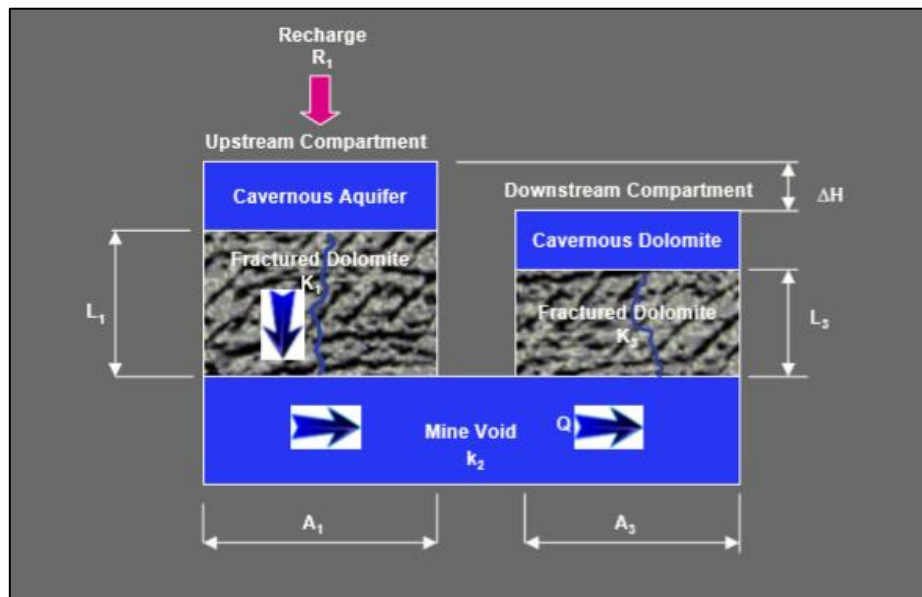


Figure 37: Calculation of IGF using Equation (1.9) as depicted by Dill, et al. (2007).

From their findings Swart *et al.* (2003a) noted:

- That all four dewatered compartments have a conjoined natural recharge of 133ML/d.
- Currently, the dolomitic aquifer is an estimated 1.14 ML (million litre) below its pre-mining storage capacity.
- In the case of a flooding event, the sinkholes within the FWR will provide major conduits for rapid recharge.
- Once the mine voids have filled, the recharge rate will be higher through the fractured rock section of the aquifer, thereafter the recharge rate will lower significantly through the rock mass of the dolomitic aquifer.

It was further noted that as long as the post-mining IGF is less than the compartment's total recharge, the relevant spring will start flowing again. An increasing IGF rate by means of the mine voids through the pieced dykes, would lead to (Dill, *et al.*, 2007; Swart, *et al.*, 2003a):

- A reduced flow rate at the relative compartments spring.
- A periodical drop in the recharge rate below that of the IGF flow rate, could lead to a temporary drop in groundwater levels.

5.2.5 Schrader *et al.* (2014a): “Determining hydraulic parameters of a karst aquifer using unique historical data from large-scale dewatering by deep-level mining – a case study from South Africa.”

Schrader *et al.* (2014a) utilized various sets of historical dewatering data from the deep-level gold mines in the FWR. This data, unlike normal pumping tests that are confined to near surface take place at the bottom of the thick aquifer. Four analytical methods designed for porous media were implemented to calculate the horizontal transmissivity and storage coefficients.

The first dataset used was the dewatering of the Bank Compartment. The data included the pumping rates from West-Driefontein mine as well as drawdown observations from 14 boreholes within a defined radius around the depression cone. The second dataset was the inrush event during 1968 that caused a water level drawdown up to 50m with eight boreholes within the defined radius of the depression cone. Figure 38 represents a magnified pumping test with a much larger drawdown than normally and the ‘*Big-Boy Fault*’ the pumping well.

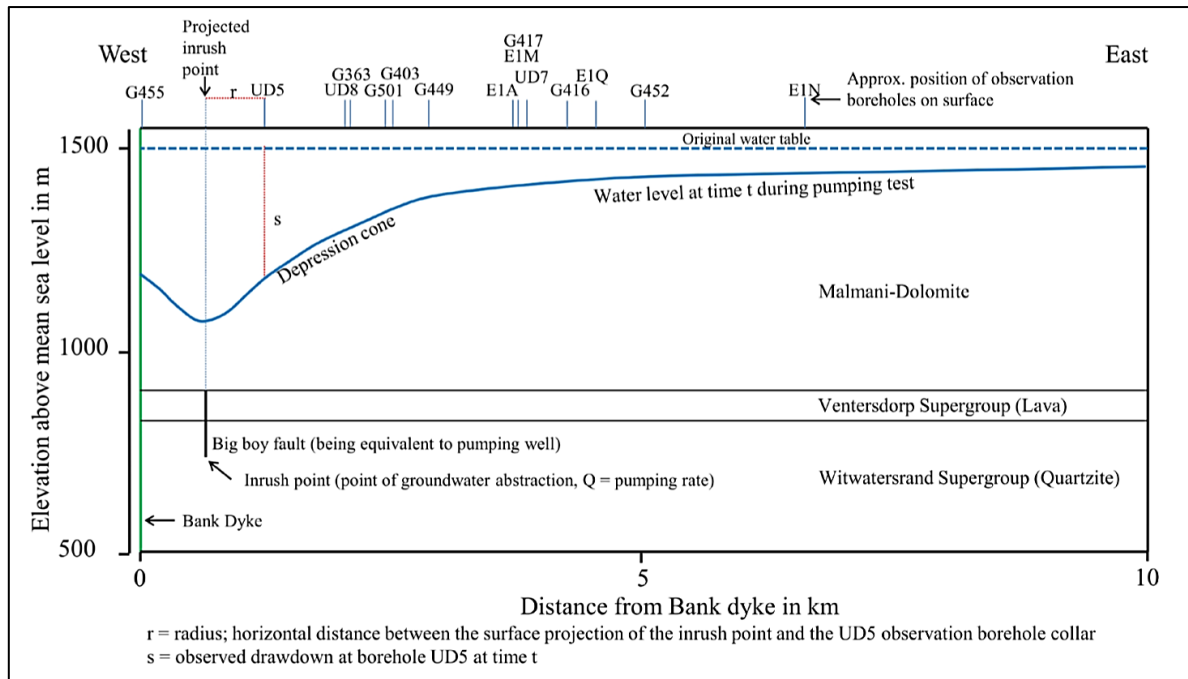


Figure 38: East-west cross-section showcasing the depression cone from the 1968 intrush, from Schrader *et al.* (2014a).

The third dataset included water level data from the steady-state dewatering during 1996 with a clear cone of depression evident in the groundwater contour map (Figure 29, Section 4.5.4). Three of the analytical methods used to evaluate the data were porous media-specific and included the well known 'Theis' method for the evaluation of transient state pumping tests as well as the 'Stallman' and 'Theim' methods. The fourth method was the analytical software tool MLU for Windows. A more detailed description of these methods is found within Schrader *et al.* (2014a).

The obtained storage coefficients were less reliable than the transmissivity values as the distance-dependency was non-satisfactory. From the obtained transmissivity results in Table 21, the values for individual boreholes exhibited high variability, as would be expected for a karstified aquifer. Large differences in the transmissivity value were due to the applied methods. The 'Theis' and 'Stallman' results for dataset one were significantly lower than that of the 'MLU' as these methods don't consider a variable pumping rate. For the second data set, the 'Theis' method was considerably lower than the 'MLU' and 'Stallman'. Although the aquifer was pumped at a stable rate, the 'Theis' method doesn't consider the influence of the no-flow boundary represented by the Bank dyke. The values for dataset two was 10 times higher on average than that of dataset one. The values from dataset two was higher and they represent the complete water-filled aquifer with all vertical zones saturated. The transmissivity of the dataset one only accounted for the aquifer beyond the first 200 meters that were

dewatered, therefore only represented the fractured and solid dolomitic zones of the aquifer (Schrader *et al.*, 2014a).

Table 21: Condensed overview of the calculated T&S values for each analytical method, after Schrader *et al.* (2014a).

Dataset: Type WL data	No. Obs. karst zone (mbs)	Transmissivity (T) (m ³ /s)				Storage (S) (%)		
		Geometric mean (min-max)				Geometric mean (min-max)		
		Theis	Stallman	MLU	Theim	Theis	Stallman	MLU
(1) Active dewatering n=161 194-863 mbs		50 (35-175)	73 (54-221)	285 (154-538)	N/A	0.58 (0.24-1.95)	0.88 (0.42-1.95)	1.12 (0.63-3.30)
(2) Inrush event 1968 n=8 53-863 mbs		1214 (618-2721)	2593 (1112-6143)	2468 (1216-5418)	N/A	0.36 (0.1-0.8)	0.65 (0.1-1.38)	0.70 (0.15-1.36)
(3) Steady state pumping n=161 194-863 mbs		N/A	N/A	N/A	40 (28-57)	N/A	N/A	N/A
All data		50-1214 (35-2721)	73-2593 (54-6143)	285-2468 (128-5418)	44 (28-57)	0.36-0.58 (0.1-1.95)	0.65-0.88 (0.1-2.62)	0.70-1.12 (0.15-3.18)
Limitations		a,b	a		b	a,b	b	
a -does not consider variable pumping rate b -does not consider flow boundary effects								

From the transmissivity results the hydraulic conductivity was calculated and plotted relative to identify vertical zones of karstification and porosity (Figure 39):

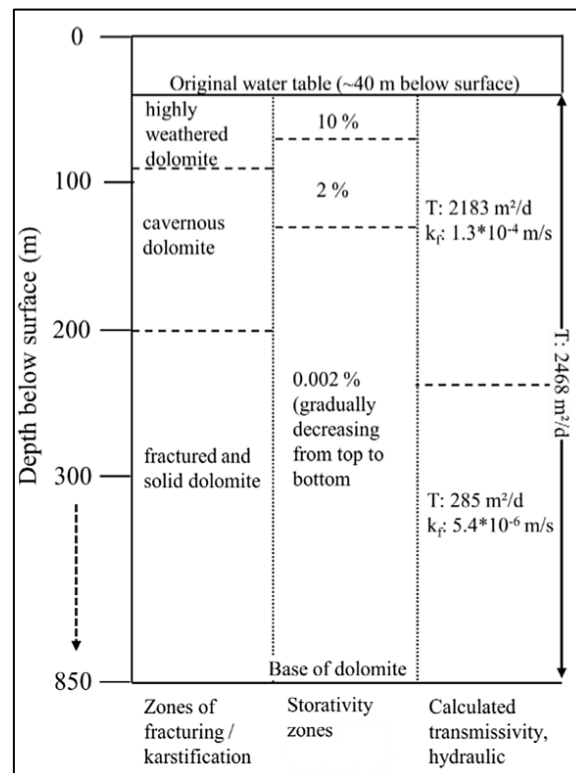


Figure 39: Transmissivity and hydraulic conductivity in relation to karstification, from Schrader *et al.* (2014a).

5.2.6 “Using impacts of deep-level mining to research karst hydrology— a Darcy-based approach to predict the future of dried-up dolomitic springs in the Far West Rand goldfield (South Africa).” Part 1: (Schrader *et al.*, 2014b) and Part 2: (Schrader *et al.*, 2014c).

This study comprised of two articles which focused on predicting a final post-mining groundwater level in the relevant dewatered aquifers of the FWR. The study used larger datasets than those previously applied, together with modified assumptions to strengthen the findings for the post-closure scenario (Schrader, *et al.*, 2014b). Similar to the previous study by Schrader, *et al.* (2014b), Swart, *et al.* (2003a) the state in which the different compartments were hydraulically connected due to mining activity was considered.

One of the major challenges was to predict the possible formation of the so called “mega-compartment” first mentioned by Jordaan, *et al.* (1960). This is a concept based on the fact that mining had pierced the watertight dykes, hydraulically connecting the different dolomitic compartments with one representative water level throughout (Schrader, *et al.*, 2014b; Schrader, *et al.*, 2014c).

To approach this problem, a conceptual model was created where the water table elevation will depend on a ratio between groundwater recharge in the form of water influx from the surface and the water loss due to IGF into downstream compartments, Figure 40.

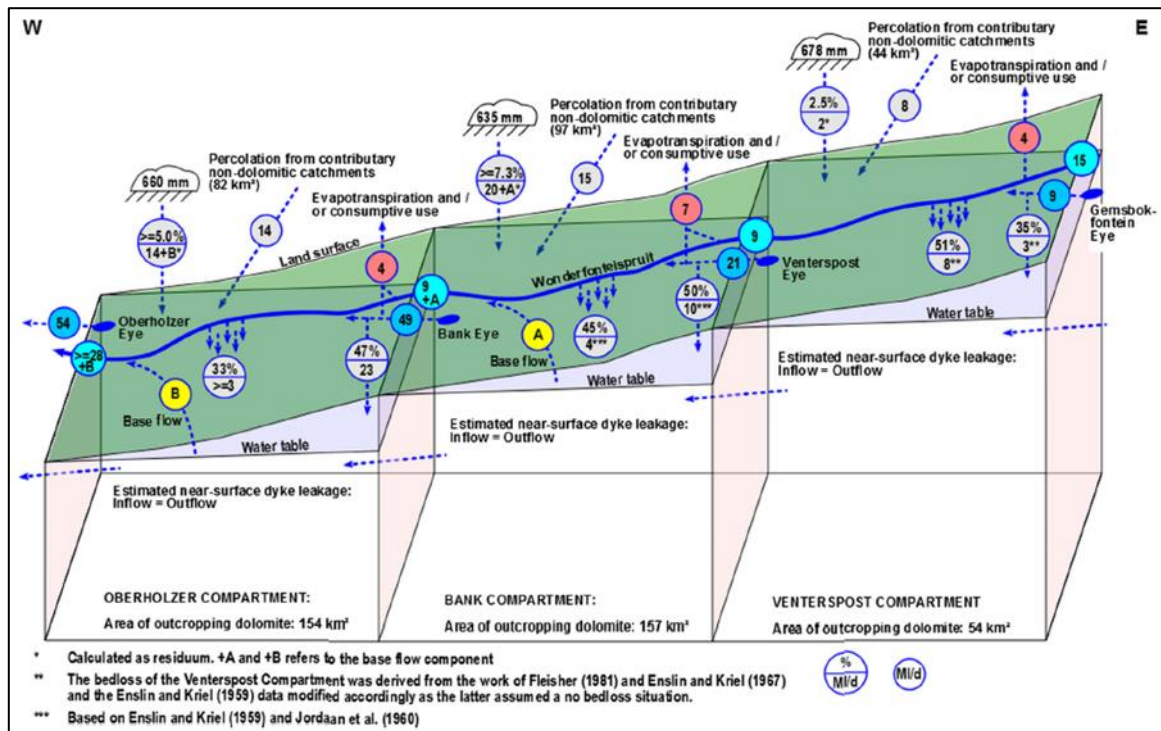


Figure 40: Conceptual model for the pre-mining water balance of the FWR (Schrader, *et al.*, 2014b).

Before mining commenced, natural recharge occurred from (Schrader, *et al.*, 2014b):

- Infiltrating rainfall on the dolomitic outcrop.
- Surface runoff and interflow intercepted by paleo karst structures.
- Stream loss through the riverbed of the WFS river, including recharge from the karst springs in upstream catchments.
- Surface runoff from non-dolomitic areas onto the dolomitic outcrops.
- Groundwater seepage through weathered near-surface dykes.

However, mining within the FWR affected the recharge of the dewatering compartments with (Schrader, *et al.*, 2014b):

- The formation of sinkholes and karst structures, increasing the surface runoff recharge.
- The diversion of the WFS river through a watertight pipeline, reducing the stream loss recharge.
- The drying up of the karst springs, reducing the recharge.
- Seepage losses through the upper-weathered dyke areas

Similar to Swart, *et al.* (2003a), IGF was calculated by subdividing the groundwater flux through each dyke into three legs (Schrader, *et al.*, 2014b; Schrader, *et al.*, 2014c):

- The downward leg, where groundwater flow down fault zones that connect the karstified dolomites with the mine void.
- The horizontal leg, where groundwater flows horizontally through the stoping area and other mine voids across the dykes.
- The upward leg, where groundwater is forced to flow upward from the mine voids through faults, fissures and fractures that connect the mine voids with overlaying karstified dolomites

The IGF was calculated for two scenarios. In the first scenario, the difference in water table elevation was estimated from pre-dewatering levels, represented by the respective karst springs. The second scenario represented the lowest possible water table elevation caused by the mega-compartment scenario, the water level was the difference between the pre-mining water level of the compartment and the Turffontein Eye, Table 22:

Table 22: Estimated IGF across mined through dykes, from Schrader *et al.* (2014c).

IGF across Dyke	Inter-compartmental groundwater flow (IGF) (Mℓ/d)		Original (pre-mining) spring flow (Mℓ/d)	IGF (% of original spring flow)	
	Scenario 1: Minimum IGF	Scenario 2: Maximum IGF		Scenario 1: Minimum IGF	Scenario 2: Maximum IGF
Venterspost	1.0	3.4	21	4.8%	16.2%
Bank	5.4	14.7	49	11.0%	30.0%
Oberholzer	0.7	0.7	54	1.3%	1.3%
Total	7.1	18.8	124	5.7%	15.2%

Because the IGF values are all well below that of the original spring flow values it is unlikely that a mega-compartment would form. The individual compartments will always receive continuous influx in the form of recharge that exceeds the water loss due to IGF through the pierced dykes as illustrated in Figure 41.

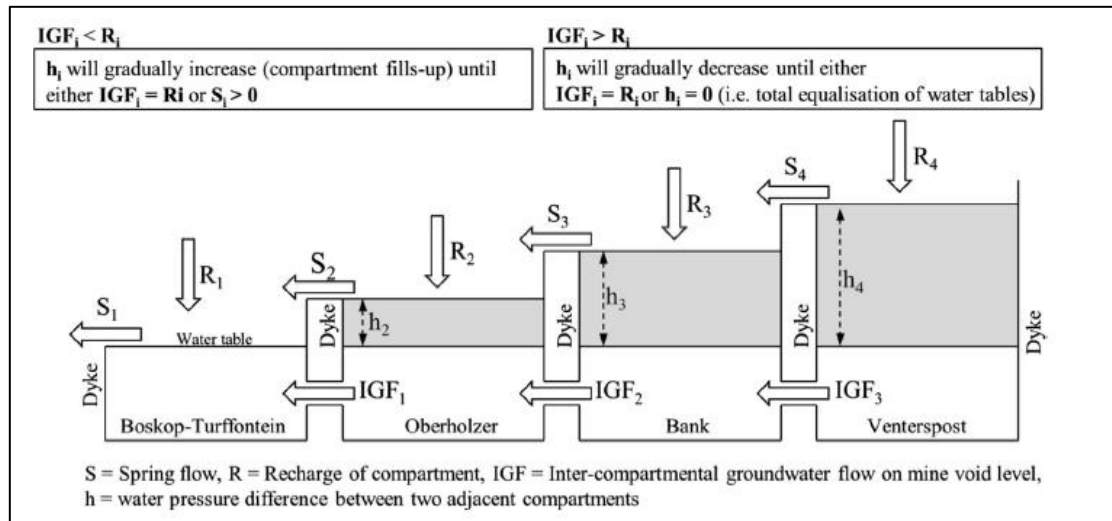


Figure 41: IGF-Recharge relationship conceptualized (Schrader, *et al.*, 2014b).

Although the IGF loss in the upstream compartment will lead to a decline in the associated spring flow, the spring flow of downstream compartments are simultaneously increased. With resulting spring flows nearly resembling that of pre-mining conditions (Figure 42) and the groundwater levels will be restored to near pre-mining levels (Schrader, *et al.*, 2014c).

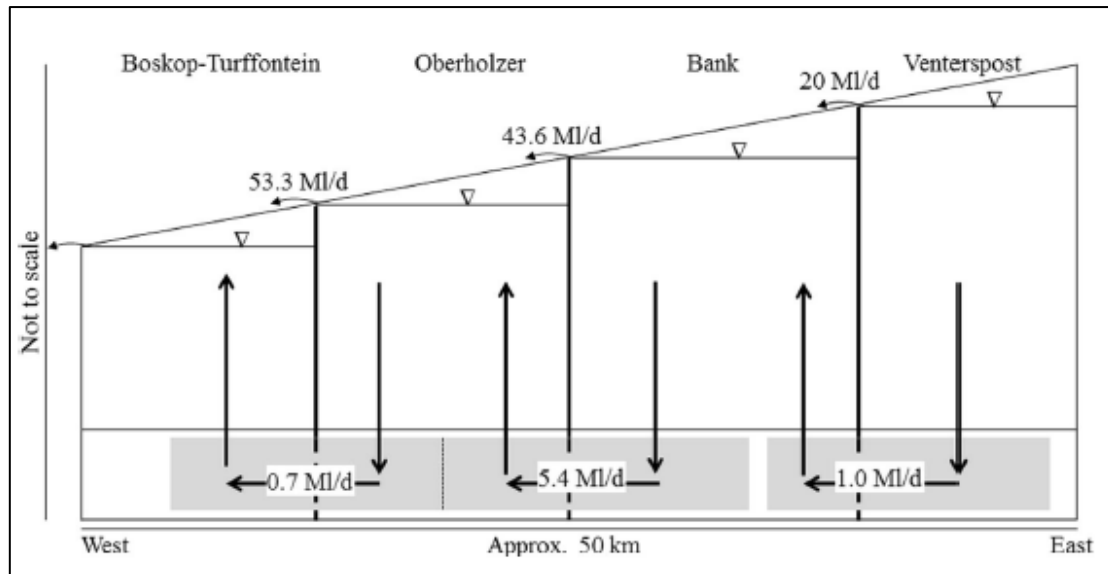


Figure 42: Predicted spring flow volumes from IGF calculations from Schrader *et al.* (2014c).

Schrader, *et al.* (2014b) did not factor in the recharge that will be attributed by the reactivated springs for the prediction presented in Figure (42). It is however considered relative to the recharge each compartment receives but subject to the environmental changes within the FWR. Factors such as sinkholes intercepting surface runoff and the diversion of the WFS river through a pipeline, all influence the post-mining spring flow. This, especially for the Bank eye, could result in the spring becoming non-perennial during the particular scenario where the WFS river remains diverted through a pipeline.

Concluding remarks by Schrader, *et al.* (2014c) stated that even though the application of Darcy's law (Equation 3.8) was based on the flow-through large conduits within the combined system. It is uncertain that the same principle is applicable for the combined karst and fractured aquifer system where different types of flow can occur. Various factors such as recharge and the size of the final mine voids will influence the final time required for rewatering, therefore it's important to be adequately prepared once mining activities cease.

5.2.7 Dennis & Dennis (2017): "Spring Flow Estimation After Mine Flooding in a Dolomitic Compartment."

Dennis & Dennis (2017) implemented the SVF method for a first-order estimation of predicting spring flow and restoration time of the dewatered Gemsbokfontein-West Compartment. The SVF method was based on a general groundwater budget. All the inflows and outflows were rendered to a change in head through the use of the aquifers specific yield and area based on the substituted groundwater budget Equation (3.10), Section 3.4.3.

The model was calibrated by using existing inflows for estimating conductance and calibrating the model's response to observed measurements by changing the specific yield and recharge parameters. Table 23 includes the known inflows and outflows together with the parameters that led to the best fit.

Table 23: SVF model inputs, after Dennis & Dennis (2017).

Inflows	(Mℓ/d)	Outflows	(Mℓ/d)	Parameter	Value
Zuurbekom Compartment	9	Cooke 4 Pumping	68	Study Area (km ²)	161
Gemsbok East Compartment	6	Cooke 3 Pumping	9	Specific Yield	0.0043
Leeuspruit Stream Loss	7.5	Gemsbokfontein Eye	0	Effective Recharge (%)	7.1
Rietspruit Stream Loss	5			Dolomite Storage (Mℓ/d)	23

Because of the model's simplicity, a few assumptions were made (Dennis & Dennis, 2017):

- All inflows were connected through faults to the mine voids.
- All dykes were treated as impermeable with inflows from other compartments controlled by a head-dependent conductance term.
- All recharge through rivers were controlled by a head-dependent conductance term.

The effective recharge observed corresponded well to values obtained by Enslin & Kriel (1968), 7.5% and the Groundwater Resources Assessment Phase II Project, 6.7% (WRC, 2005; Dennis & Dennis, 2017). The Results from the SVF model was compared to a numerical model created on MODFLOW (Figure 43). The implemented MODFLOW model accounted for the Gemsbokfontein-West, Gemsbokfontein-East and Zuurbekom Compartments with the transmissivity values from Table 24.

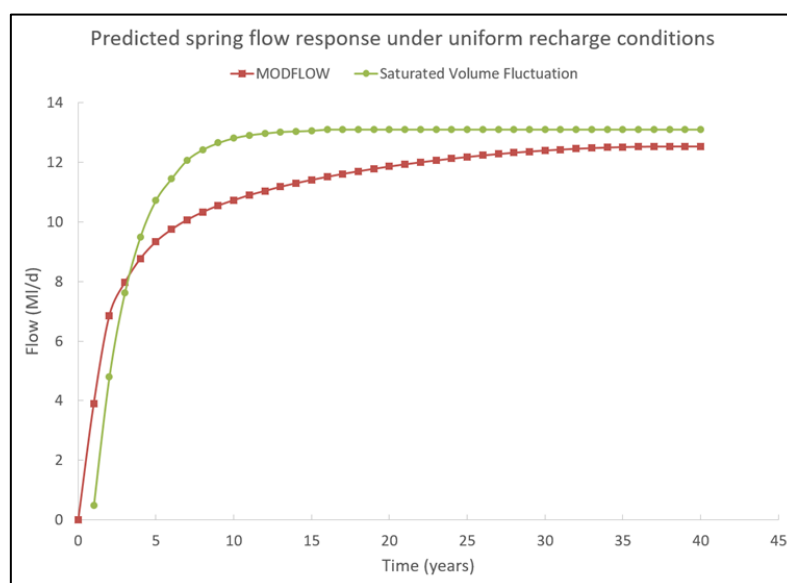


Figure 43: Spring flow comparison between the SVF and MODFLOW scenario, from Dennis & Dennis (2017).

Table 24: Transmissivity values used in MODFLOW model, from Dennis & Dennis (2017).

Stratigraphy/Geology	Transmissivity (m ² /d)
Transvaal Group	15
Weatherd dolomite	50
Deep lying dolomite	20
Gryke structures	500
Lineaments and Fault zones	100
Dykes	impermeable

The SVF model proved a quick and effective way of obtaining first-order spring flow estimates, predicting within a year's difference to the applied numerical model. Although steady-state estimates were off by tens of years the SVF model compared well considering its simplicity and assumptions (Dennis & Dennis, 2017).

5.2.8 Van Biljon (2017): “Managing groundwater inflow risks of the gold mines in the far West Rand Basin, South Africa.”

In the Ph.D. thesis of Van Biljon (2017), Van Biljon calculated recharge for the dolomitic compartments by using a pre-mining water budget. It was assumed that although dewatering has lowered the groundwater levels causing the springs to dry up, the natural recharge would remain similar. Furthermore, if it was assumed that no abstraction occurred within the compartment, the difference between the compartments inflow and outflow would reflect its natural recharge from rainfall. Van Biljon (2017) kept the area and rainfall constant where after Van Biljon altered the recharge percentage until it matched the recorded spring flow (Table 25).

Table 25: Pre-mining groundwater budget recharge calculations, after Van Biljon (2017).

Compartment	Original springflow (Mℓ/d)	Compartment Volume Balance			Area (m ²)	Rainfall (m)	Calculated Daily Recharge	
		In (Mℓ/d)	Out (Mℓ/d)	Balance (Mℓ/d)			(%)	(Mℓ/d)
Gemsbokfontein-West	8.64	-	8.64	8.64	54 601 082	0.646	9.00%	8.70
Venterspost	20.90	8.64	20.90	12.26	54 389 000	0.646	12.74%	12.26
Bank	49.10	20.90	49.10	28.20	156 690 000	0.646	10.17%	28.20
Oberholzer	54.10	49.10	54.10	5.00	153 845 000	0.646	1.84%	5.01
Turffontein	75.60	54.10	75.60	21.50	704 476 000	0.646	1.73%	21.57

5.2.9 (De Klerk, 2019): “An integrated rainfall-runoff water quality model in a mine impacted karst environment.”

De Klerk (2019) created a distributed rainfall-runoff model for the Wonderfonteinspruit Catchment. The Ph.D. thesis of De Klerk aimed at modelling rainfall-runoff and mass transport within the WFS by making use of the stormwater modelling software: EPA SWMM™.

De Klerk (2019) delineated the WFS according to the drainage network of the WFS river by means of the catchment delineation process in ArcGIS™ as opposed to the Department of Water Affairs (DWA) quaternary catchments. Within his delineation, De Klerk included the large Endoreic area identified by Middleton & Bailey (2008). Endoreic areas commonly referred to as pans are found throughout South Africa. Characterised as natural shallow depressions with no in- or outlet, these endoreic areas only receive inflow via precipitation with evaporation and seepage as the sole outflows, Figure 44.

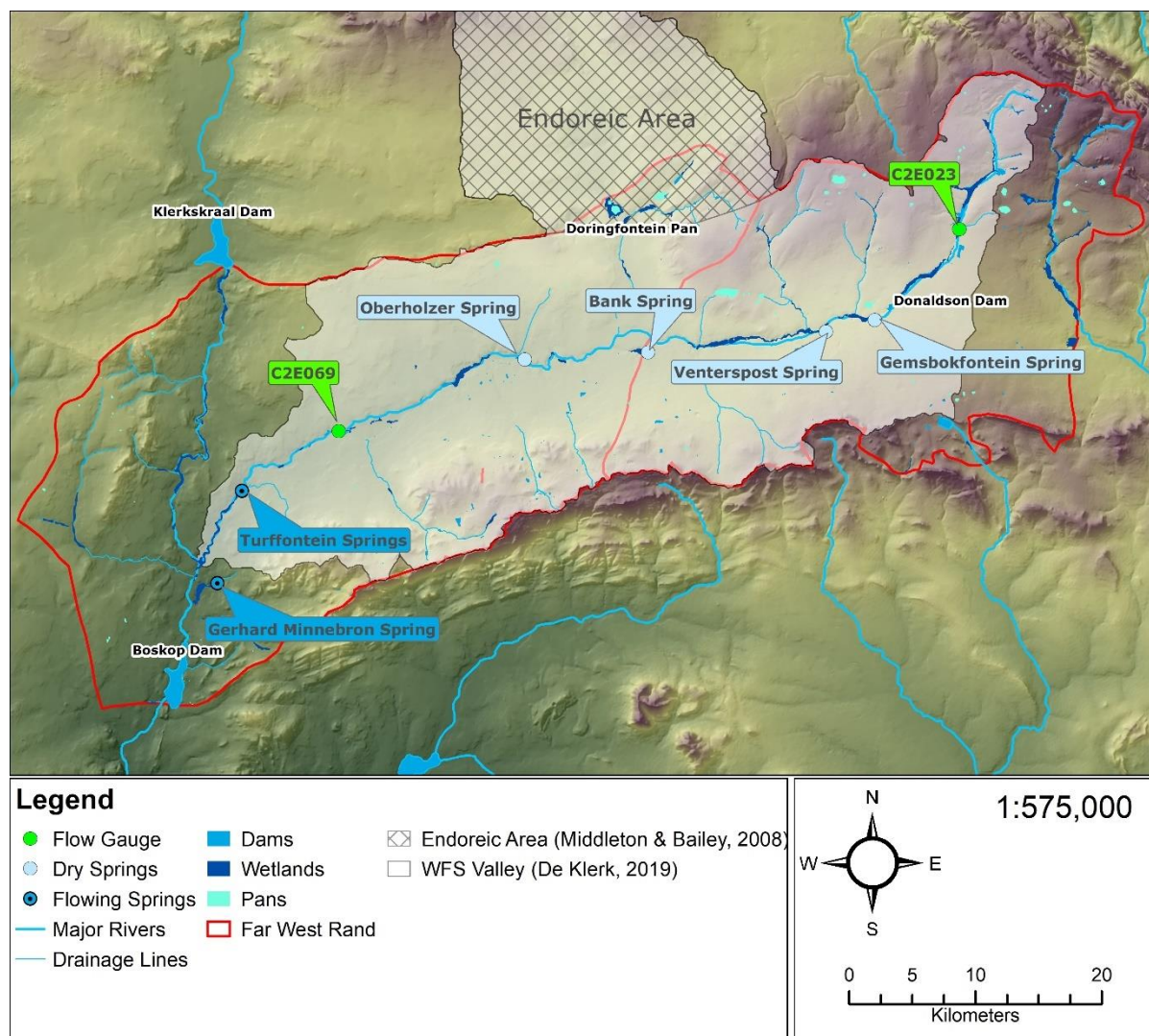


Figure 44: Wonderfonteinspruit Catchment delineation by (De Klerk, 2019).

For his thesis, De Klerk did an in-depth review on the hydrological workings of the FWR. He further sub-delineated the WFS into an upper and a lower part. The sub-division followed the general consensus previously followed by Coetzee, *et al.* (2006); Krige (2006); Usher & Scott (2001); Swart, *et al.* (2003b) where Donaldson Dam serves as the dividing term between the two sub-divisions. The Upper-Wonderfontein mainly comprises of the headwaters of the WFS as well as a low abundance of dolomite resulting in few sinkholes and caves. Although other sources of ingress such as mine shafts and quarriers still influence the natural runoff. The Lower-Wonderfontein makes up the largest area of the two sub-divisions, spanning across the majority of the dolomitic compartments, including the identified endoreic area. With the large number of sinkholes present, De Klerk used a Sinkhole Modification Value calculated from the sinkhole's area to determine the infiltration rates.

De Klerk (2019) highlighted challenges within the calibration of his SWMM model development with factors such as the overland flow, seepage drainage as well as sinkhole infiltration sensitive to change during calibration. After extensive calibration De Klerk ran two recovering compartment scenarios:

- The "Mega-compartment" scenario, consisting of a single mega-compartment opposed to multiple aquifer compartments separated by impermeable dykes. This scenario first introduced by Jordaan, *et al.* (1960) as described by Wolmarans (1984) is mainly due to the mine voids found within these dykes hydraulically connecting the different compartments (Usher & Scott, 2001; De Klerk, 2019).
- The "Recovered compartment" scenario, as discussed in Section 5.2.4 describes that the groundwater level within each compartment will return to its initial water level and the dry springs will start flowing again. This spring flow is estimated to be lower than initial volumes with the possibility of no spring flow during low rainfall periods (Swart, *et al.*, 2003a; Dill, *et al.*, 2007; Schrader, *et al.*, 2014c).

The results from De Klerk's scenarios confirmed that the Upper-WFS is a mainly rainfall driven sub-catchment. De Klerk (2019) also noted that no spring flow occurred for the mega-compartment scenario. The flow measured at the outflow point of the Lower-WFS flow gauge C2H069 indicated a large difference between the two scenarios. For the recovered compartment scenario, it was found that even though the springs would reactivate, three out of the four springs' flows would be subjected to the impact of the sinkholes within the streambed.

5.2.10 APLIS application on the FWR

The method developed by Andreo *et al.* (2008) was implemented and adapted for estimating recharge in the FWR. The APLIS method was applied to derive an accurate recharge range for each compartment whilst validating the effectiveness of a GIS-based recharge method in Southern Africa.

GIS overlay maps were created according to the variable ranking systems mentioned in Table 4 within Section 3.3.1. The variable ranking system was adapted according to the topography, lithology and soil types found within the FWR (Table 26).

Table 26: APLIS implementation variable ranking system.

A		P		L	
Altitude (m)	Rating	Slope (%)	Rating	Lithology	Rating
(1200-1500)	6	(<3)	10	Dolomite (Karstified)	7
(1500-1800)	5	(3-5)	9	Gravels and sands	4
		(5-10)	8	Plutonic and metamorphic rocks	2
		(10-15)	7	Shales, silts and clays	1
		(15-20)	6		
		(20-30)	5		
		(30-45)	4		
I		S			
Infiltration landforms	Rating	Soils		Rating	
Sinkholes, Dykes and	10	Leptosols		10	
Dolomite	5	Leptosols to Calcareous regosols		9	
Non-dolomite	1	Cambisols		6	

The highveld of the FWR is situated at a high elevation between 1369 and 1792 mamsl, therefore only two of the proposed ten altitude variables could be implemented. Slopes within the study area rarely exceed 10% but do occasionally reach a percentage of 45% with seven of the possible slope variables implemented. For the lithology, shales together with silts and clays were given a value of one, metamorphic rocks were valued at two with gravels and sands valued at four as proposed by Andreo *et al.* (2008). The dolomitic strata depended on the degree of karstification as well as fracture occurrence. Compared to the aquifer systems on which the APLIS method was based by Andreo *et al.* (2008), the dolomites of the FWR fitted a value of seven.

Areas dominated by fault lines, sinkholes and other intrusion structures were assumed as abundant infiltration landforms, assigned a value of ten. The dolomitic strata, considered moderate infiltration landforms assigned a value of five with other scarce infiltration landforms assigned a value of one. Lastly, the soil types were classified according to the World

Reference Base soil groups (Table 27) based on the soil characteristics map, Figure 4 in Section 2.4.

Table 27: Implemented APLIS soil type classification (Andreo, *et al.*, 2008; Fey, 2010).

SORTER soil association map for the FWR	Correlation of South African soil groups with World Reference Base soil groups (Fey, 2010)	APLIS (Andreo, <i>et al.</i> , 2008)
Red and yellow, massive or weakly structured soils with low to medium base status (association of well drained Ferralsols, Acrisols and Lixisols and one or more of Regosols, Leptosols, Calcisols and Durisols)	Humic	Cambisols
Red and yellow, massive or weakly structured soils with low to medium base status (association of well drained Ferralsols, Acrisols and Lixisols)	Humic	Cambisols
Red, massive or weakly structured soils with high base status (association of well drained Lixisols, Cambisols, Luvisols)	Humic	Cambisols
Red, yellow and greyish soils with low to medium base status (association of Ferralsols, Acrisols, Lixisols and Plinthosols. In addition, other soils with plinthic and gleyic properties may also be present)	Humic	Cambisols
Rock with limited soils (association of Leptosols, Regosols, Durisols, Calcisols and Plinthosols)	Lithic	Leptosols
Soils with minimal development, usually shallow on hard or weathering rock, with or without intermittent diverse soils. (association of Leptosols, Regosols, Calcisols and Durisols)	Lithic	Leptosols to Calcareous regosols

The correction factor, Fh distinguished between the aquifer and non-aquifer properties (Figure 45).

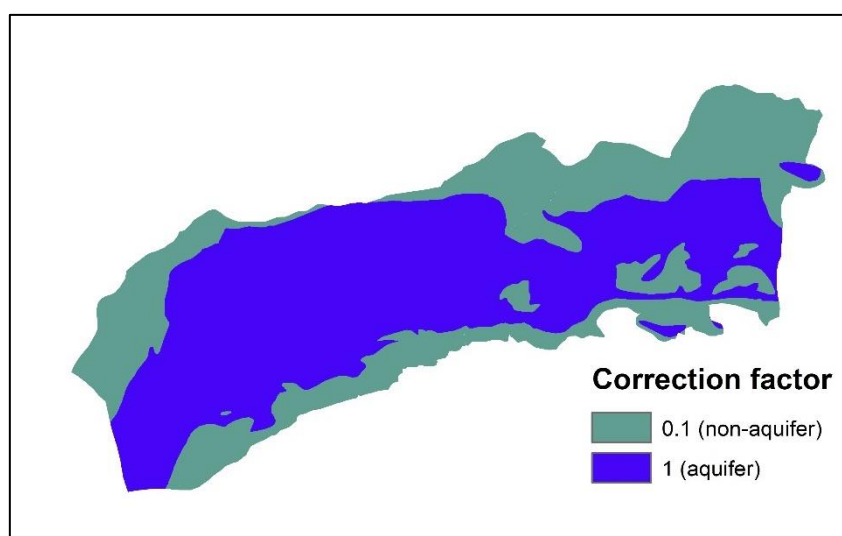


Figure 45: Implemented APLIS correction factor.

The resulting maps seen in Figure 46 together with the correction factor were overlaid using the APLIS mathematical expression (Equation 3.2), to obtain uncalibrated autogenic groundwater recharge values for the FWR. This recharge estimation was further calibrated against the known recharge values for the Gemsbokfontein West Compartment (Figure 47).

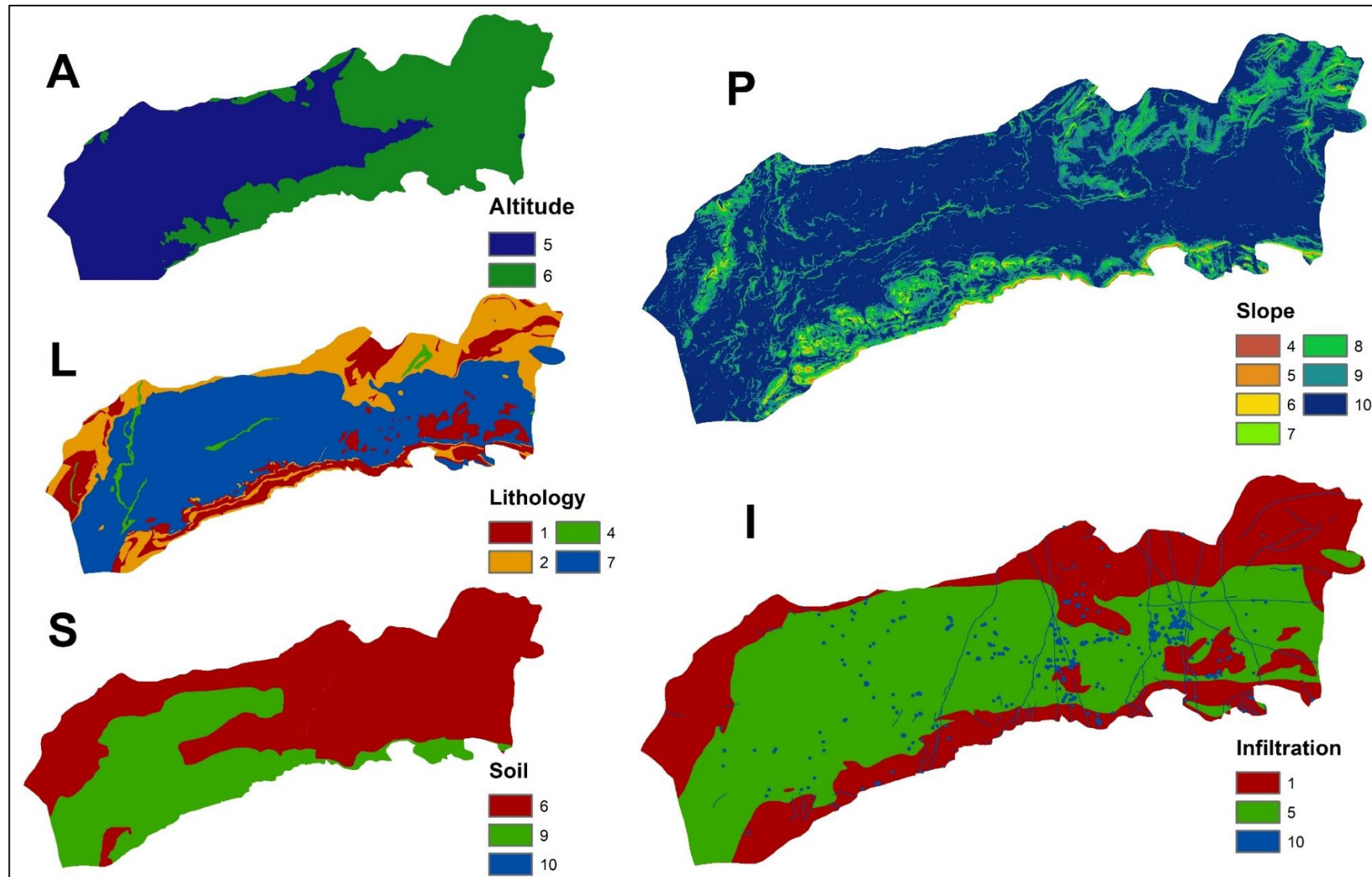


Figure 46: Implemented APLIS overlay maps.

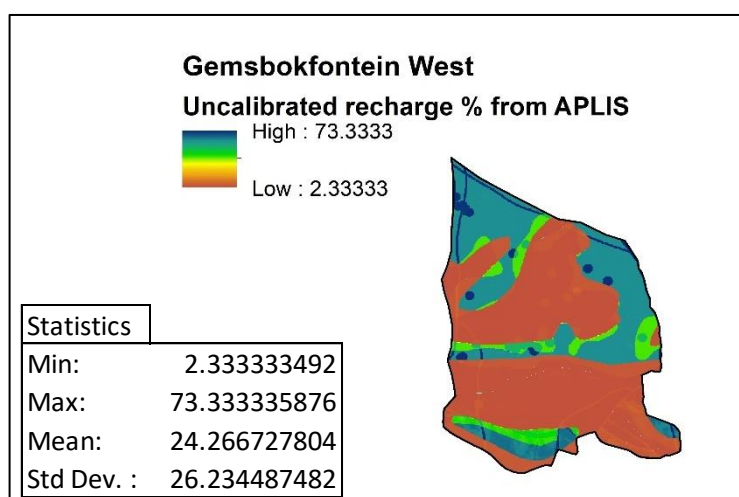


Figure 47: Uncalibrated APLIS output for the Gemsbokfontein West Compartment.

The uncalibrated recharge estimation for APLIS obtained values ranging between 2.3% and 73.3%. These values are quite high as the method estimates the percentage autogenic recharge resulting from direct precipitation and doesn't account for any evaporation and evapotranspiration losses as well as impermeable man-made surfaces as mentioned in Section 3.3.1. To account for these and other possible losses the Gemsbokfontein Compartment was calibrated against its measured recharge values to obtain a calibration coefficient to be implemented across the FWR. The Gemsbokfontein Compartment was chosen for the calibration as previous recharge estimations by Dennis & Dennis (2017) were obtained by implementing the SVF method discussed in Section 5.2.7. An effective recharge of 7.1% obtained by Dennis & Dennis (2017) corresponded well with an estimation of 7.5% done by Enslin & Kriel (1968) as well as the 6.8% obtained by the GRAII implementation done by WRC (2005) as mentioned by Dennis & Dennis (2017).

The effective recharge of 7.1% was divided by a mean 24.26% obtained from APLIS for Gemsbokfontein West to obtain a calibration coefficient of 0.292. This calibration coefficient was implemented across the FWR to obtain the final effective recharge estimation as seen in Figure 48:

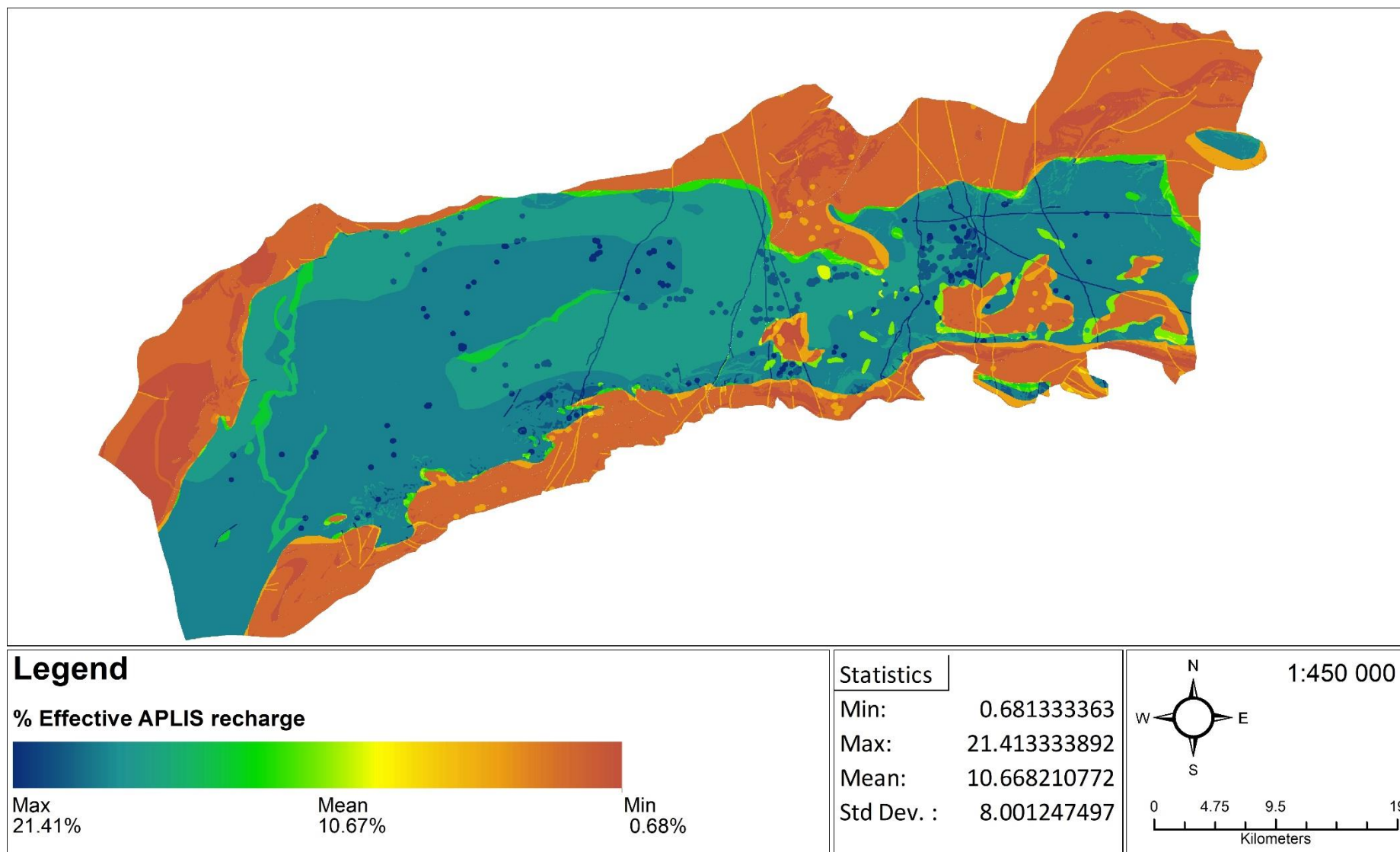


Figure 48: Calibrated APLIS output for the study area.

The results from the calibrated APLIS effective recharge indicate a strong connection between recharge and the dolomitic geology and associated features with the maximum autogenic recharge occurring at fault lines, sinkholes and heavy karstified terrain with the minimum autogenic recharge occurring over neighbouring non-dolomitic rock as seen in Figure 49:

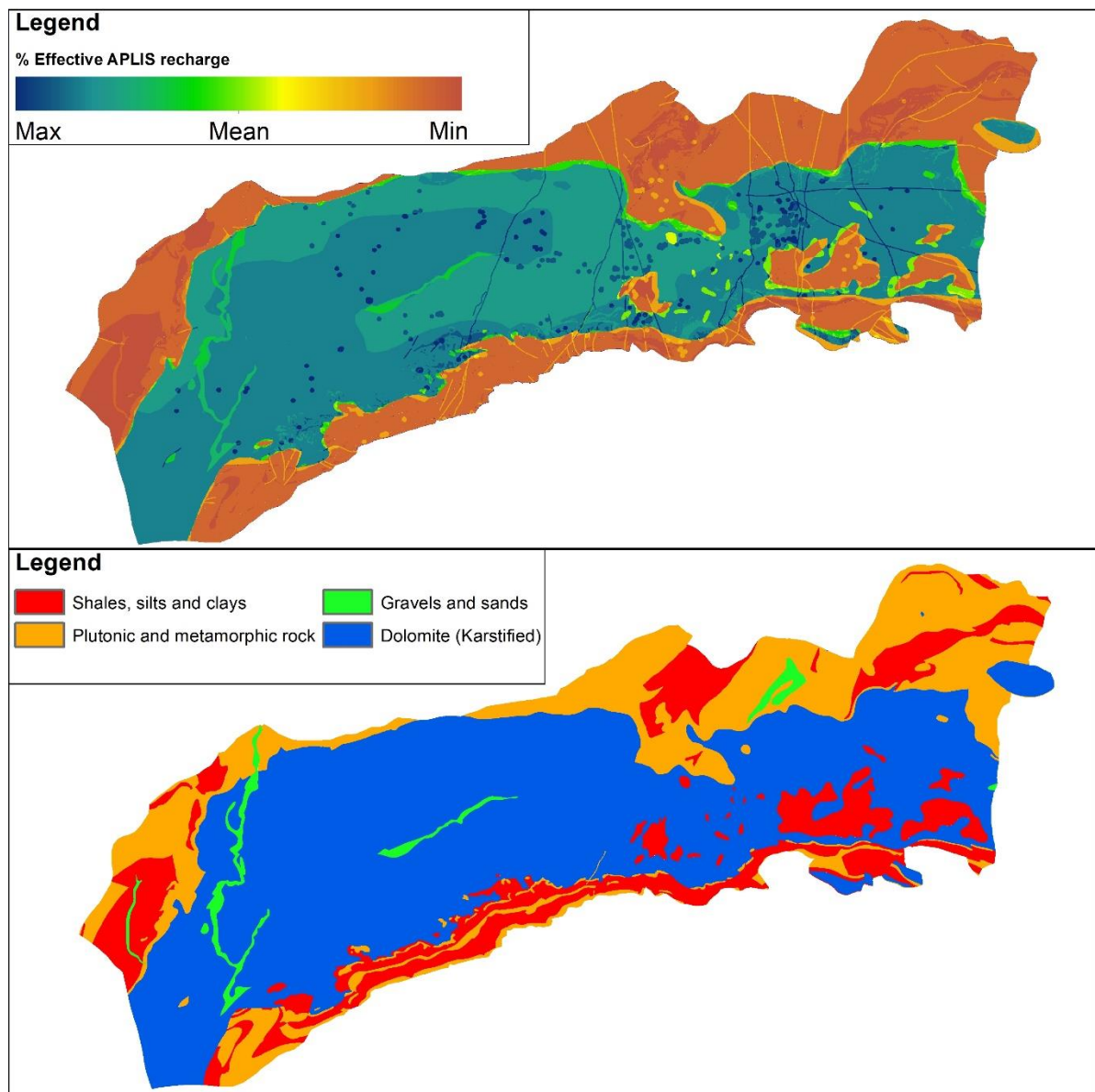


Figure 49: APLIS effective recharge correlation to dolomitic geology.

The effective recharge obtained from the APLIS overlay technique for each dolomitic compartment is summarized in Table 28.

Table 28: APLIS Compartmental effective recharge results.

Compartment	Recharge Max %	Recharge Min %	Recharge Mean %
Zuurbekom	20.44%	0.71%	7.25%
Gemsbokfontein- West	21.41%	0.68%	7.09%
Venterspost	20.44%	0.68%	11.10%
Bank	21.41%	0.78%	10.09%
Oberholzer	21.41%	0.74%	9.59%
Turffontein	21.41%	0.68%	12.89%

5.2.11 Estimated recharge values obtained by Vegter (1995)

As mentioned in Section 3.4.1, Vegter (1995) composed a national groundwater recharge map of South Africa by means of the ACRU model. The model was run for a total of 712 zones, each reflecting a different homogeneous rainfall response into which the country has subdivided.

The national recharge map developed by Vegter (1995) was based on actual annual precipitation per quaternary catchment. From Figure 12 (Section 3.4.1) recharge values for the majority of the FWR were estimated at 95 mm/a except for the Turffontein Compartment which varies between 65 mm/a and 95 mm/a. Based on the WR2012 mean annual precipitation values obtained from Figure 6 (Section 2.7), the approximate recharge % can be assigned to the estimated values by Vegter (1995), Figure 50.

Based on the groundwater recharge map composed by Vegter (1995), recharge varies between 7.5% and 16.46% based on the spatial variability of the mean annual precipitation from WR2012 (2012). The average recharge percentage for each of the compartments of the FWR based on Figure 50 is included in Table 29.

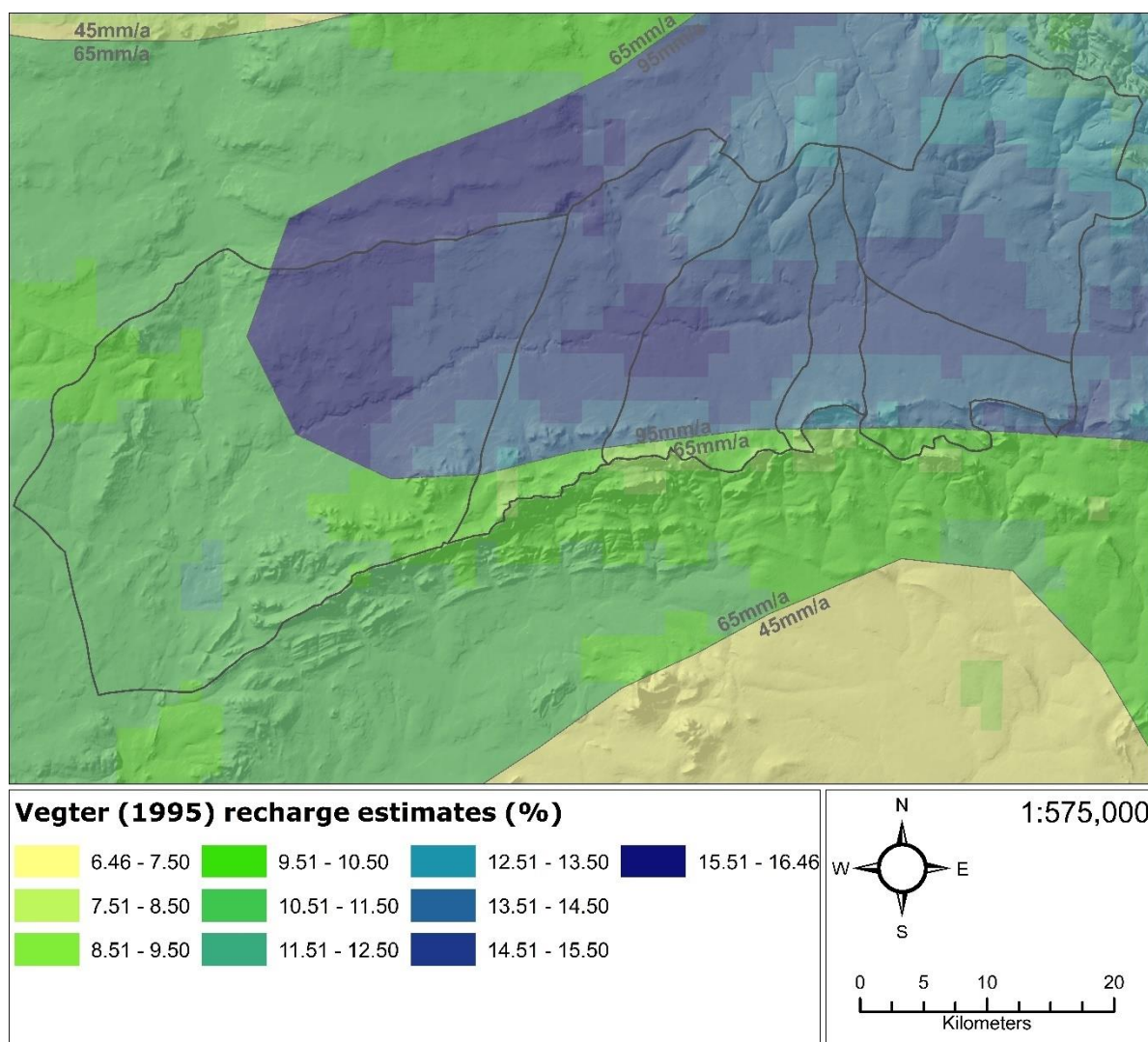


Figure 50: Vegter (1995) recharge % estimates.

Table 29: Vegter (1995) compartment groundwater recharge estimates.

Compartment	Vegter (1995) recharge % estimates			Mean annual precipitation (mm)
	Minimum	Maximum	Average	
Zuurbekom	11.7	15.3	13.8	693.5
Gemsbokfontein	9.2	15.2	13.4	661.9
Venterspost	9.1	15.1	13.8	671.6
Bank	8.8	16.5	13.7	657.0
Oberholzer	8.9	15.8	13.5	651.1
Turffontein	9.5	16.2	12.6	614.2

5.2.12 The CMB method to estimate recharge in the FWR

DWAF (2006) implemented the CMB method in a bid to estimate recharge within different homogeneous rainfall zones in South Africa for the creation of a national scale map of groundwater recharge, the Groundwater Resource Assessment II (GRAII). The resulting CMB values were combined with applicable GIS filters including the groundwater table depth, soil type, rainfall season, geology, landcover and slope. In addition, empirical rainfall/recharge relationships were assessed with a cross-calibration of the results to field measurements as well as detailed catchment values. Figure 51 showcases the calibrated recharge percentage obtained from the GRAII for South Africa:

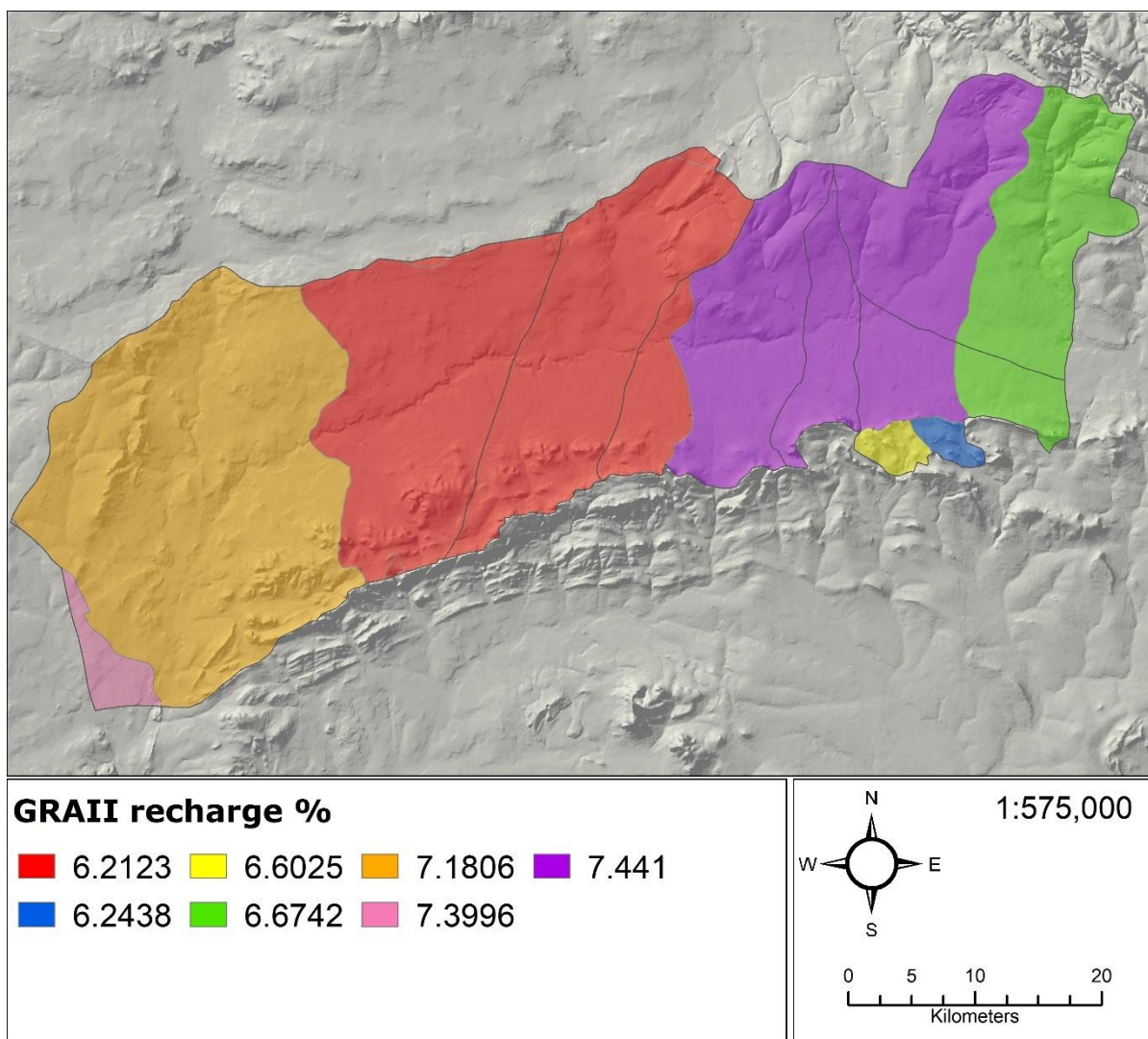


Figure 51: GIS calibrated recharge % obtained from GRAII.

5.2.13 Findings summary

The APLIS results were compared to the methods and recharge findings previously discussed in Section 5.2 in Table 30:

Table 30: APLIS results comparison.

Recharge (%) Study	Dolomitic Compartments					
	Zuurbekom	*Gemsbokfontein	Venterspost	Bank	Oberholzer	Turffontein
Enslin & Kriel (1968)	13	7.5	8.5	5.8	3.6	-
Fleisher (1981)	16.8	12.8	27	24	-	-
Wolmarans (1984)	15	5.3	20	16.3	18.3	5.6
Foster (1988)	-	-	54.6	27.3	12.9	-
Bredenkamp (1993)	15.8	27	-	-	-	-
Dennis & Dennis (2017)	-	7.1	-	-	-	-
van Biljon (2017)	-	9	12.74	10.17	1.84	1.73
Vegter (1995)	11.7 - 15.3	9.2 - 15.2	9.1 - 15.1	8.8 - 16.5	8.9 - 15.8	9.5 - 16.2
**GRA II DWAf (2006)	8.87	8.91	8.09	7.48	7.04	6.77
Average	13.9	11.3	20.7	15.0	9.5	6.7
APLIS	7.25	7.09	11.1	10.09	9.59	12.89
*Most estimates were done for the Gemsbokfontein West Compartment						
**For 656mm annual rainfall						

The results obtained from the APLIS recharge estimation were overall lower than that of the average recharge percentages for each compartment with exceptions for the Venterspost- and Turffontein Compartments. The results for the Zuurbekom-, Gemsbokfontein-, Oberholzer- and Turffontein Compartments closely resembled that of the recharge estimation by the GRAII study mentioned in Section 5.2.12. Moreover, the results for the Venterspost- and Bank Compartments resemble that obtained by calculations done by Van Biljon (2017), Section 5.2.8.

Due to the many uncertainties and predictions as to the average dewatering rate that occurred within each dolomitic compartment during recent years a summary of the most applicable values were chosen for use within the study (Table 31).

Table 31: Pumping rates for the applicable dolomitic compartments.

Compartment	Pumping (M ³ /d)	Additional comments
Gemsbokfontein West	77	From (Dennis & Dennis, 2017).
Venterspost	28	From (Swart, <i>et al.</i> , 2003a).
Bank	41 or	From (Swart, <i>et al.</i> , 2003a).
	69	From (Usher & Scott, 2001) (Schrader <i>et al.</i> , 2014b)
Oberholzer	25	From (Swart, <i>et al.</i> , 2003a).

For the Gemsbokfontein West compartment, the dewatering rate is unclear as to whether it was established during steady-state condition or as a result of over-pumping due to the large difference between the original spring flow of 8.6 and the pumping volume.

5.3 Model selection

The method previously applied to the Gemsbokfontein-West Compartment for recharge estimations by Dennis & Dennis (2017) will be used to predict groundwater level recovery within the dolomitic compartments of the FWR. Due to its low input data requirements and simple analytical solution, the SVF method's water budget methodology lumps parameters together. This enables simple calibration based on groundwater level fluctuations within monitoring boreholes. With each dolomitic compartment being regarded as individual systems together with the overall lack in distributed data, the lumped parameter approach used by the SVF method is best suited in this study. The method can also be easily adapted to simulate groundwater level recovery by keeping the pre-defined parameters constant throughout.

From the findings discussed in Section 5.2, it's evident that the creation and calibration of a model using pre-mining data results in uncertainties. Using pre-mining data leads to assumptions as to how mining activities impacted the aquifer system. The implementation of the SVF method will be based on the most recent conditions (as per data availability) whilst dewatering activities are ongoing. This not only takes into account the ongoing dewatering by the mines but also how the mining affected the aquifer with the growing development of karst features and mine voids.

5.4 Time dependent data

5.4.1 Monitoring Boreholes

To successfully calibrate the water budget model, monitoring boreholes within each dolomitic compartment were required. A total of 1298 boreholes were obtained from the National

Groundwater Archive (NGA) within the FWR. An initial analysis found that only 7% of all the boreholes had monthly measured groundwater level data after the year 2000 (Figure 52).

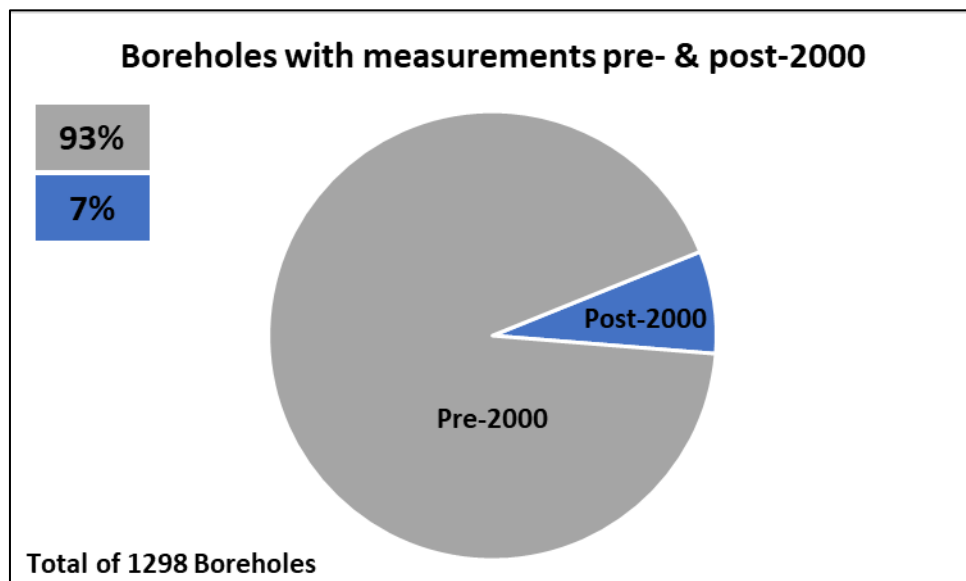


Figure 52: Boreholes with measurements pre- & post-2000.

Of the remaining boreholes that contained groundwater level data from the year 2000 onwards, 73% were of use. The other 27% of boreholes only contained a single month or zero monthly measurements, as seen in Figure 53:

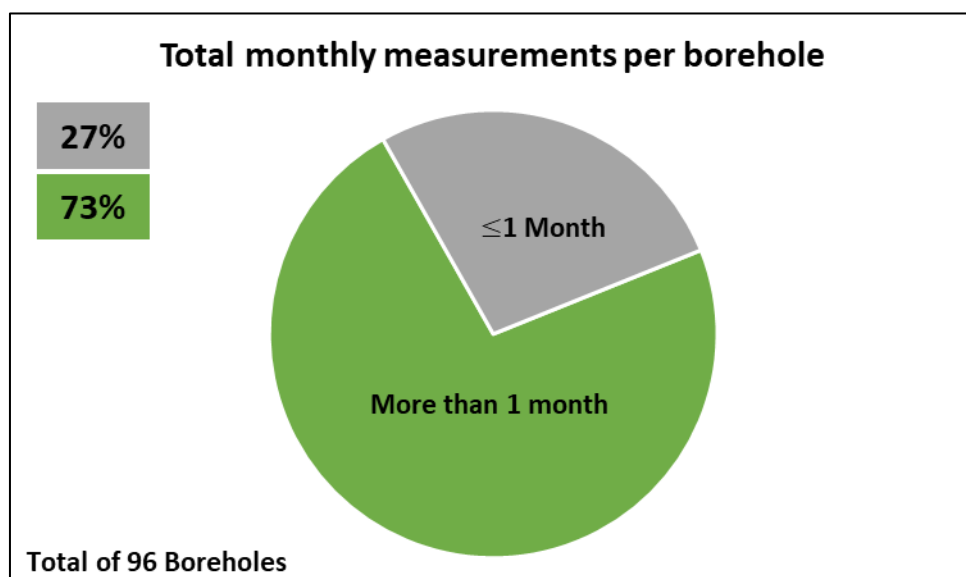


Figure 53: Total post-2000 monthly measurements.

Ideal monitoring boreholes were not only unaffected by the dewatering that takes place within each compartment but also had available overlapping monitoring data for calibration. A total of 21 monitoring boreholes across all six compartments (Figure 54) were eligible. Each of these

monitoring boreholes had a near-constant groundwater level unaffected by dewatering and three boreholes are located within each dolomitic compartment.

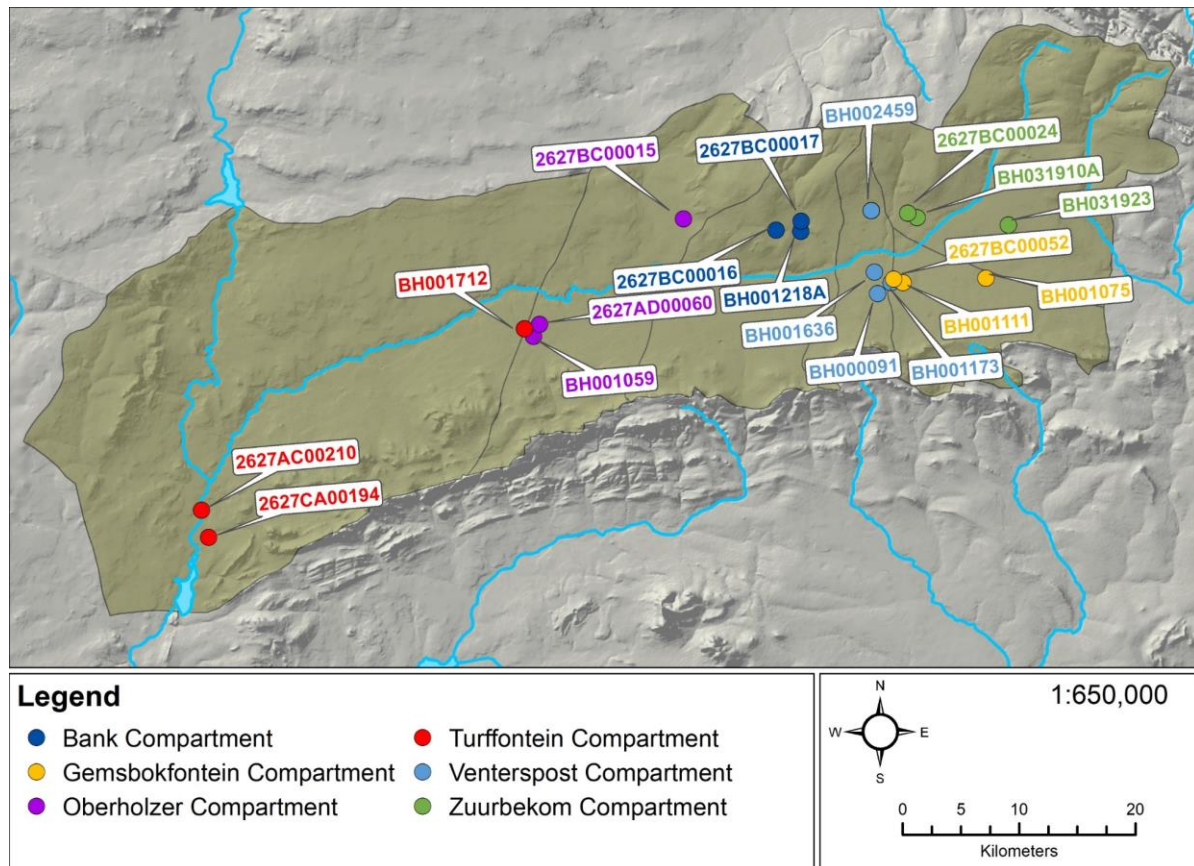


Figure 54: Shortlisted monitoring boreholes.

Figure 55 indicates the close to constant groundwater levels from BH001218A required for steady-state calibration. This is a characteristic of a borehole mainly unaffected by dewatering against a monitoring borehole (BH487) situated within the cone of depression formed by the dewatering of the Bank Compartment. Although a steady drop in the groundwater level in borehole BH001218A is witnessed, the drop is less significant when compared to that of BH487 which is closer to the point where dewatering of the Bank Compartment took place (Figure 56).

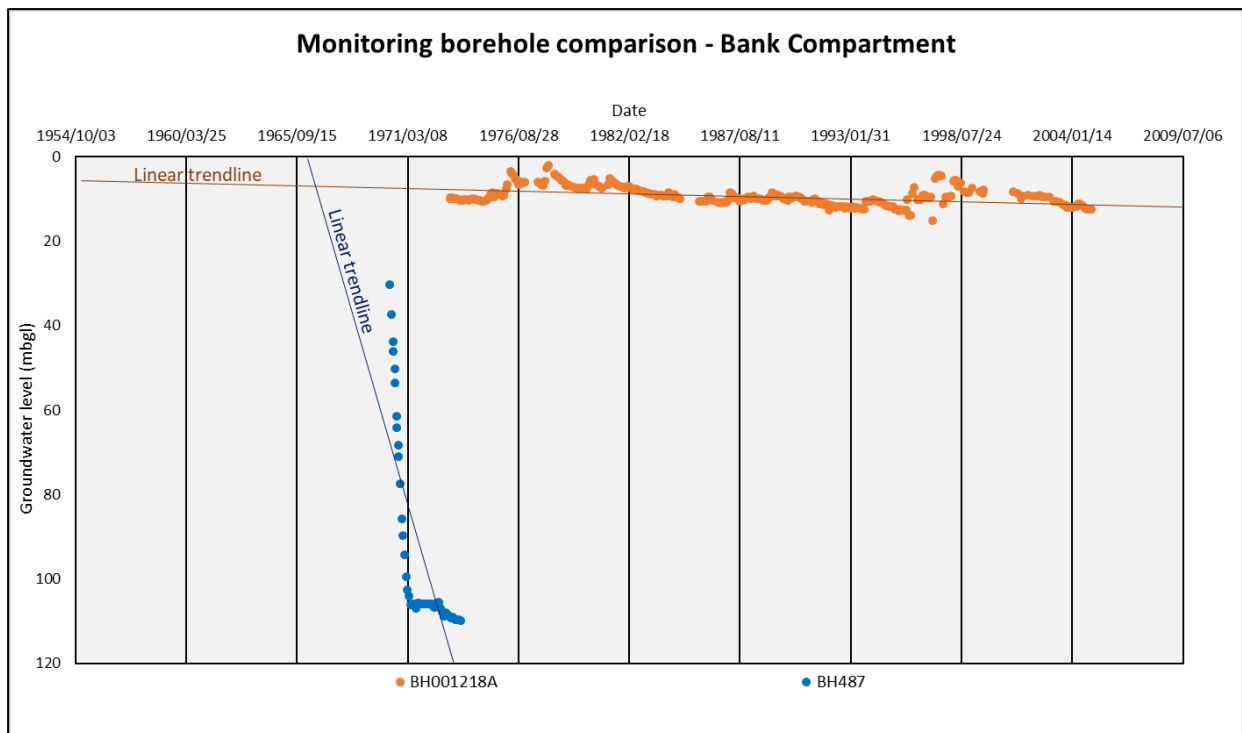


Figure 55: Monitoring borehole comparison - Bank Compartment.

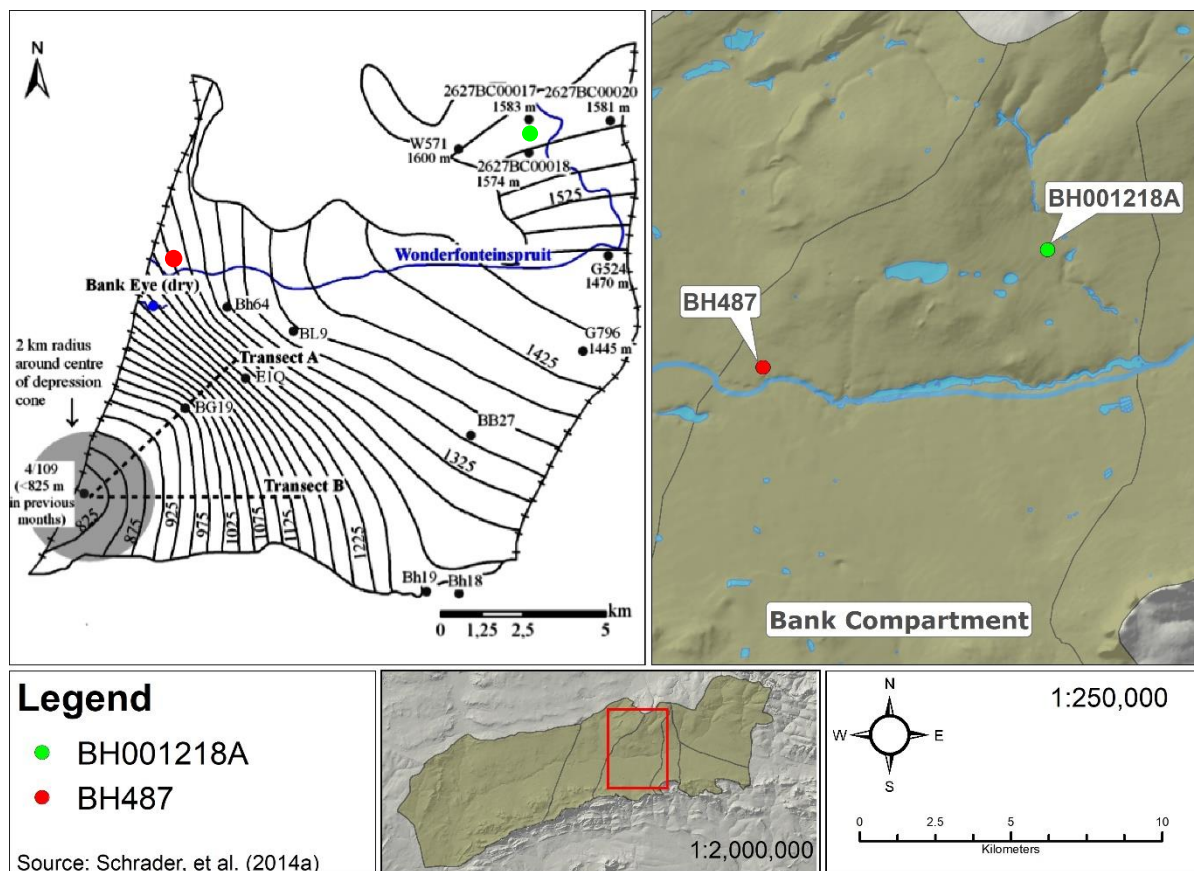


Figure 56: Monitoring borehole comparison locations.

Of the 21 monitoring boreholes, one within each compartment was selected on the basis of overlapping data. Because all the compartments are interconnected, each of the water budgets needs to be calibrated within the same time period. Figure 57 includes the monitoring boreholes selected for calibration as well as the time period chosen.

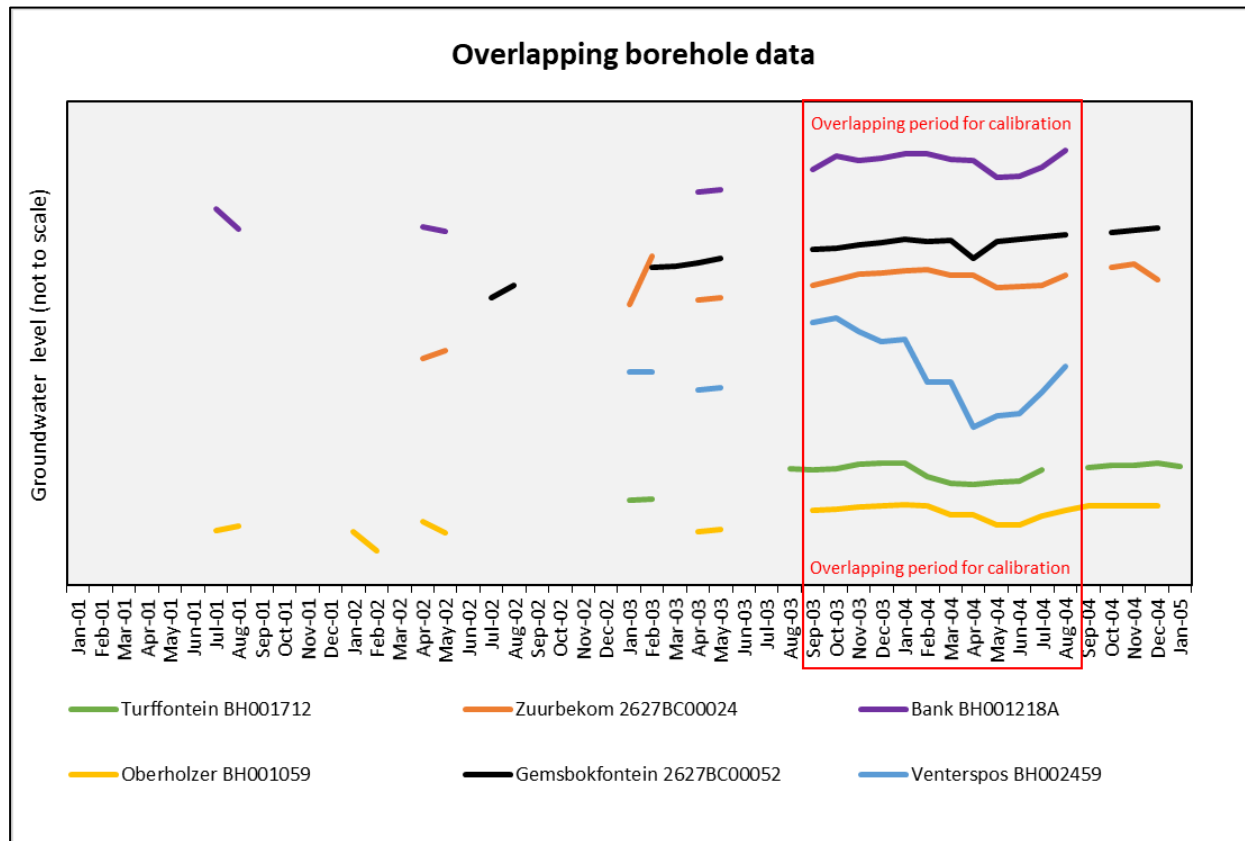


Figure 57: Overlapping borehole data for calibration.

5.4.2 Spring-representative boreholes

For the groundwater level recovery scenario, a borehole within each of the relevant compartments was chosen to represent the groundwater level of its representative spring. These boreholes obtained from the National Groundwater Archive (NGA) were chosen according to their distance from the eye/spring, the availability of groundwater level data, as well as the most recent groundwater level available. Figure 58 indicates four boreholes, one for each of the relevant springs.

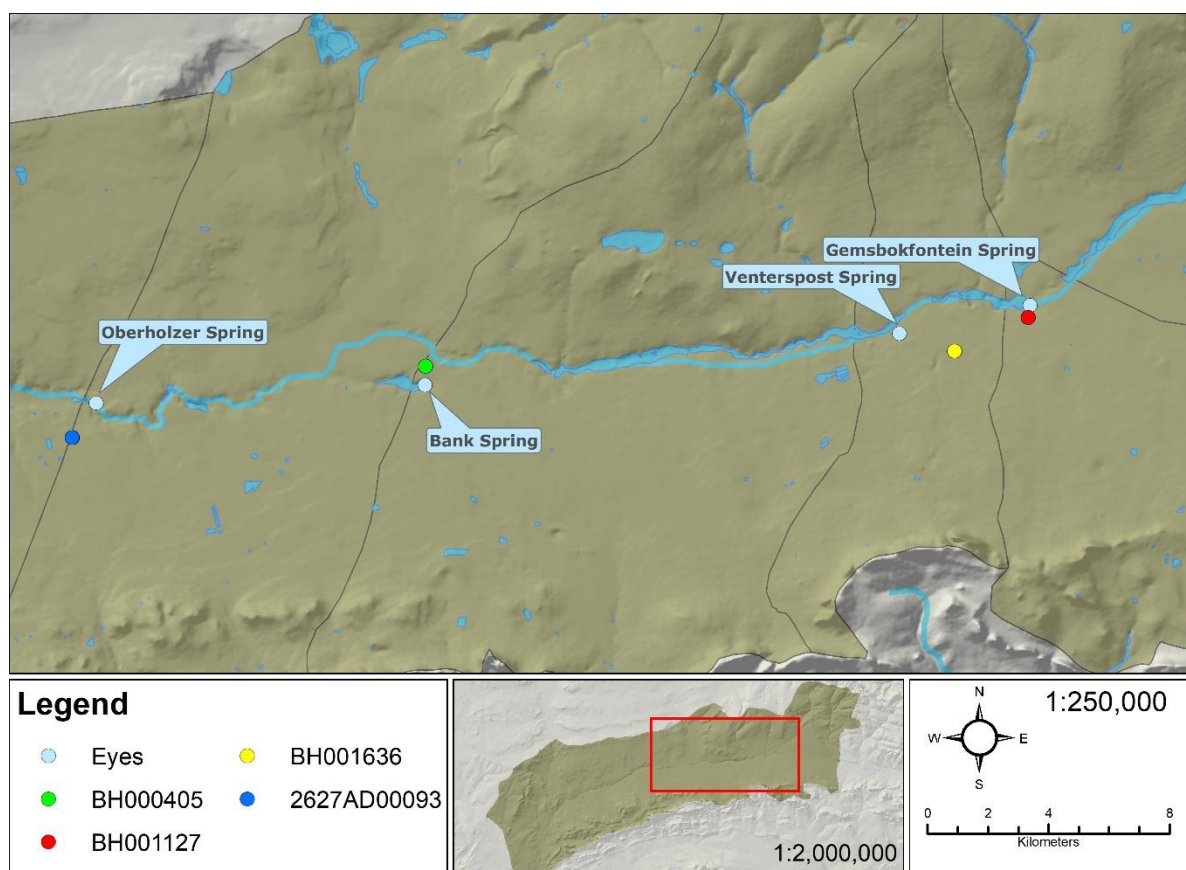


Figure 58: Spring-representative boreholes.

Due to a lack of boreholes with available groundwater level data, boreholes were used up to 1.7 kilometres from the eye with the overall available groundwater level data limited to the 1990s and in some cases 1970 and 1973, as seen in Table 32:

Table 32: Representative borehole information.

Eye	Closest Borehole	Distance between borehole and eye (m)	Recent Groundwater level (mbgl)	Date of measurement
Gembokfontein Eye	BH001127	53	22.2	1990/09/28
Venterspost Eye	BH001636	1730	19.25	1990/08/12
Bank Eye	BH000405	631	81.4	1970/10/28
Oberholzer Eye	2627AD00093	1333	54.6	1973/03/15

5.4.3 Rainfall

The FWR receives an average of 656mm precipitation annually. Furthermore, the rainfall ranges depending on the location within the study area with annual precipitation up to 1108mm.

In order to set up the SVF model, monthly precipitation data for the study area is required to correlate with the obtained monthly monitored water levels. Due to a decline in rainfall monitoring in recent years, precipitation data from both DWAF as well as SAWS was acquired.

Due to the varying precipitation across the area as mentioned in Section 2.7, two rainfall/weather stations were selected within different areas of the FWR based on data availability during the calibration period (Figure 59).

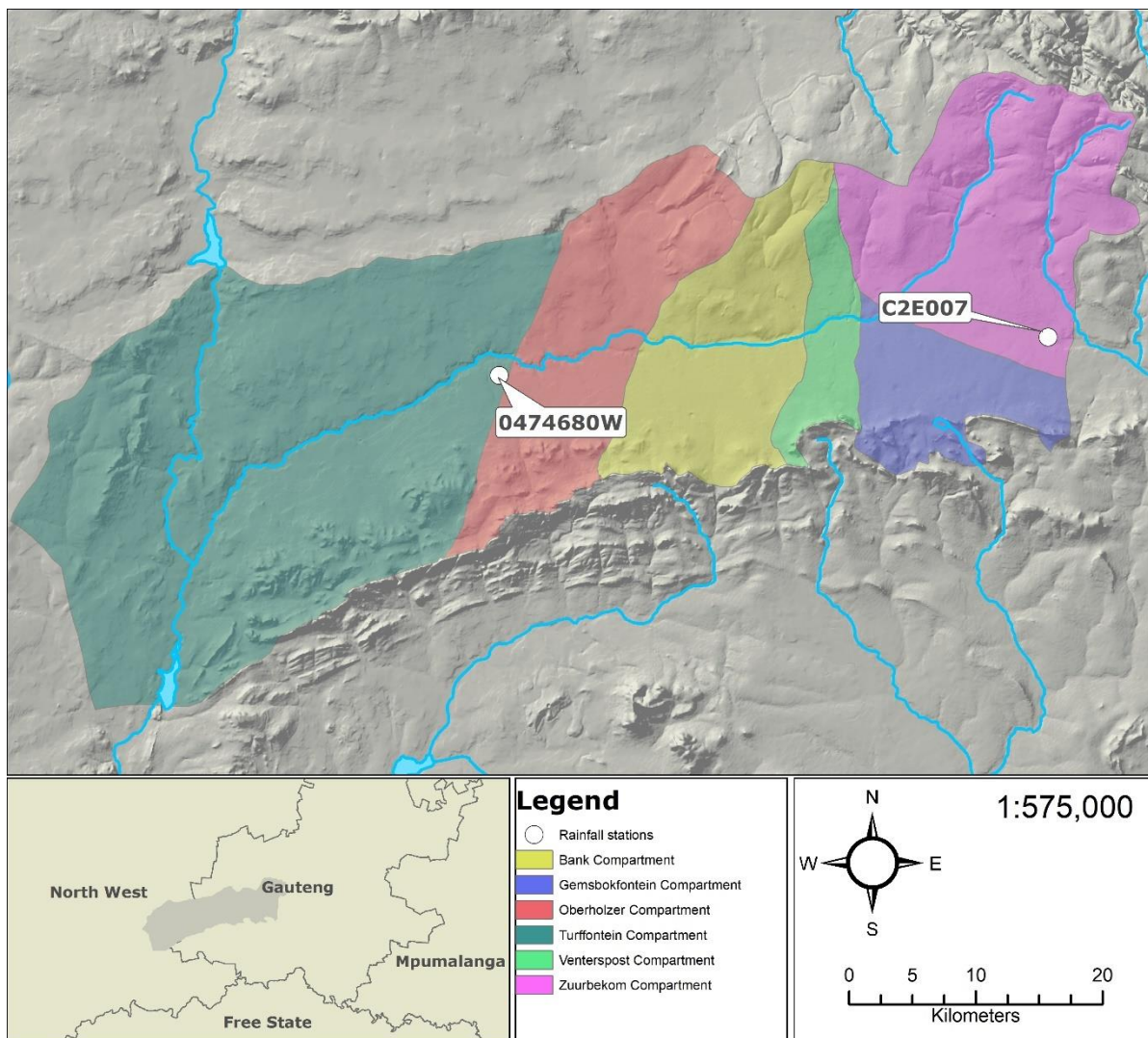


Figure 59: Rainfall station locations.

Closer to the headwaters of the WFS river in the Zuurbekom Compartment, rainfall station C2E007 from DWA (2019) was selected for the input precipitation data for the Zuurbekom-, Gembokfontein-, Venterspost-, and Bank Compartments. Rainfall station C2E007's data consists of monthly measured precipitation data dating back to 1962. Figure 60 includes the annual precipitation data for this compartment between 1962 and 2011.

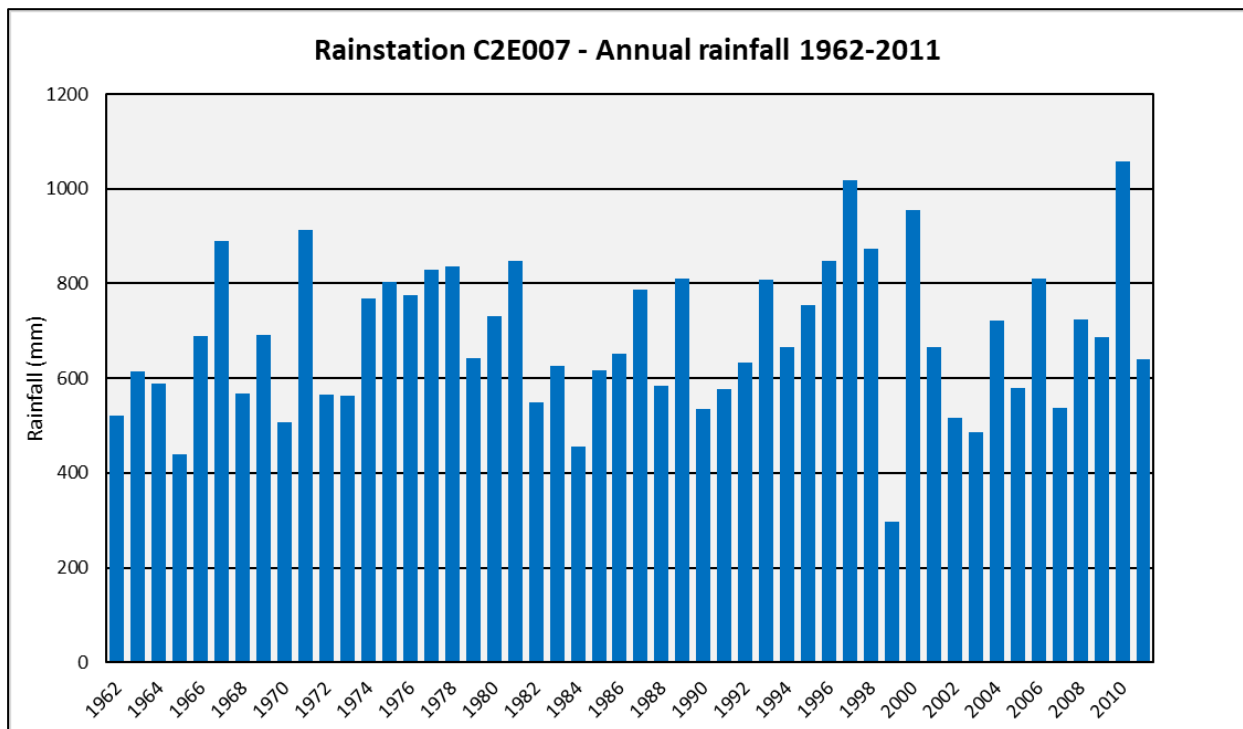


Figure 60: C2E007 – Annual rainfall 1960-2018.

On the Eastern border of the Turffontein Compartment next to the WFS river, rainfall station 0474680W from WR2012 (2012) was selected for the input precipitation data for both the Turffontein and Oberholzer Compartments. The rainfall stations data consist of daily measured precipitation data with several daily reading missing throughout (Figure 61).

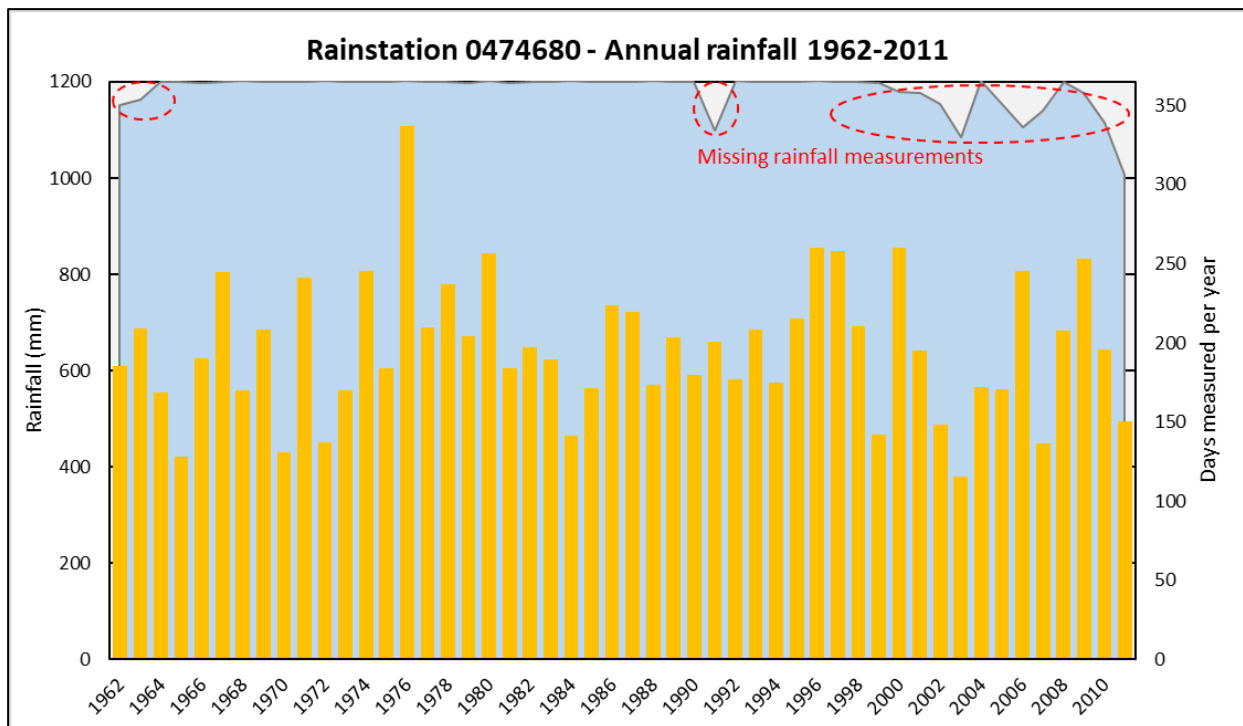


Figure 61: Rain station 0474680 – Annual rainfall 1962-2012.

Characteristic to the highveld region of South Africa, the study area receives the majority of its rainfall during the summer months of September until April/May (Section 2.7, Figure 8). Both of the selected rainfall stations contain sufficient precipitation data for the calibration period identified in Section 5.4.1 (Figure 62).

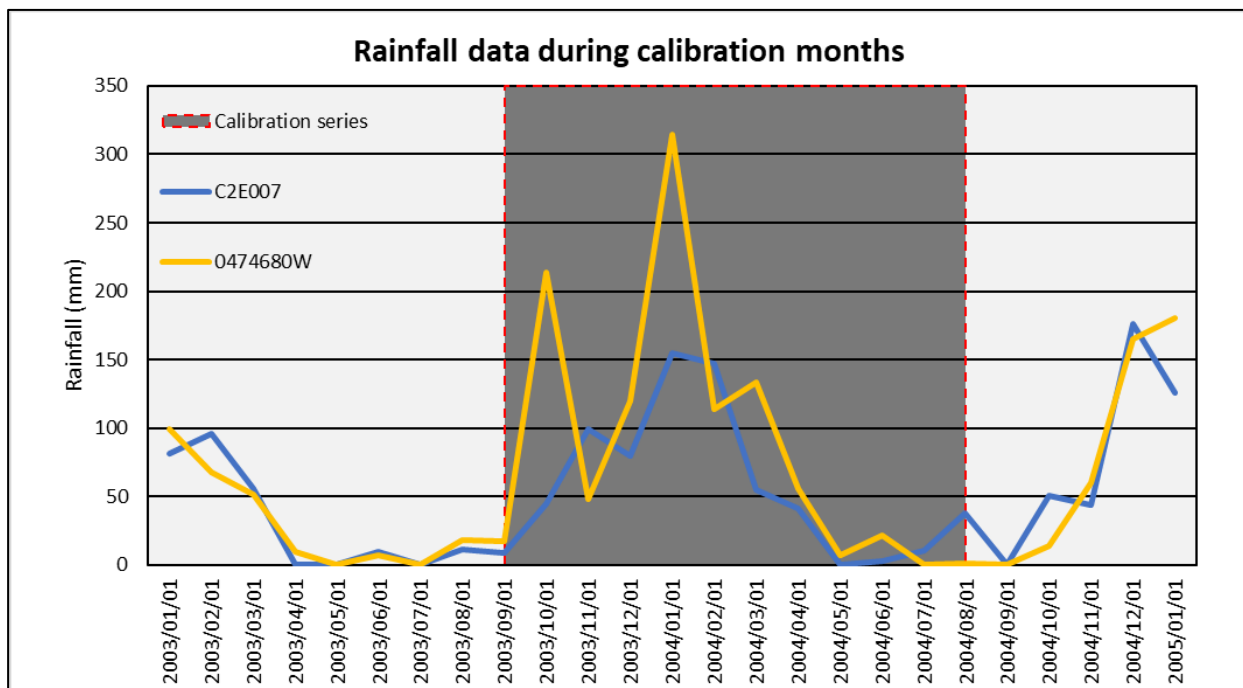


Figure 62: Rainfall data during calibration months.

Although the calibration period occurs during the peak rainfall summer months of the 2003/2004 hydrologic year, a delay can occur as mentioned by Van Biljon (2017) between the actual rainfall and the response within the groundwater levels. Therefore the months prior to the calibration period were also included for the calibration.

5.5 Future rainfall forecasting

Annual rainfall within the study area has stayed constant throughout the years and rarely goes below 400mm and above 800mm per annum as mentioned in Section 2.7. This aids the prediction for future rainfall predictions using the historical rainfall patterns.

To simulate future rainfall patterns for the use in the groundwater level recovery step, a histogram was created containing 10 years of recent historical rainfall events and the standard deviation determined. Random monthly rainfall patterns were created using the histogram of the historical rainfall sequences by changing the standard deviation one standard deviation up for wet annual rainfall seasons and one standard deviation down for dry annual rainfall seasons. The process was also repeated for estimating more extreme wet and dry annual figures by making use of two standard deviations up and down. In total five rainfall sequences were created (extremely dry, dry, normal, wet, extremely wet) from these random sequences to create a single 100-year rainfall sequence based on historical values. This process was used to create future forecasting rainfall data for both of the rainfall stations used during calibration. Figure 63 includes the first 10 years of simulated future forecasted rainfall patterns for both rain gauge C2E007 and rainfall station 0474680W.

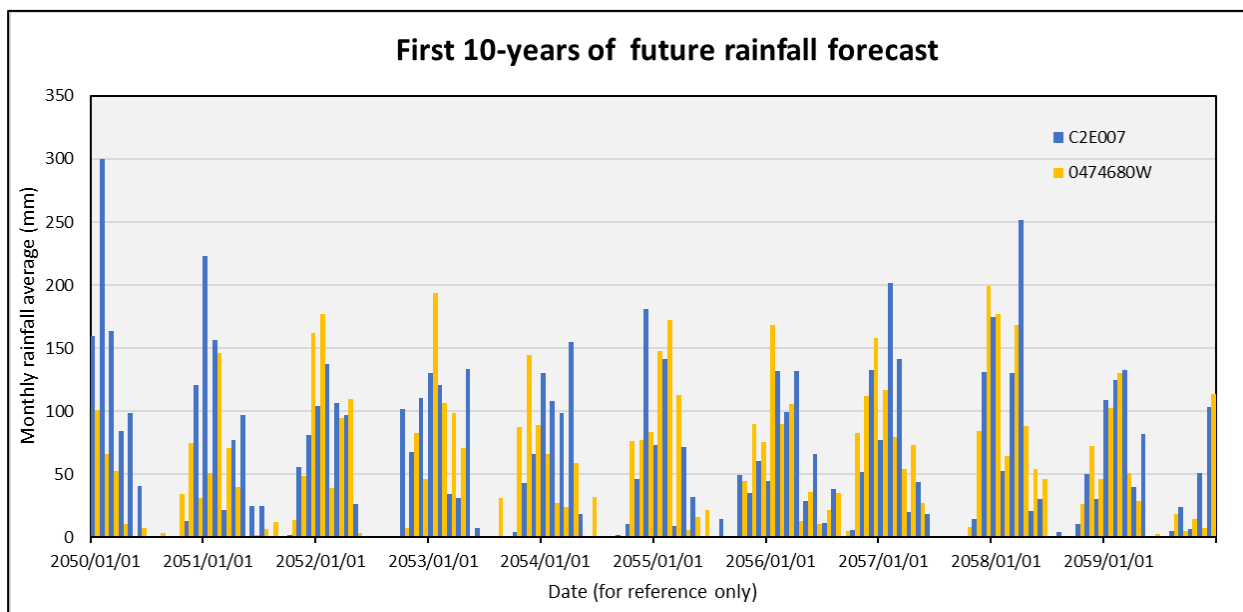


Figure 63: First 10-years of future rainfall forecast.

5.6 Conclusion

This chapter focused on the different methods and studies that were conducted on the FWR. It was noted that groundwater interaction between adjacent dolomitic compartments is largely influenced by IGF flows resulting from mined-out voids within the impermeable dykes. These IGF flows would largely affect any rewatering process as well as the possibility of full recovery. Most of these studies assessed the recharge and storage capabilities of the compartments based on their respective mine dewatering rates and pre-mining spring flow volumes.

It was estimated that the majority of recharge the aquifer receives comes from rainfall infiltration. The recharge value was found to differ across the entire study area due to the high heterogeneity found within the upper section of the aquifer system. Each of the dolomitic compartments developed according to their respective geomorphological characteristics and the amount of recharge depends greatly on the karst development within each of the compartments. Recharge ranges for each of the compartments were estimated based on the previous studies conducted on the study area as well as the implementation of the GIS-based APLIS method for recharge estimation in karst environments.

The SVF method was selected due to its ability to group parameters regardless of data deficiencies. The SVF method's lumped parameter approach will be implemented on each of the compartments of the FWR and calibrated according to present-day data. Data from monitoring boreholes for each of the compartments were selected based on time and spatial variability together within precipitation data. The next chapter will discuss how the FWR can effectively be conceptualised as well as how each of the SVF water budget equations were created and calibrated for each of the dolomitic compartments.

CHAPTER 6:

SVF WATER BUDGET MODEL

Based on the research conducted within the study area, the selection of the SVF water budget model as well as the specific data requirements was discussed in the previous chapter. This chapter focusses on understanding how all the gathered data and information aided in the creation of a water budget model by means of conceptual modelling. Based on the conceptual model, the methodology of how the SVF water budget model was created and calibrated for each compartment followed. Finally, the calibration results are discussed and compared to that of previous studies.

6.1 Conceptual Model

6.1.1 Introduction

By conceptualising the study area, all relevant information obtained in Chapters 4 and 5 was combined for an initial idea of how the system will function as well as the assumptions that should be implemented. Therefore, the conceptual model aimed at describing the topographical, hydrological, geological and hydrogeological environments and their interactions, whilst answering questions such as (Mandle, 2002):

- Is there enough data to describe the current hydrogeological conditions?
- Can the groundwater movement be characterized as one-, two- or three dimensional?
- How many aquifer systems are present within the study area and is the interaction between them important?
- How does the aquifer receive recharge? By precipitation or leakage from a river, drain, dam, or infiltration pond?
- What are the main aquifer outflows? Seepage to a river or lake, flow to a drain, or extraction by a pumping?
- Does the aquifer's hydrogeological characteristics remain uniform through the study area, or is there considerable variation over the site?
- Have the boundary conditions been defined based on a hydrogeological basis?

6.1.2 Model conceptualization

Similar to the pre-mining models created by Dill, *et al.* (2007) and Swart, *et al.* (2003a) the conceptual model representing present day conditions (Figure 64) for the study area consisted of several interconnected boxes, each representing a different compartment.

These boxes represent each of the main dolomitic aquifers' storage capacity with the groundwater level determined by the observed level within each borehole discussed in Section 5.4.1. Below these karst systems, the less permeable fractured system occurs that connects the groundwater flow between the dewatered compartments with the mine voids. Naturally, groundwater flow occurs in three directions with the majority of the flow along vertical and horizontal flow paths between the interconnecting dykes. Interaction between the different compartments dates back to pre-mining times in the form of seepage within the near-surface weathered dykes. To a greater extent mining activity along the dykes lead to subsequent mine voids within these dykes leading to IGF.

The inflows and outflows for each dolomitic compartment differ as its recharge characteristics are influenced by the current geology, hydrology, hydrogeology as well as human activities. Whilst recharge by means of precipitation contributes to all the compartments' basic inflows, other inflows are also existing and differ from compartment to compartment. At the present day these inflows include:

- The leakage from surface water bodies; mainly the WFS river, dams, wetlands and TSF on top of the karst dolomites.
- Leakage from the diverting canals as well as runoff/ percolation from the adjacent non-dolomitic rocks and endoreic areas.
- The above-mentioned seepage and IGF from adjacent compartments.

The outflows for each compartment differ with dewatering which is the largest outflow within the majority of the compartments. Additional outflows occur within the non-dewatered compartments: Zuurbekom and Turffontein, where the groundwater table attributes baseflow to the WFS river. Although there is ongoing pumping by the Rand Water board, it is assumed that the current dry conditions of the spring accounts for all pumping actives, resulting in a neglectable balance (Enslin & Kriel, 1968).

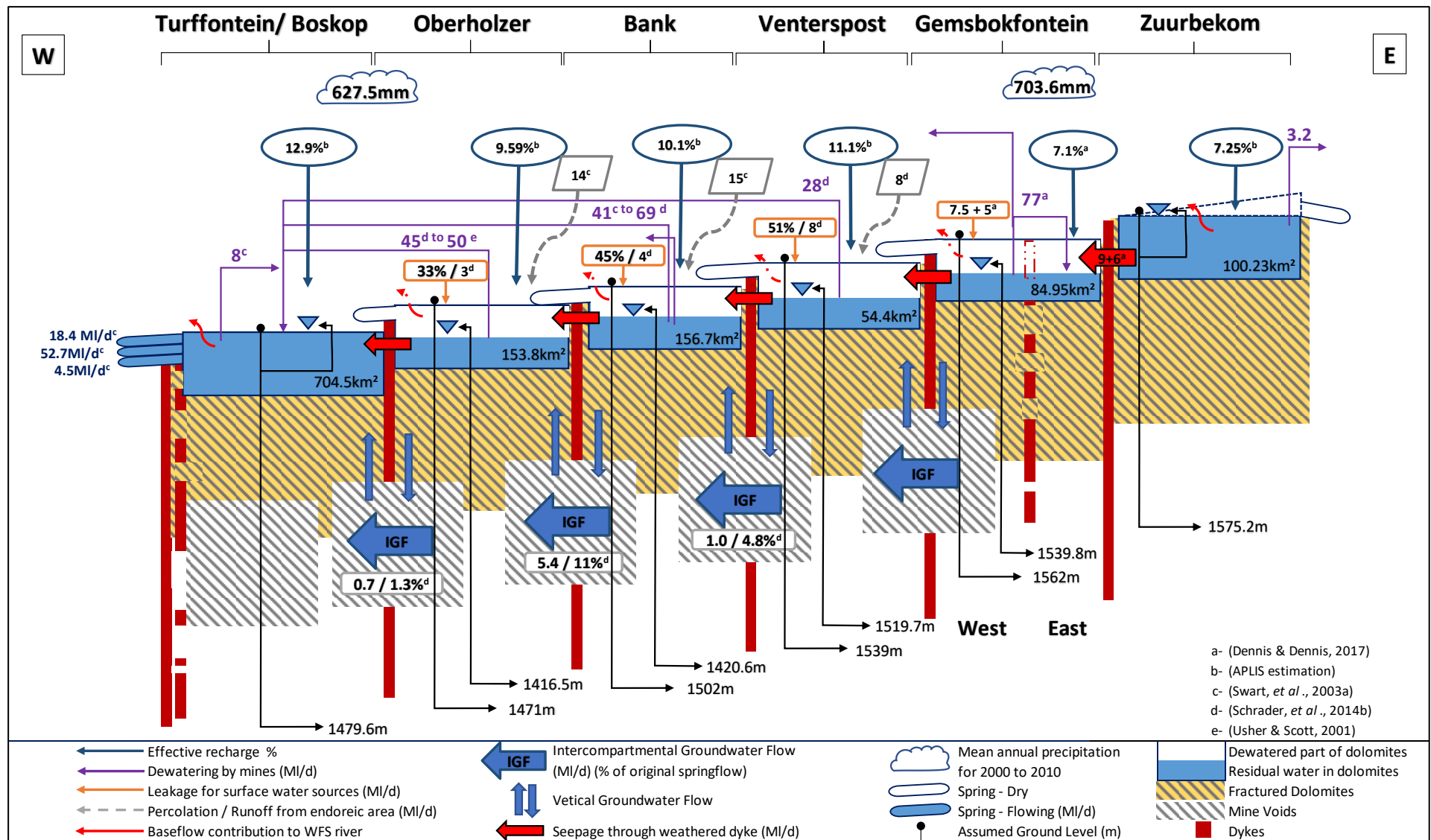


Figure 64: SVF conceptual model – present day workings.

Due to the SVF method's simplicity as well as the uncertainties regarding data accuracy, other assumptions included:

- Each compartment was assumed to be homogeneous with the same aquifer and geological properties throughout the compartment.
- The groundwater table for each compartment was based on the monitoring borehole water levels from Section 5.4.1.
- All inflows other than precipitation were grouped as a single “leakage” inflow term due to the absence of individual characteristics.
- During calibration and the recovery simulation, the Gemsbokfontein Compartment was used as a whole, consisting of both the Gemsbokfontein West and East.
- The contribution to baseflow from non-dewatered compartments were grouped together with the relevant compartments' groundwater interflow term.
- Groundwater interflow terms were controlled by conductance terms specified by groundwater head differences between neighbouring compartments.
- The change in storage or ΔS would remain zero. Implying that the storage component was present during simulation but did not participate in the flow between compartments.
- River/ waterbody leakages were head-dependent and controlled through a conductance term.

6.2 Model Development

6.2.1 Introduction

The modelling approach made use of several SVF water budget equations, each representing a different compartment. Based on an idealized flow assumption where generalized flow occurs from compartment “a” to “b”, “b” to “c” up to compartment “f”. Individual water budgets for each compartment enabled each of them to function as individual systems, each system with its own set of inflows and outflows governed by a set of aquifer parameters. In addition, each of these individual water budgets effected its downstream neighbour, resulting in an interconnected system for a total of five calibrated inflow/outflow terms (Q) between the six compartments.

6.2.2 Model Creation and Calibration

Each SVF water budget equation representing a dolomitic compartment: “a”- Zuurbekom, “b”- Gemsbokfontein, “c”- Vensterspost, “d”- Bank, “e” - Oberholzer and “f”- Turffontein was created within Microsoft Excel® according to Equation (3.7), Figure 65:

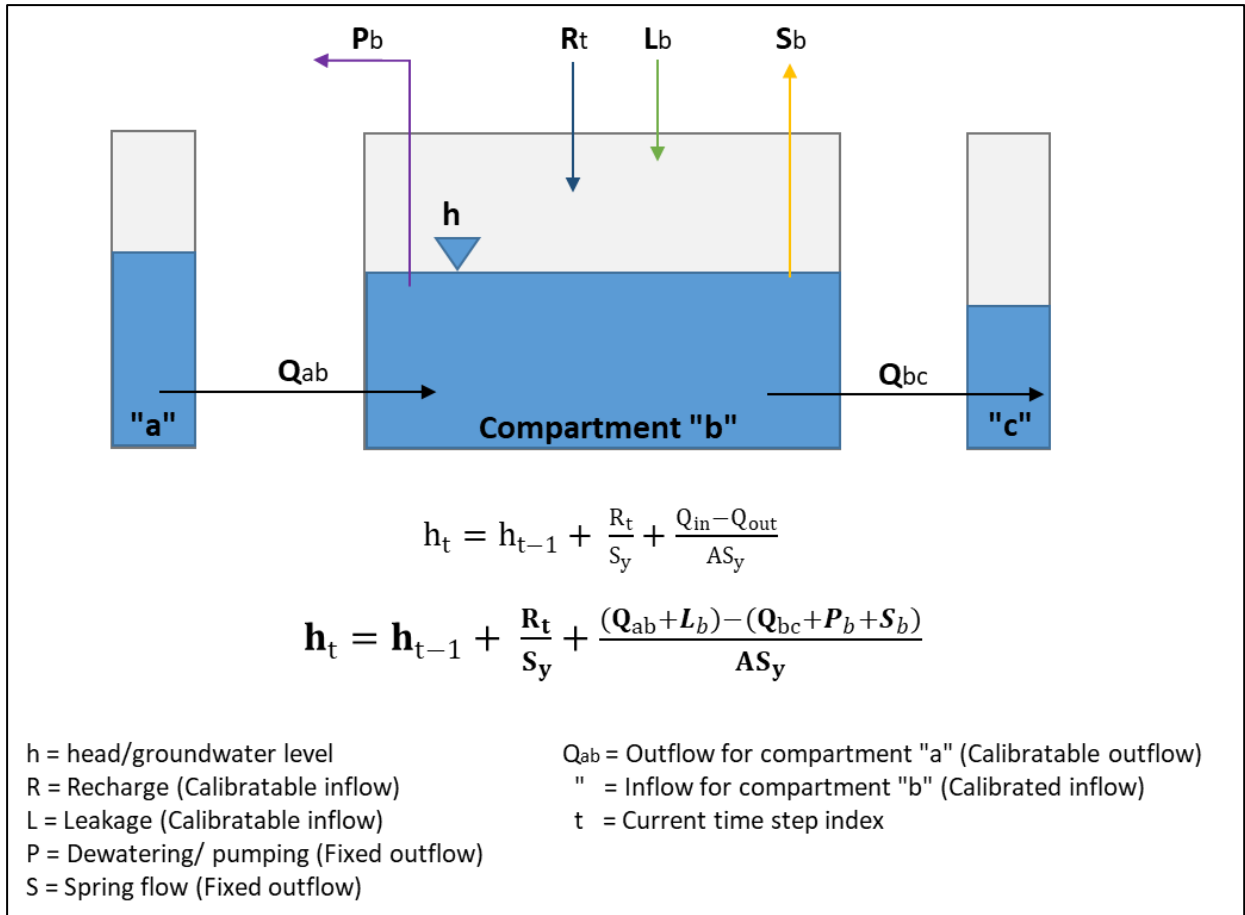


Figure 65: SVF-equation application.

Each groundwater level/head (h) at a specific time step (t) was influenced by the resulting simulated head value of the previous time step (h_{t-1}). Figure 66 includes the basic flow diagram used to construct and calibrate each compartment's SVF model.

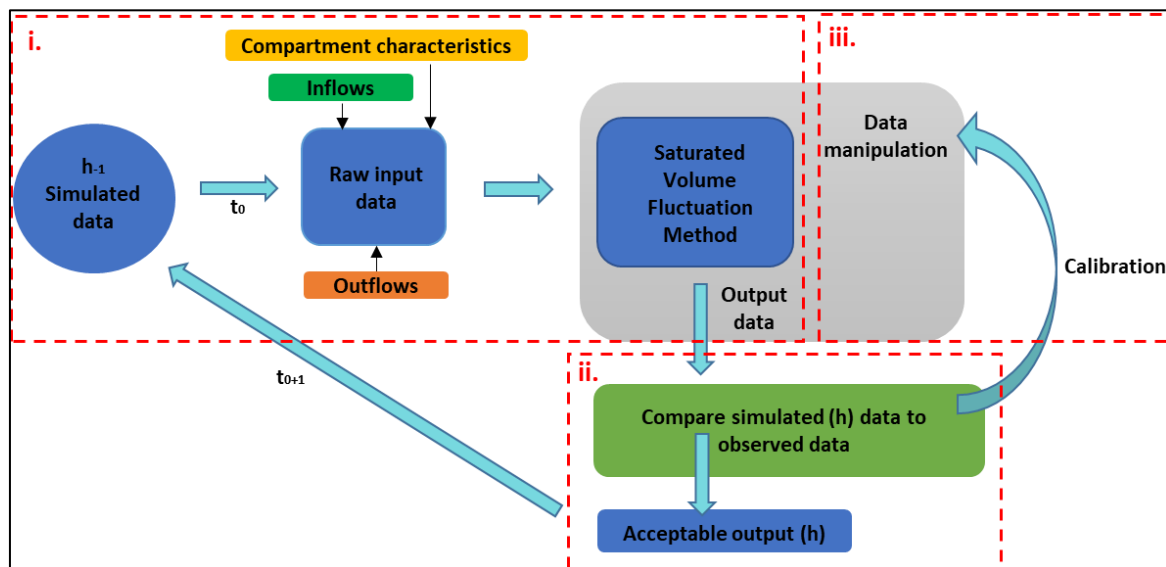


Figure 66: Calibration flow diagram.

- i. During t_0 the parameters: inflows, outflows and compartment characteristics were assigned within the SVF water budget along with the head in the previous time step (h_{-1}) to produce a head value (h) at the current time step of t_0 .
The parameters were further sub-divided into fixed and calibrated parameters. Fixed inflows included the monthly rainfall amount on the compartment (mm) as per the rainfall data from the applicable rainfall station mentioned in Section 5.4.3. Fixed outflows included the average pumping amount by mines determined in Section 5.2.13 as well as spring flow (where applicable). Fixed compartment characteristics included the outcropping dolomite area (m^2) for each of the compartments as mentioned in Section 5.2. Adjustable inflows included the percentage of rainfall that would recharge the aquifer or recharge percentage as well as inflows from adjacent compartments and surface water body within the leakage component. The latter was driven by a conductance flow term according to the water level within the compartment. Adjustable outflows included the outflow of groundwater into adjacent compartments based on a general conductance driven by the difference on the head (Δh).
- ii. This simulated head (h) was compared to the value within the selected observation borehole as mentioned in Section 5.4.1 at time step (t_0). A tolerance of less than 0.1m head difference between the simulated head and the observed groundwater level was deemed acceptable.
- iii. To calibrate the simulated head value to best fit the observed value at (t_0). Each of the adjustable parameters were altered, whilst the fixed parameters such as area size were kept constant throughout the simulated time series for the specific compartment.

To further calibrate the simulated head values to that of the observed values a general delay between the rainfall derived inflows and outflows were added to address the phase in which rainfall recharges the system as mentioned in Section 3.3. Once the groundwater level was calibrated for a specific time series, the conductance (C) could be calculated for both the inflow into each compartment as well as the outflows contributed by adjacent compartments by means of Equation 3.8, Figure 67:

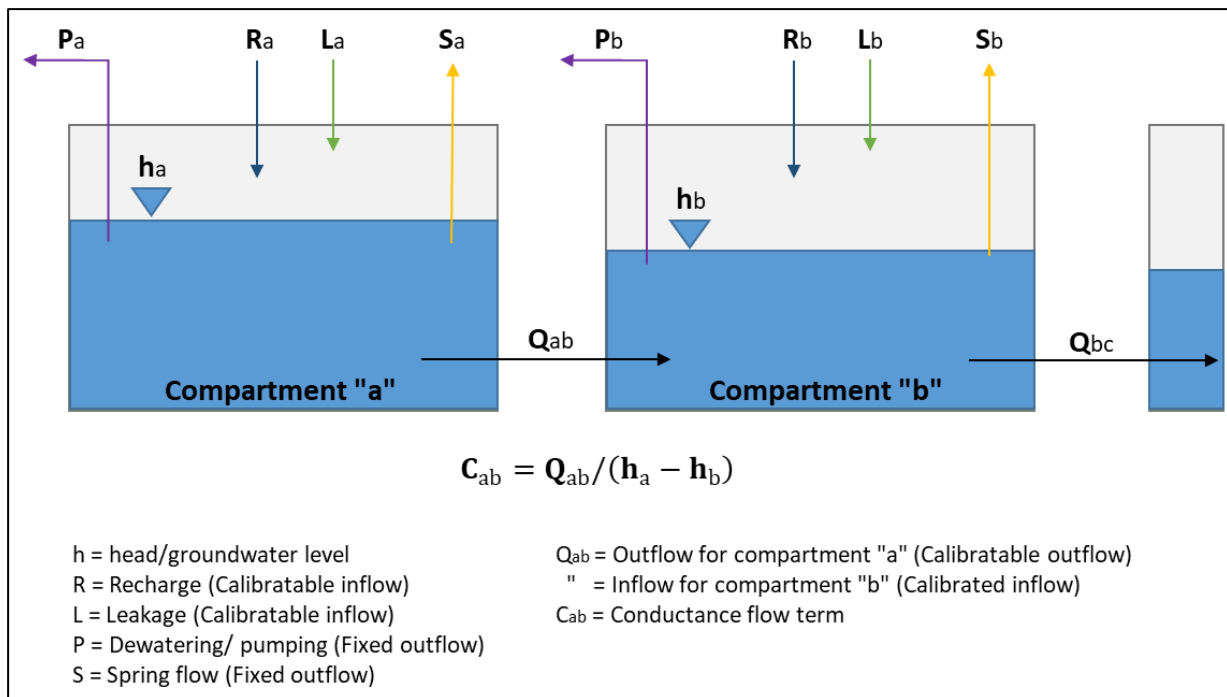


Figure 67: Conductance term application.

This conductance flow term governed the head-dependent flow (Q) that occurred between adjacent compartments. Each conductance between relevant compartments will be fixed throughout the recovery simulation with the Q flow between adjacent compartments driven by a change in the head (Δh).

A similar approach was taken to determine the conductance that governed leakage inflows. Where instead the Δh was established by the height of the dewatered groundwater table relevant to the "full" groundwater level determined by the ground elevation of the compartments particular spring.

Figure 68 includes the created and calibrated SVF model of the Gemsbokfontein Compartment within Microsoft Excel®.

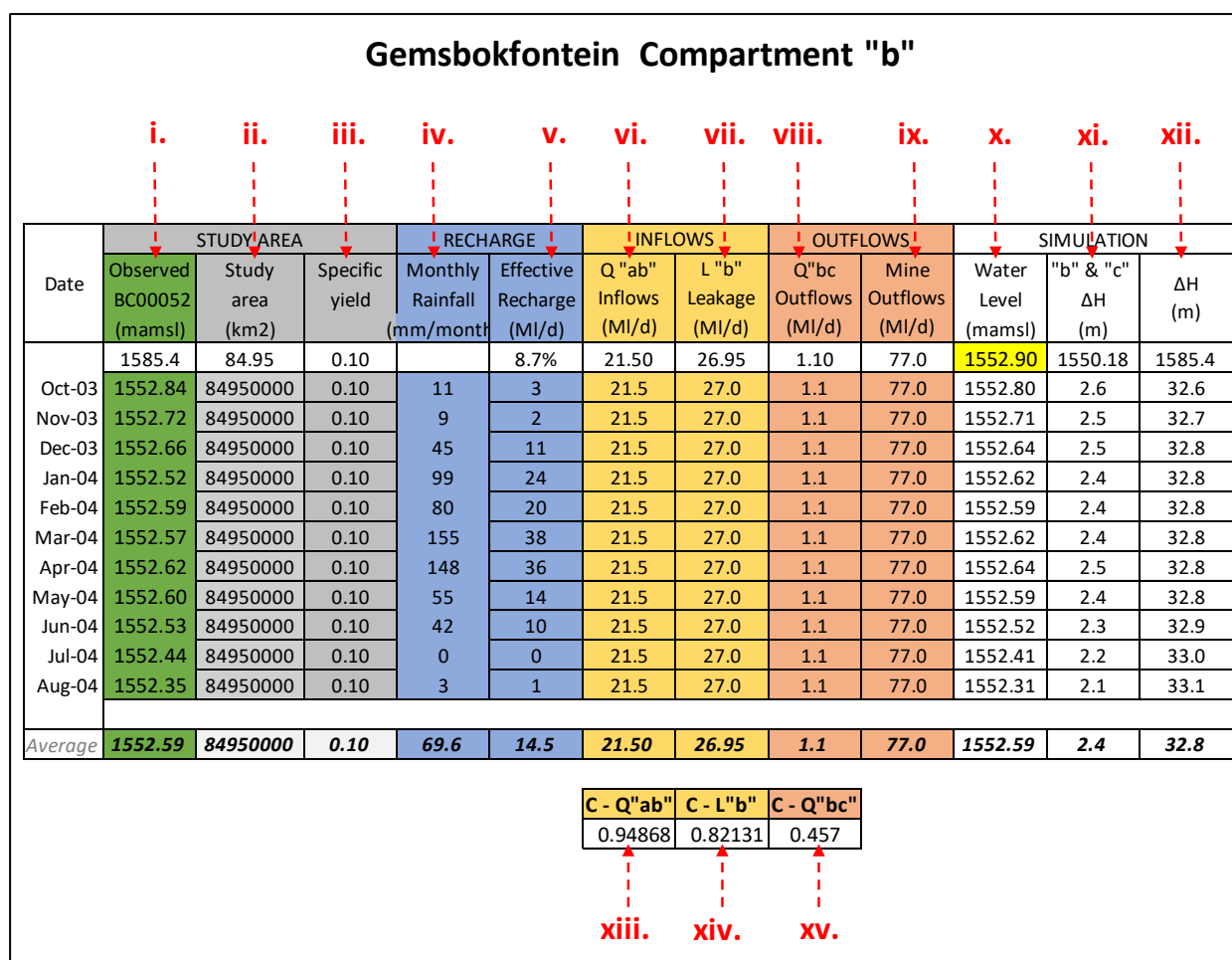


Figure 68: Gemsbokfontein Compartment "b" calibration breakdown.

Where:

- i. Observed groundwater levels at the monitoring BH-2627BC00052 during the specified calibration time period.
- ii. Study area size (fixed parameter).
- iii. Specific yield (Sy) for the compartment (calibration parameter).
- iv. Monthly rainfall (mm) obtained from rainfall station C2E007.
- v. Effective recharge, percentage (%) rainfall from the monthly value (calibration parameter).
- vi. Q"ab" Inflow from adjacent compartment (calculated).
- vii. Leakage from compartment surface water bodies and rivers including all other inflows (calibration parameter).
- viii. Q"bc" Outflow into the adjacent compartment (calculated).
- ix. Average active dewatering by mines within compartment (fixed parameter).
- x. Model simulated groundwater table specific to calibration time period.

- xi. Groundwater level difference between compartment “b” and “c”, for the calculation of conductance flow term C^{bc} .
- xii. Head difference between average simulated groundwater level and the ground level at the monitoring BH, for calculation of compartment leakage conductance flow term L^b .
- xiii. Calculated conductance term for Q^{ab} inflow.
- xiv. Calculated conductance term for L^b .
- xv. Calculated conductance term for Q^{bc} outflow.

6.2.3 Calibration Sensitivity Analysis

The initial parameter adjustments during the calibration process included the adjustment of the rainfall percentage ranging between the minimum estimated percentage from Table 26 up to the maximum. This was aided by the addition of leakage inflows to balance the outflows from the dewatering activities. Specific yield was initially assigned values of 0.05 and adjusted to best match the observed data's trend with additional adjustments to the inflows and outflow terms for further refinement. During the calibration process, it was observed that the effective recharge percentage - model parameter was the most sensitive to change.

A sensitivity assessment was performed on the four main adjustable inputs: specific yield, effective recharge percentage and leakage inflows and outflows (Q between compartments). The informal sensitivity assessment made use of the RMSE calculated for each calibration. By changing the parameters within fixed intervals, the change in the model simulation could be observed by the changing RMSE value. The largest change in RMSE reflected the most sensitive parameter (Figure 69). From the sensitivity analysis, it was evident that effective recharge percentage was the most sensitive parameter, followed by leakage inflows, outflow (Q between compartments) and specific yield. The sensitivity of the Q - outflow parameter did however mainly depend on how much volumes of water was removed by the pumping outflow term.

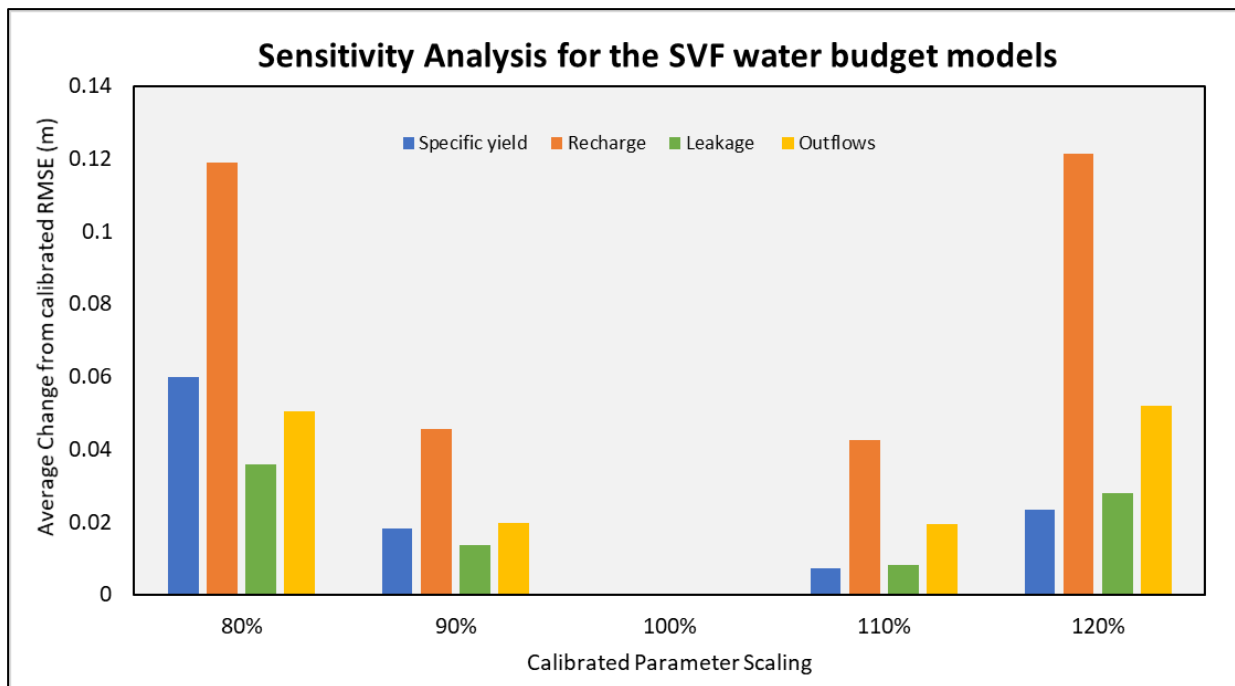


Figure 69: Sensitivity analysis results graph.

6.2.4 Calibration Results

The calibration process mentioned in Section 6.2.2 was repeated for each of the compartment's SVF water budgets. Adjustable parameters were changed using a “trail-and-error” approach. Each compartment was deemed calibrated once the obtained results, not only visually matched that of the observed values but also achieved good statistical matching. The three statistical methods discussed within Section 3.5 were used to evaluate each compartments' calibrated outputs based on the overall “goodness of fit” and error across the output.

6.2.4.1 Zuurbekom Calibration

The calibration of Zuurbekom Compartment during the 11-month time series from October 2003 until August 2004 made use of a 2-month precipitation delay and yielded an overall Pearson correlation of 0.944 or 94.4% between the simulated data and the observed data from BH-2627BC00024. The SVF-simulated groundwater sequence also obtained a low RMSE of 0.074m together with a very good NSC value of 0.89. These statistical methods indicated a well-defined fit between the simulated and observed data as evident in Figure 70.

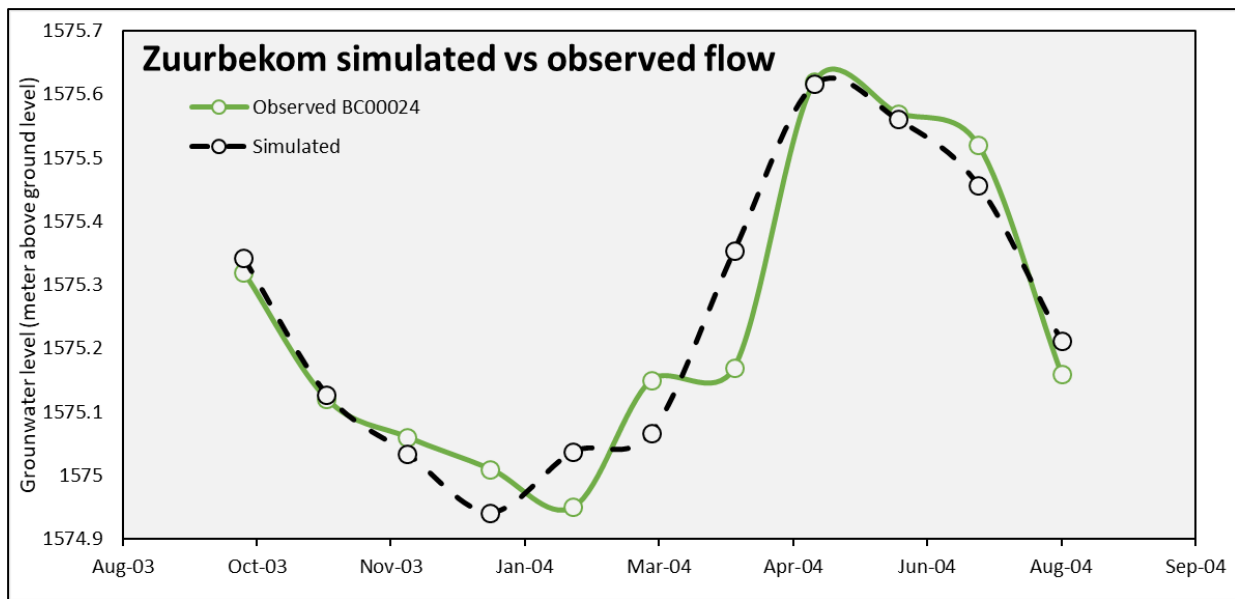


Figure 70: Zuurbekom simulated vs. observed flow.

A lower specific yield of 0.027 was obtained compared to those calculated for the dewatered compartments discussed later in the section. This may indicate lower amounts of karstification, and karst features directly influenced by the absence of extensive mine dewatering.

An effective recharge of 9.3% was determined, below the average recharge estimates from Table 27, Section 5.2.13. Resulting in a recharge value closer to that determined by GRAII than the APLIS estimate.

A leakage value of 2.94 MI/d in conjunction with an effective recharge of 19.4 MI/d nearly balanced the outflows represented by the pumping by Rand Water (3.6 MI/d) and the much higher Q_{ab} outflow of 21.5 MI/d. The resulting high Q_{ab} outflows was a direct result of the calibrated conductance term along with the large head difference between the non-dewatered Zuurbekom Compartment and the Gemsbokfontein Compartment. This in large parts represented the seepage that occurs through the near-surface weathered areas of Diagonal, Panvlakte and Magazine dykes; approx. 9+6 MI/d as previously estimated - indicated on Figure 64, Section 6.1.2.

6.2.4.2 Gemsbokfontein Calibration

As indicated on the graph in Figure 71, the simulated groundwater level matched the downward trend observed in BH-2627BC00052 after a 2-month precipitation delay was added. A Pearson correlation of 95% was obtained as well as a RMSE value of 0.040m and an NSC value of 0.90. This downward trend could have been caused either by an increase in pumping rate or due to seasonal fluctuation from the lower rainfall period that occurred during 2003's winter months (Figure 62, Section 5.4.3). With the pumping rate kept constant throughout the entire simulation run, the latter could have caused this small downward trend in the water level also evident within the Venterspost and Bank Compartments. These fluctuations in the water table due to rainfall indicate that the dewatered compartments are more sensitive to inflow and outflow changes that may indicate higher porosity characteristics and subsequently higher specific yield values as seen in Section 3.2.1.

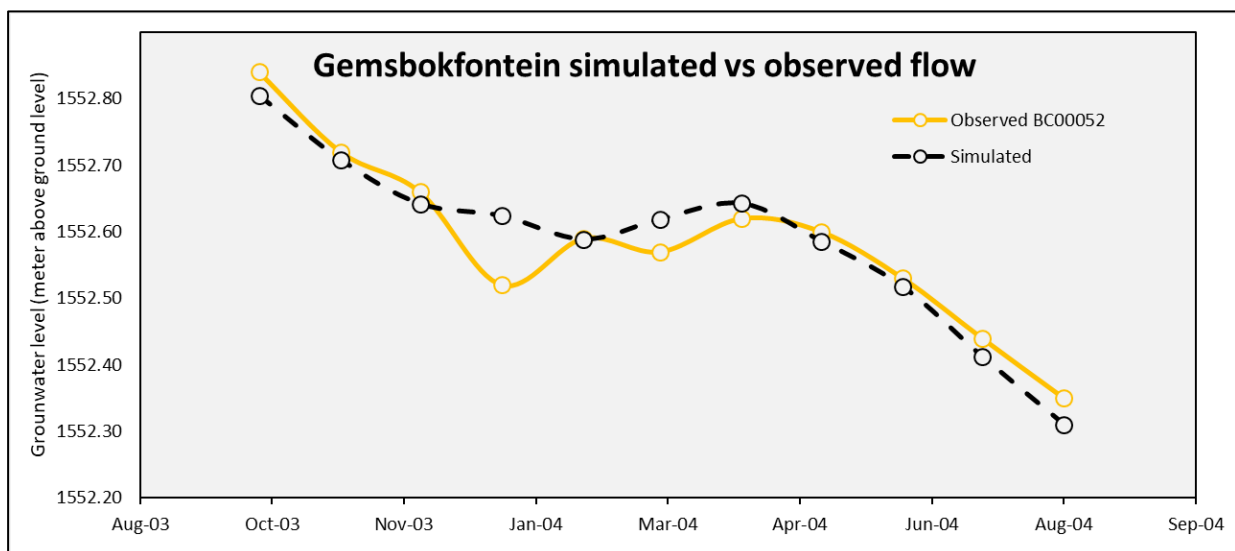


Figure 71: Gemsbokfontein simulated vs. observed flow.

Due to incredibly large volumes of water pumped from the Gemsbokfontein Compartment, high inflows had to be assigned throughout the calibration to counterbalance this outflow. Not only did this include the high amount of inflows attributed from Zuurbekom but also included high leakage amounts. Some of these leakages can be directly explained by the dewatering discharge, estimated at 8 ML/d by Swart, *et al.* (2003a) that occurs within the Gemsbokfontein East compartment. Resulting in recurring recharge back into the Gemsbokfontein West Compartment. Other reasons for such high leakage values can be directly attributed from a high seepage volume through the near-surface weathered Magazine dyke as well as the stream loss for approx. 7.5 ML/d from the Leeuspruit river and 5 ML/d from the Rietspruit river (Dennis & Dennis, 2017), Section 5.2.7.

A high specific yield of 0.10 was obtained from the calibration of the Gemsbokfontein Compartment. This can be directly attributed from the recirculation or due to high porosity within the aquifer or a combination of the two. A high porosity value can explain the higher leakage volumes into the compartment as well as the high required dewatering pumping rates estimated by (Dennis & Dennis, 2017). This possibly results from the large amount of karst features within the area close to the western boundary (Figure 34, Section 4.6).

A calculated recharge of 8.7% correlated well to the values obtained by Van Biljon (2017) and Dennis & Dennis (2017), taking into account that most estimates were only for the Gemsbokfontein West Compartment.

6.2.4.3 Venterspost Calibration

The observed groundwater level within the monitoring BH002459 was similar to the downward trend observed in the Gemsbokfontein compartment. The simulation of the water level closely matched the observed trend with small deviations where it was underestimating in the early stages. Regardless, the overall fit obtained a very high Pearson correlation of 97.6%; a RMSE value of 0.090m and a very good NSC value of 0.93, Figure 72.

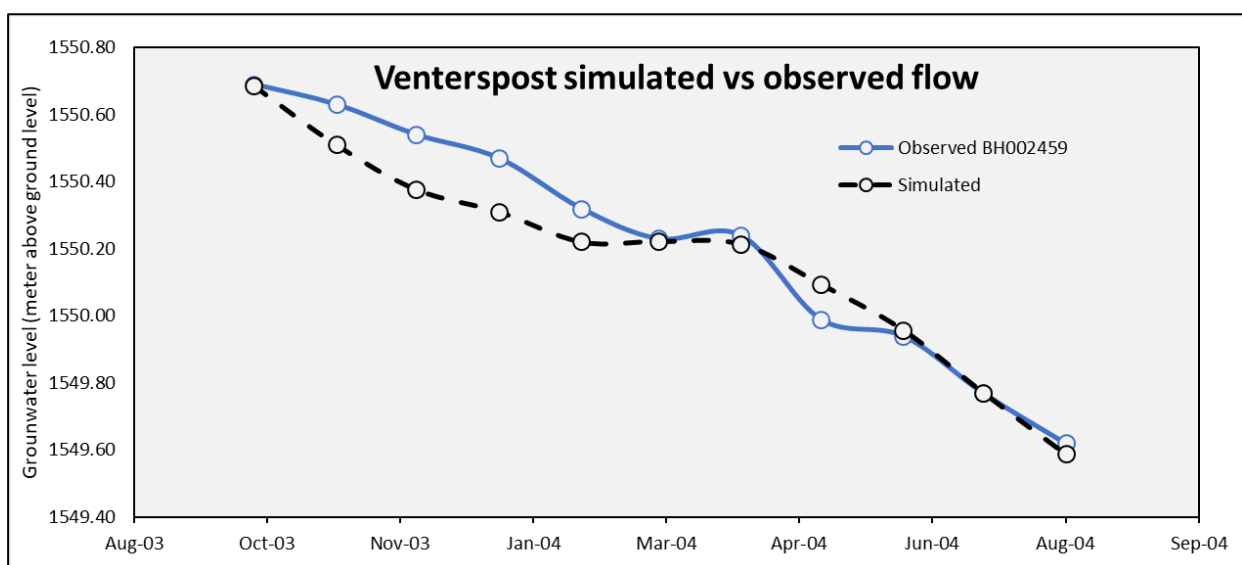


Figure 72: Venterspost simulated vs. observed flow.

An effective recharge of 7.9% closely matched that of the GRAII and the estimate made by Enslin & Kriel (1968). A relatively high specific yield value of 0.065 can be due to the increase in karst features such as sinkholes across the landscape (Section 4.6). This can be supported by the large amount of leakage inflows of 7.0 MI/d present with inflows, considering the non-perennial state of the WFS river and various spot recharge areas. Inflow from the Gemsbokfontein Compartment (Q_{bc}) amount to a little over 1.1 MI/d, this low interflow is

possibly due to the mainly intact Gemsbokfontein Dyke No. 2. The calibration also included the implementation of a 2-month precipitation delay between the rainfall event and effect on the groundwater table.

6.2.4.4 Bank Calibration

The simulation of the Bank Compartment closely resembled that of the Venterspost Compartment. The overall downward trend from the observed BH001218A values were closely matched with underestimation occurring in early times and some overestimations during late times, Figure 73.

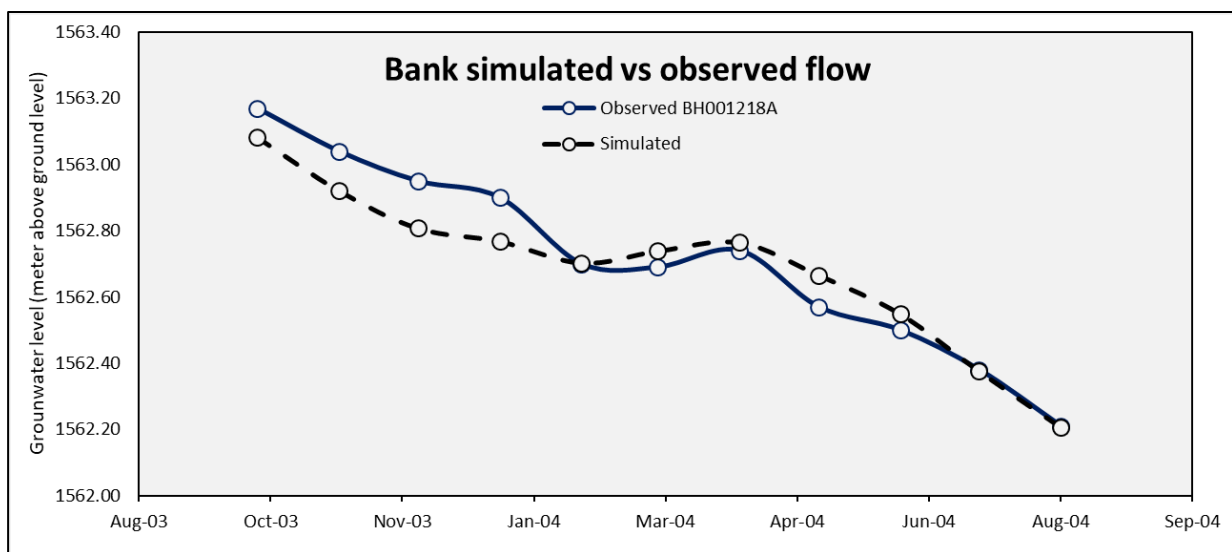


Figure 73: Bank simulated vs. observed flow.

Nevertheless, a good Pearson correlation of 96.8% was obtained indicating a strong correlation, supported by a good RMSE of 0.082m and a very good NSC of 0.91. The obtained value of 0.063 for the specific yield was very similar to that of the Venterspost Compartment. A higher recharge of 8.5% correlates well to that of the GRAII and is less than that by Van Biljon (2017) as well as the APLIS estimation (Section 5.2.10). A leakage inflow value of 7 MI/d supports the increased inflow due to a larger abundance of karst features but to a lesser extent than that of the Venterspost Compartment. The Bank Compartment made use of a similar 2-month precipitation delay applied to the above-mentioned compartments.

To address the different pumping values between 45 and 69 MI/d (Table 27), the average pumping rate was adjusted alongside the Q_{de} outflow. Finally, a pumping rate of 60 MI/d was used during calibration, in between the higher estimate by Schrader, *et al.* (2014b) and the lower estimate by Swart, *et al.* (2003a) with an acceptable Q_{de} outflow of 6.2 MI/d. This value for Q_{de} correlates well with that of Schrader, *et al.* (2014b), where Q_{de} (Bank- Oberholzer) is

approximately five times that of Q_{cd} (Venterspost-Bank). This can mainly be due to the extensive mining activities along the “Big Boy” fault where a large volume of mine voids attribute to the increased IGF flow between the compartments.

6.2.4.5 Oberholzer Calibration

The Oberholzer Compartment differed from those previously discussed with its rainfall sequence obtained from Rainfall station 0474680W as appose to C2E007. The simulated sequence best fitted the observed water levels from BH001509 once a 3-month precipitation delay was implemented to account for the recharge movement into the aquifer (Figure 74).

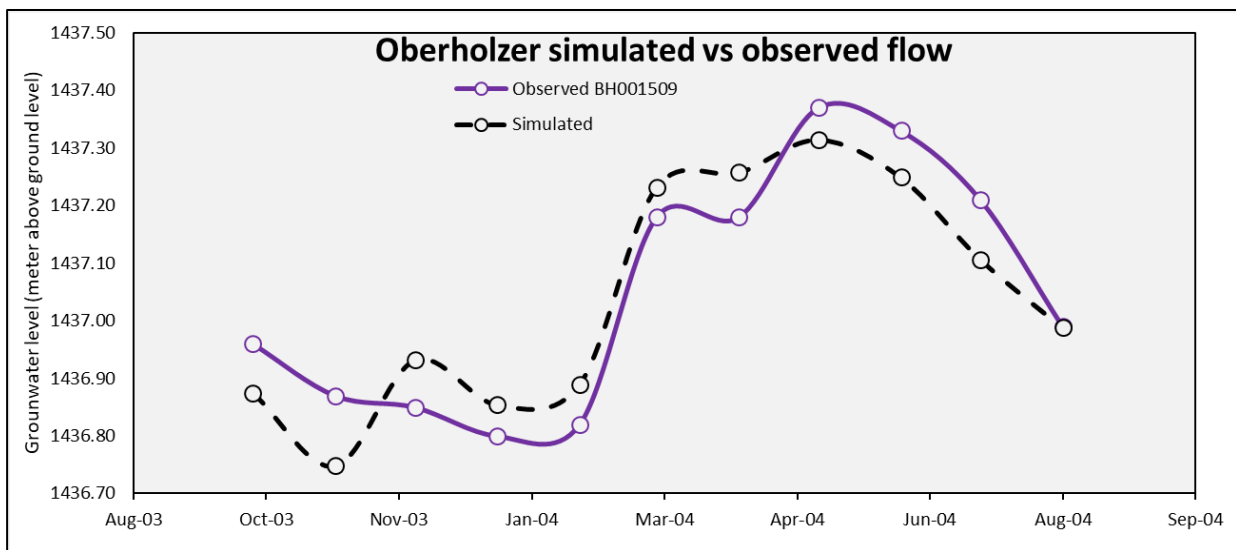


Figure 74: Oberholzer simulated vs observed flow.

The simulated water levels loosely fit the overall trend and shape of the observed. Although the simulation overestimated during the initial stages and underestimated during the final stages, it mainly fitted to the bulk of the observed water levels. A Good Pearson correlation coefficient of 0.924 or 92.4% was obtained alongside a RMSE for 0.077m and a good NSC value of 0.85.

The effective recharge of 8.2% (48MI/d) was close to that of the GRAIL as well as the APLIS estimate. An inflow of 4 MI/d was estimated for all other inflows represented by the Leakage inflows. Together with the 50 MI/d estimated pumping component, a further 1.1MI/d contributed to the outflows for the Oberholzer Compartment as Q_{ef} outflow.

6.2.4.6 Turffontein Calibration

The simulation of the Turffontein Compartment performed the worst with regards to the previous compartments with an acceptable Pearson correlation 89.2%, RMSE value of 0.15m

and a good NSC value of 0.78. As seen from the visual match (Figure 75), the simulated water levels matched the bulk of the observed water levels from BH001712 based on the same rainfall sequence used for Oberholzer. The simulation included the implementation of a 4-month precipitation delay to best match the simulated water levels to the observed water levels, accounting for the general delay that occurs between a rainfall event and its effect on the groundwater table as mentioned in Sections 3.4.3 and 5.4.3.

Due to the Turffontein Compartment's large size, a 6.0% effective recharge rate was obtained resulting in 101.8 Ml/d, together with Q_{ef} inflows (1.1 Ml/d) as well as 3 Ml/d inflows due to leakage. The Compartment's large amounts of inflows were balanced by similar outflow values, most is attributed by the Gerhardminnebron Spring (52.7 Ml/d approx.); Turffontein Springs (17.4 Ml/d approx.); and Boskop Spring (45 Ml/d approx.) (Usher & Scott, 2001). Additional outflows amounted to 38.6 Ml/d with a low specific yield of 0.30.

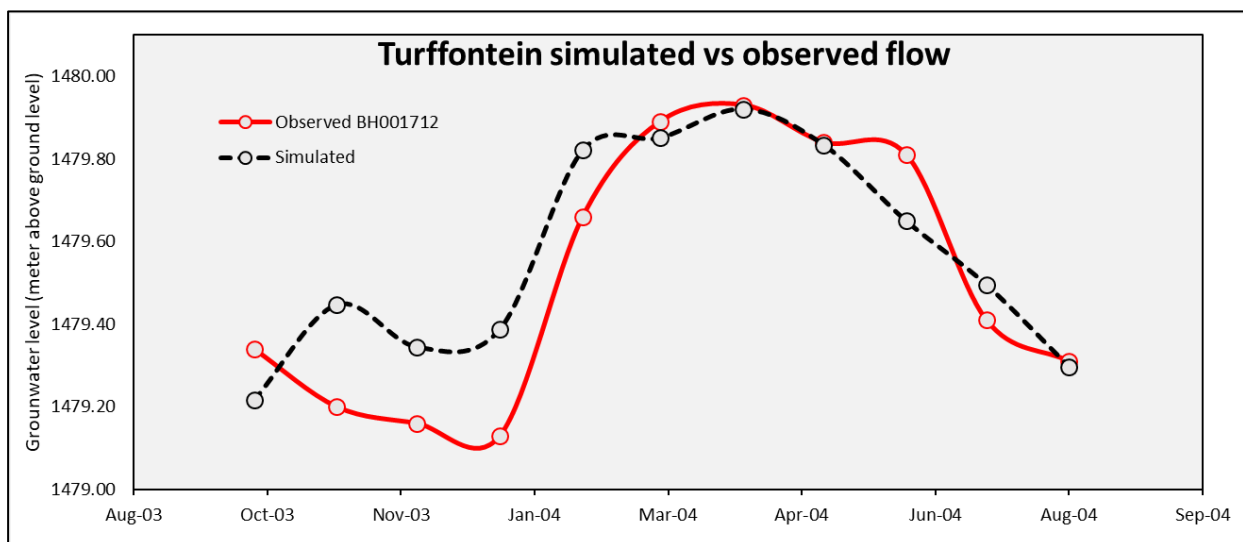


Figure 75: Turffontein simulated vs observed flow.

6.2.5 Conclusion

By using a combination of estimated as well as assumed parameters, the SVF water budgets for each of the dolomitic compartments were created. Calibration parameters were assessed based on their sensitivity and calibrated accurately to the observed water level data. The simulated results were considered satisfactory based on a visual match against the observed values from the monitoring boreholes as well as the calculated statistical methods. Typical deviation was observed throughout, due to the nature of time and spatial differences in rainfall patterns as mentioned by Erickson, *et al.* (2010). The parameter grouping and assumptions also contributed to the over and underestimations throughout but overall, each compartment

was deemed calibrated with Pearson correlation values rarely going below 90%, RMSE values mostly below 0.1m and NSC values above 0.75.

Table 33 includes a summary of the calibrated parameters for each of the compartments as well as the calculated conductance terms for both the intercompartmental flow Q and the Leakage inflows L.

Table 33: Calibration parameter results.

	Zuurbekom	Gemsbokfontein	Venterspost	Bank	Oberholzer	Turffontein
Specific Yield	0.027	0.100	0.065	0.063	0.052	0.030
Effective Recharge %	9.30%	8.70%	7.90%	8.50%	8.20%	6.02%
Effective Recharge (MI/d)	19.42	14.47	9.96	30.88	48.04	143.86
Q Inflows (MI/d)	n/c	21.5	1.1	2.2	6.2	16.1
"C" Q- Inflows	n/c	0.949	0.457	0.173	0.049	0.144
L Leakage (MI/d)	2.943	27.0	7.0	7.0	20.0	3.0
"C" Q Leakage (MI/d)	n/c	0.821	0.191	0.511	0.275	n/c
Q- Outflows (MI/d)	21.5	1.1	2.2	6.2	16.1	28.6
"C" Q- Outflows	0.949	0.457	0.173	0.049	0.144	n/c
Mine Outflows (MI/d)	3.6*	77	28	60	50	n/a
*Pumping by Rand Water Board					Averages across simulation period	
n/c = not calculated						

In the next Chapter, the calibrated parameters listed in Table 33 in conjunction with the forecasted rainfall predictions will be used in simulating a post-mining scenario where all dewatering pumping has ceased, rendering the largest outflow component observed in Chapter 6 inoperable.

CHAPTER 7:

FUTURE SCENARIO MODELLING

7.1 Introduction

After each of the SVF water budgets were created and calibrated in Chapter 6, the obtained parameters (Table 33) were used to simulate groundwater level recovery for each of the dewatered dolomitic compartments. Modelling of the FWR in terms of the current hydrogeological conditions can not only assist in understanding the current status of these karstified aquifers but also assist in understanding how these aquifers will react in the foreseeable future. The scenario implemented in this chapter assumes a future where the mining within the FWR has ceased and subsequently the ongoing dewatering activities related to the mining activities have stopped. Since the majority of outflows within the dewatered compartments no longer apply, the overall levels within each compartment will raise, resulting in a recovery/rewatering process.

Similar to the model consisting of several water budgets created in Section 6.2.2, the SVF water budget equation was used based on the established aquifer parameters (Table 29). The recovery scenario's purpose is to simulate water level recovery and spring reactivation by means of a water budget model.

7.2 Background

The ongoing mining activities that dominated the FWR for almost a century (Section 2.2) has caused the large water-bearing dolomitic aquifers to be dewatered due to flooding issues occurring within the mine voids underneath the vast aquifers. The dewatering of the dolomitic compartments and consequent recirculation of discharge water as well as piercing of the impermeably compartmental dykes (Section 4.5) has altered these aquifers' hydrogeological properties. These alterations have extended beyond all previous knowledge with the exact present state of these aquifers systems unknown.

This is of importance as gold mining within the WR and FWR along the WWL is slowly coming to an end, resulting in the termination of dewatering activities within the relevant compartments. The prospect of a rewatering phase comes with a list of uncertainties. Several attempts to understand and even predict these uncertainties have been undertaken over the years. Studies by Swart, *et al.* (2003a) and Schrader, *et al.* (2014b) mathematically forecasted recovery phases of the dewatered aquifers. Both studies by Swart, *et al.* (2003a) and Schrader, *et al.* (2014b) as well as studies by Dennis & Dennis (2017) and Van Biljon (2017) estimated

that the compartments in question would rewater to pre-mining water levels and the once high yielding springs of the WFS catchment will resume flow with little changes in the overall spring yield. Another study worth mentioning is that of Jordaan, *et al.* (1960) who estimated the formation of a “mega-compartment”, where all the individual dolomitic compartments would in theory act as one system as a result of being hydraulically connected (Section 5.2.6). According to Jordaan, *et al.* (1960) as mentioned by Schrader, *et al.* (2014b) this would not only lead to one universal groundwater level across all compartments but also cause most of the springs to remain dry.

Therefore, it's not only important to estimate the current properties that have resulted from mining activities within the area (Chapter 6) but also to conduct a recovery scenario for approximating how the different compartments will rewater. How long the rewatering process will take place? If the water level will return to pre-mining levels? What is the resulting spring flow and how does it compare to the pre-mining estimates?

7.3 Recovery Scenario Development

At its core, the SVF-model contains several water budget equations consisting of inflows and outflows. Based on the conservation of mass (Section 3.4), the water budget accounts for all water entering the aquifer such as recharge from precipitation together with surface water leakages and IGF and all water exiting the aquifer in the form of spring flow, IGF, dewatering and or other uses. Any change in the system's steady-state balance results in a transient change in the groundwater level, causing an increase or decrease depending on the net water balance.

Based on the sensitivity analysis conducted in Section 6.2.3, it was found that recharge played the most significant role in the model due to the SVF being a mainly rainfall driven model in regard to the study area. With no future rainfall data available to create a future recovery scenario, a forecasted future rainfall sequence was created from long-term historical rainfall sequences for the area (Section 5.5). The forecasted rainfall sequence includes 100-years of future rainfall data, based on the existing data with added dry and wet season variability for each of the data sets from the two rainfall stations as shown in Figure 76.

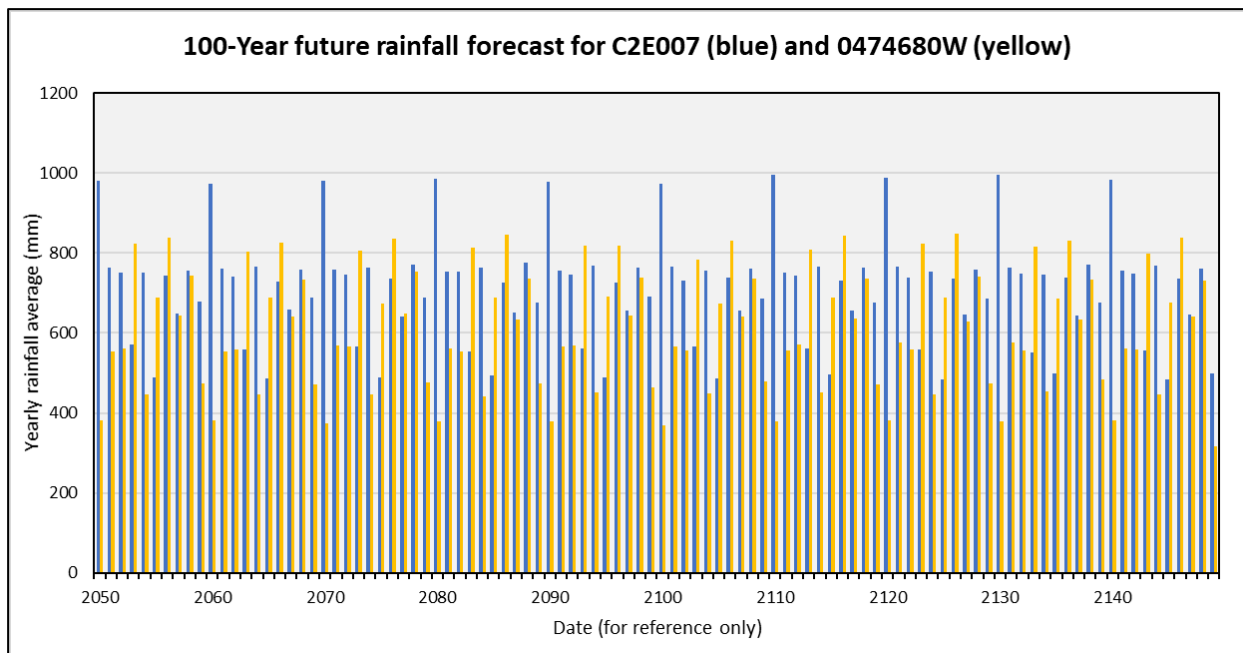


Figure 76: 100-year future rainfall forecast.

During the development of the recovery scenario, the following assumptions were taken into account:

- Each compartment was assumed to be homogeneous with the same aquifer and geological properties throughout the compartment.
- The initial dewatered water table for each compartment was based on the spring-representative borehole water level from Section 5.4.2.
- A leakage inflow term governed all inflows other than precipitation, as determined in Section 6.2.2.
- The recovery of the Gemsbokfontein Compartment was considered as a whole.
- The pipeline across the dewatered compartments will remain operational, conveying the majority of the water from the upstream WFS up to Donaldson dam.
- All mining activities are set to stop simultaneously, therefore dewatering will conclude within all dewatered compartments at the same time resulting in a simultaneous recovery start.
- No decanting across the mining infrastructure will occur as a result of rewatering.
- As mentioned in Section 6.2.2 groundwater interflow terms, as well as leakages, are controlled by the calculated conductance terms (Table 29) as well as the groundwater level between neighbouring compartments determined in Section 5.4.2.
- The water level for each dewatered compartment was assumed recovered once/if the water level reaches the relevant compartment springs' ground elevation.

The scenario made use of the SVF method based on Equation (3.10), Section 3.4.3. This linear equation made use of the SVF water budget Equation (3.7) developed by Breckenkamp, *et al.* (1995) with the addition of the calculated inflow/outflow conductance terms using Equation (3.8). Resulting in a water budget equation that can simulate fluctuations within the water level with the addition of head-dependent inflows and outflows on top of the normalised inflows and outflows. Figure 77 show cases the equation used for each of the dewatered compartments for the recovery scenario.

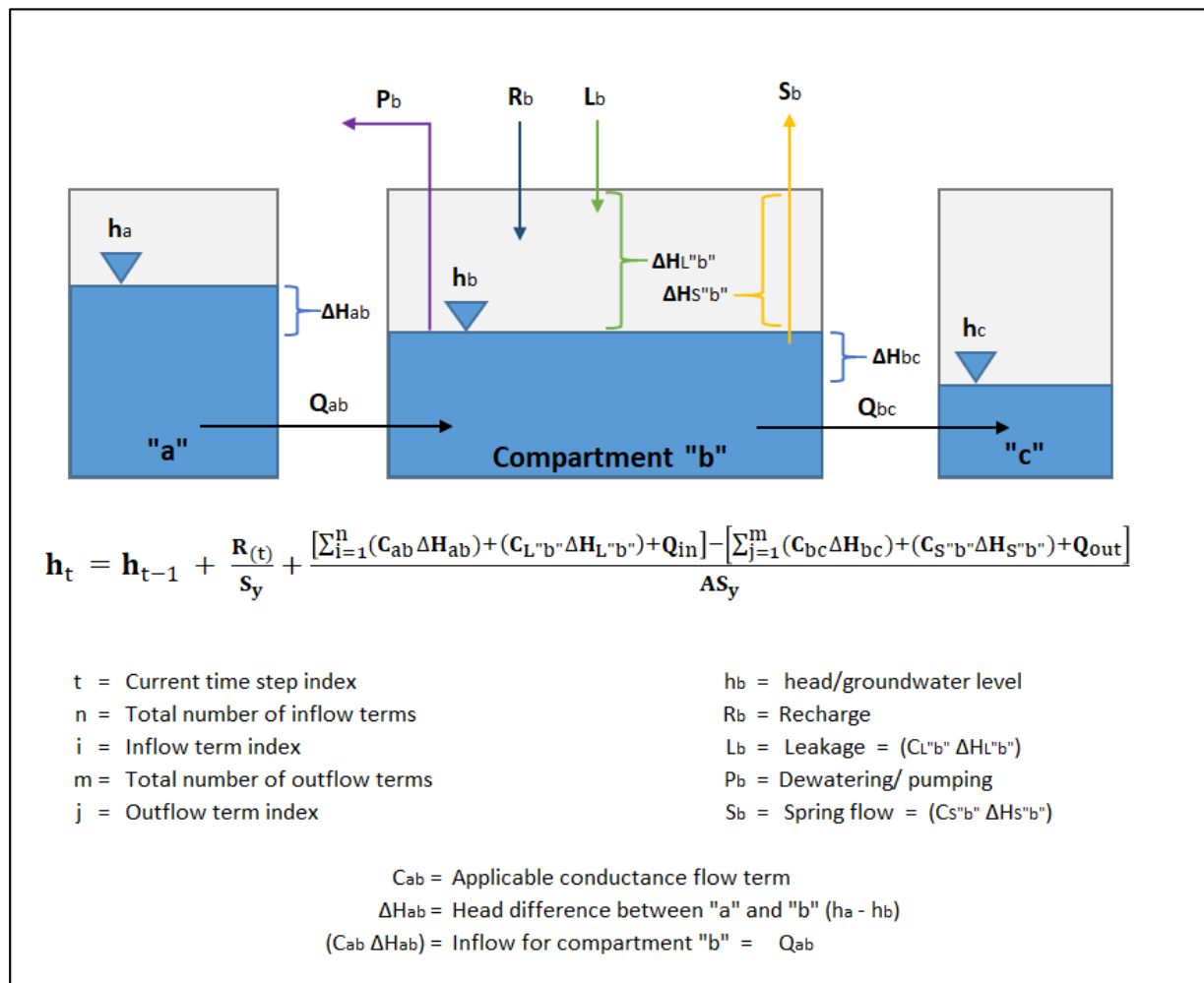


Figure 77: SVF - applied recovery equation.

The head dependent inflow and outflow for the intercompartmental flow terms (Q_{ab} & Q_{bc}) depend on the head difference between the compartment and the neighbouring compartment that serves as the inflow/ outflow point. The head-dependent leakage term is defined by the difference between the current groundwater level at time t_0 and the ground level at the associated compartment spring. As the water level recovers, the resulting head difference will become smaller resulting in lesser leakage driven by head difference. Similarly, the spring outflow was also driven by a conductance term as it can

be assumed constant for the unsaturated/dewatered zone within the applicable aquifer. Spring flow would therefore occur once the groundwater table reaches the ground level height of the spring.

Based on Figure 77, Equation (3.10) was implemented for each of the dewatered compartments within Microsoft Excel®. The compartments were directly linked to each other with continuous calculation occurring between each compartment based on the resulting t_1 values influenced across the model's simulation for estimating the t_0 water level for each compartment. Figure 78 includes the first year of recovery simulation for the Gembokfontein Compartment, directly affected by inflows from the assumed "full" Zuurbekom Compartment as well as outflows to the recovering Venterspost Compartment.

Gembokfontein Compartment "b"												
Date	ZUURBEKOM "a"		STUDYAREA		RECHARGE		INFLOWS		OUTFLOWS		RECOVERY	
	Water level BC00024 (mamsl)	ΔH_{ab} (m)	Study area (km ²)	Specific yield	Monthly Rainfall (mm/month)	Effective Recharge (ML/d)	Q_{ab} Inflows (ML/d)	L_b Leakage (ML/d)	Q_{bc} Outflows (ML/d)	S_b Springflow (ML/d)	Water level BH001111 (mamsl)	ΔH_{bc} (m)
	1575.2		84.95	0.1000		8.7%	0.9487	0.82	0.4566	1562.0	1539.80	
2050/01/01	1575.24	35.44	84950000	0.1000	159	39	33.6	18.0	9.2	0.0	1540.09	20.1
2050/02/01	1575.24	35.15	84950000	0.1000	300	74	33.3	17.7	9.2	0.0	1540.50	20.2
2050/03/01	1575.24	34.74	84950000	0.1000	163	40	33.0	17.4	9.3	0.0	1540.78	20.3
2050/04/01	1575.24	34.46	84950000	0.1000	85	21	32.7	17.2	9.3	0.0	1541.00	20.4
2050/05/01	1575.24	34.24	84950000	0.1000	99	24	32.5	17.1	9.4	0.0	1541.23	20.5
2050/06/01	1575.24	34.01	84950000	0.1000	41	10	32.3	16.9	9.4	0.0	1541.41	20.7
2050/07/01	1575.24	33.84	84950000	0.1000	0	0	32.1	16.8	9.5	0.0	1541.54	20.8
2050/08/01	1575.24	33.70	84950000	0.1000	0	0	32.0	16.7	9.6	0.0	1541.68	21.0
2050/09/01	1575.24	33.56	84950000	0.1000	0	0	31.8	16.6	9.7	0.0	1541.82	21.2
2050/10/01	1575.24	33.42	84950000	0.1000	0	0	31.7	16.5	9.8	0.0	1541.95	21.4
2050/11/01	1575.24	33.29	84950000	0.1000	13	3	31.6	16.3	9.8	0.0	1542.10	21.5
2050/12/01	1575.24	33.14	84950000	0.1000	121	30	31.4	16.1	9.9	0.0	1542.34	21.7

Figure 78: Gembokfontein Compartment "b" recovery simulation breakdown.

Where,

- Assumed "full" groundwater level of the non-dewatered Zuurbekom Compartment.
- Head difference between the groundwater level of Zuurbekom Compartment and the Gembokfontein Compartment at time step t_0 .
- Constant compartment characteristics.
- Forecasted future monthly rainfall (mm).
- Effective recharge inflow, percentage (%) rainfall from monthly forecasted value.
- Q_{ab} Inflow from Zuurbekom Compartment "a" based on the calculated conductance value of 0.95 as well as ΔH_{ab} .

- vii. Leakage inflows based on the groundwater level at time t_0 as well as the calculated conductance value of 0.82.
- viii. Q_{bc} Outflow to the Venterspost Compartment “c” based on the calculated conductance of 0.46 as well as ΔH_{bc} .
- ix. Spring flow based on the conductance term L_b multiplied by the head difference between the current groundwater level and the spring ground level of 1562 mamsl.
- x. Model simulated recovery groundwater level based on all parameters at time t_0 and the previous groundwater level at t_{-1} .
- xi. Head difference between the groundwater level of Gemsbokfontein Compartment and the Venterspost Compartment at the same time step.

7.4 Recovery Scenario Results

To observe the SVF-methods effectiveness in estimating groundwater level recovery within the dewatered compartments of the FWR, the modelling approach mentioned in Section 7.3 was used to simulate groundwater recovery. Recovery occurred within the dewatered dolomitic compartments: Gemsbokfontein, Venterspost, Bank and Oberholzer over a period of 100 years. Within this forecast window, each of the compartments recovered at different rates up to the spring ground level resulting in spring reactivation and spring flow across the entire FWR.

7.4.1.1 Gemsbokfontein Recovery

The following recovery results were obtained for the dewatered Gemsbokfontein Compartment based on the idealized flow conditions including inflow from the “full” non-dewatered Zuurbekom Compartment as well as recharge from precipitation and surface water leakages, Figure 79.

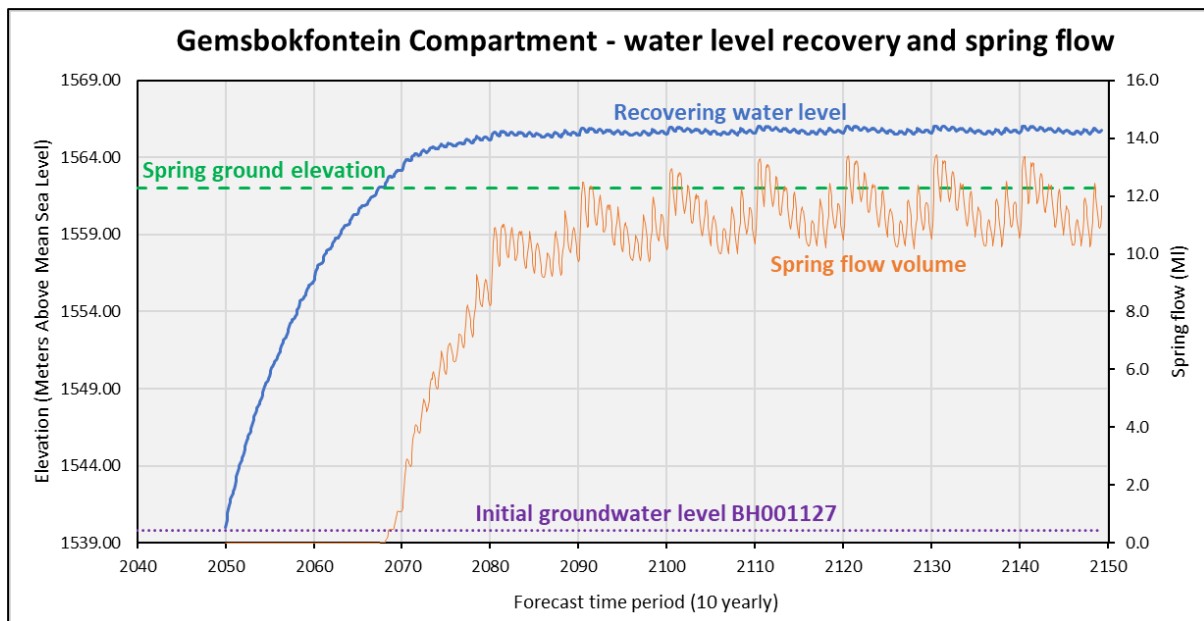


Figure 79: Gemsbokfontein Compartment groundwater recovery scenario.

The groundwater level of the Gemsbokfontein Spring, represented by BH001127 recovered from its initial groundwater level situated at 1539.80 mamsl to the spring surface elevation level defined at 1562 mamsl, the groundwater level further recovered up to 1565.5 mamsl where it remained in a steady-state condition. The total head of 22.2 meters recovered within a period of 17 years and 11 months with steady-state conditions reached after approximately 30 years. This recovery extent falls in between the 15-year estimate made by Dennis & Dennis (2017) during SVF-model calculations and the 35-year estimate made by their numerical model calculation based on a similar total head recovery. Initial high IGF flows between the Gemsbokfontein and Venterspost Compartments diminished recovery within the compartment but as the Venterspost Compartment recovered this IGF value slowly decreased with a marginal increase in recovery.

Spring flow slowly started to occur once the compartment was fully recovered up to the spring ground elevation of 1562 mamsl and steadily increased with an observed peak spring flow of 13.2 Ml/d during heavy rainfall months. Similar to the groundwater level, the spring flow rate reached steady-state conditions with fluctuations occurring as a direct result from wet and dry seasons from the forecasted rainfall data. An average spring flow of 11.4 Ml/d was obtained during the steady-state flow conditions.

The modelled spring flow of 11.4 Ml/d was higher than the original spring flow of 8.6 Ml/d during pre-mining conditions but lower than earlier flow estimates of 17 Ml/d as mentioned by Swart, *et al.* (2003a). Fitting within the same estimate of ± 12 Ml/d by Dennis & Dennis (2017) who conducted a similar SVF-modelled scenario solely on the Gemsbokfontein West Compartment.

7.4.1.2 Venterspost Recovery

At BH001636 a total head of 19.25 meters recovered within approximately 25 years after the recovery started. The groundwater table within the Venterspost Compartment recovered from an initial level of 1519.75 mamsl assumed from Section 5.4.2., Figure 80.

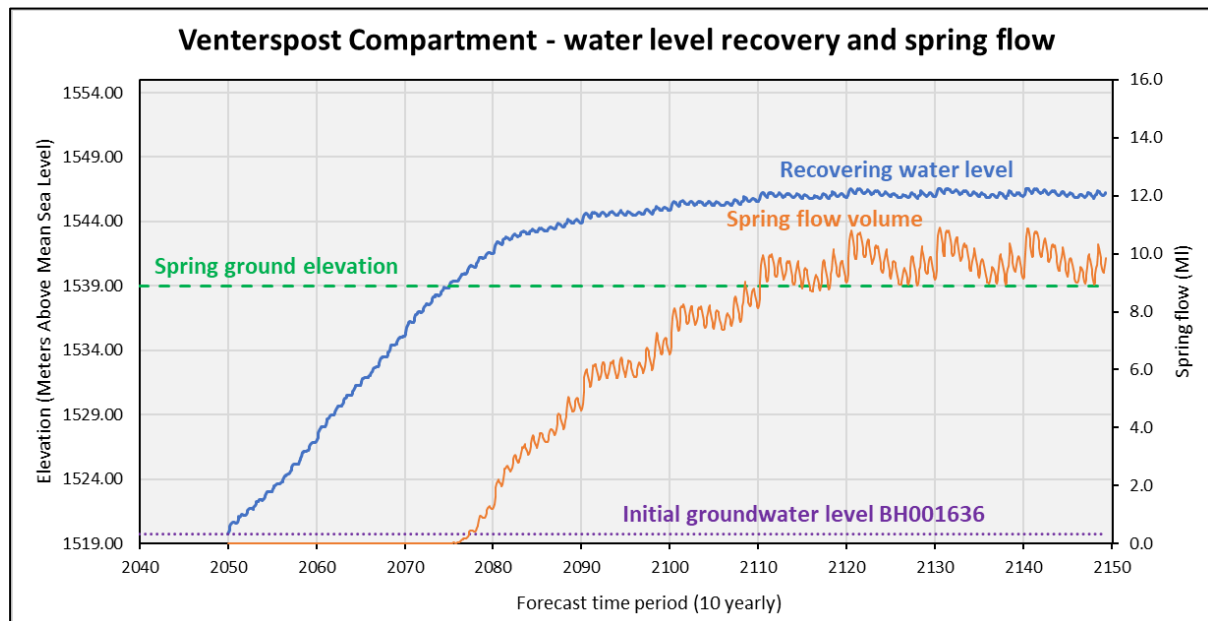


Figure 80: Venterspost Compartment groundwater recovery scenario.

The groundwater level within the Venterpost Compartment recovered up to the spring ground elevation of 1539.8 mamsl and further recovered up until a steady-state level of 1546 mamsl in equilibrium with the resulting spring flow with small fluctuations based on seasonal variability within the rainfall data. The compartment recovered fairly slow due to an abundance of IGF flow between the Venterspost Compartment and the Bank Compartment driven by the large head difference (100m during initial stages) between the two compartments. Even though the lateral IGF flow peaked at 17.1 Ml/d, the horizontal compartment's inflow amounted to 35.8 Ml/d during the same time step, resulting in $R > \text{IGF}$ (Section 5.2.6). The water level (h) therefore increased and the compartment filled up, resulting in a recovered compartment as opposed to lower levels resulting from a mega-compartment scenario with water levels up to 70m below that of the ground level as predicted by Wolmarans (1984) as mentioned by Schrader, *et al.* (2014b).

Once the compartment recovered beyond the relative spring's ground elevation, spring flow started to occur and increased steadily up to a steady-state average of 9.8 Ml/d. The spring flow fluctuated according to rainfall fluctuation up to a peak spring flow of 10.8 Ml/d. The recovery scenario spring flow was significantly lower than the original 20.9 Ml/d and half the

20 Ml/d estimated by Schrader, *et al.* (2014c). It should be noted that the spring flow estimates made by Schrader, *et al.* (2014c) ignored the effects IGF would have on the spring flow volumes. The considerable high IGF that occurs between the Venterspost Compartment and the Bank Compartment within the recovery scenario largely caused the lower flow observed at the Venterspost Spring.

7.4.1.3 Bank Recovery

The recovery scenario of the Bank Compartment lasted 57 years for a total head recovery of 81.4m. Represented by BH000405, the groundwater level at 1420.6 mamsl recovered up to the spring elevation level at 1502 mamsl. Figure 81 includes the groundwater level recovery for the Bank Compartment as well as the resulting spring flow volumes.

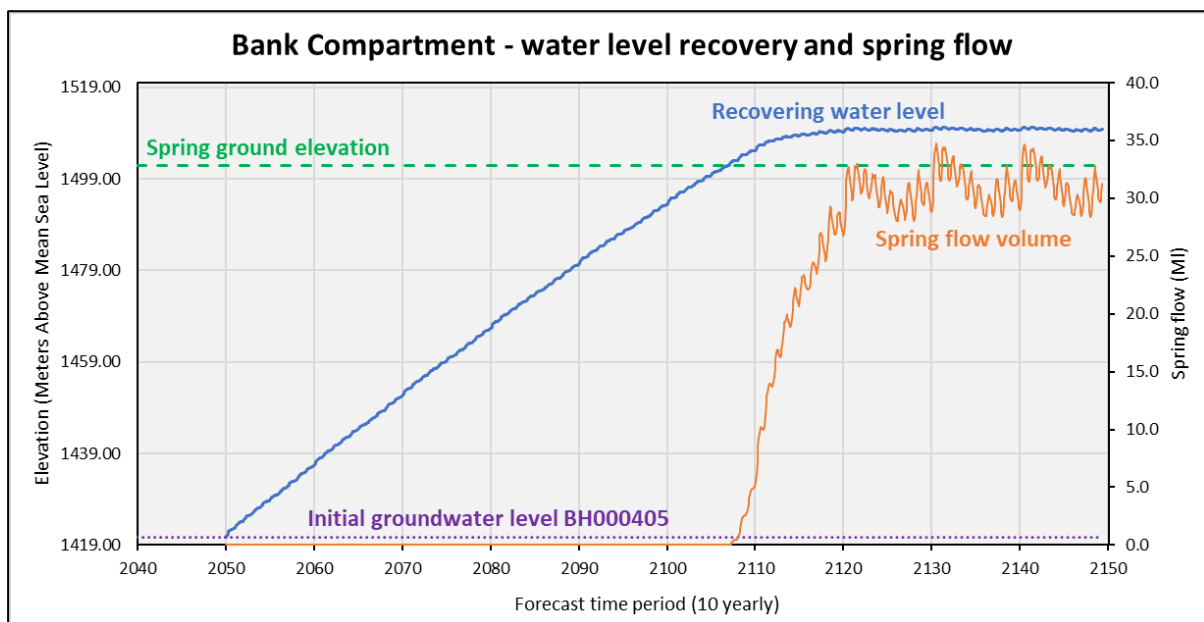


Figure 81: Bank Compartment groundwater recovery scenario.

Compared to the Gemsbokfontein and Venterspost Compartments, the Bank Compartment recovered with the fastest rate of 1.42m/year against 1.23m/year and 0.77m/year respectively. The large recovery rate was mostly due to the large head difference between the Venterspost- and Bank Compartment with large inflows from IGF aiding the major contributing effective recharge inflow. With the initial groundwater level within the Oberholzer Compartment 4m below that of Bank, a lower IGF between these two compartments resulted in the fast recovery of the Bank Compartment taking into account that the compartment is three times the size of the Venterspost Compartment. The groundwater level reached its steady-state conditions at 1510 mamsl.

Spring flow occurred once the ground elevation of the Bank Spring was reached, this resulted in a steady flow increase until it reached equilibrium with the groundwater level. The resulting steady-state spring flow yielded an average of 31.4 MI/d. The resulting scenario spring flow was lower than the original spring flow of 49.1 MI/d but closer to the estimation by Schrader, *et al.* (2014c) taking into account that Schrader, *et al.* (2014c) ignored the effects from IGF as previously mentioned.

7.4.1.4 Oberholzer Eye

Similar to the above-mentioned compartments, the groundwater level within the Oberholzer recovered up to the ground elevation of its associated spring within approximately 30-years and 4-months. The groundwater level recovered from the dewatered level represented by BH-2726AD00093 at 1416.5 mamsl up to the Oberholzer Springs ground elevation at 1471 mamsl amounting to a total head recovery of 54.5m, Figure 82.

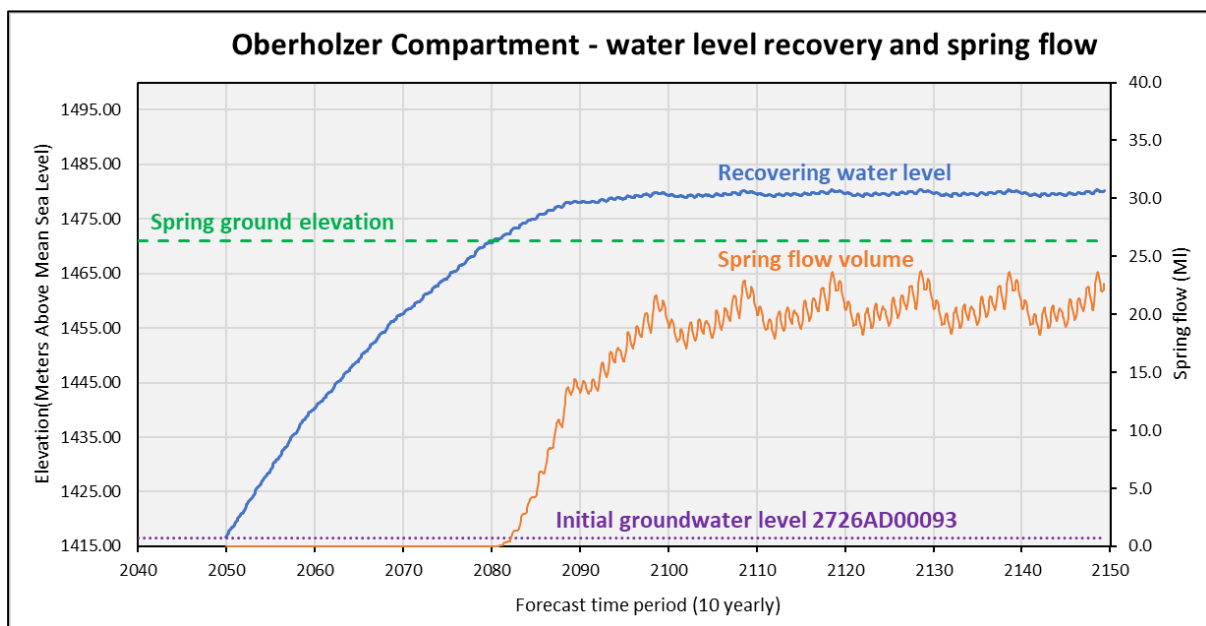


Figure 82: Oberholzer Compartment groundwater recovery.

Unlike the previously mentioned compartments, the Oberholzer Compartments' recovery scenario was based on a forecasted future rainfall data set based on historical data obtained from rainfall station 0447680W, Section 5.5. During recovery, not only were there IGF inflows from the neighbouring Bank Compartment but also from the downstream "full" non-dewatered Turffontein Compartment. These dual IGF inflows resulted in the rapid recovery of the Oberholzer Compartment at 1.84m/year.

Spring reactivation occurred once the groundwater level reached the spring elevation level and with flow steadily increasing up to steady-state conditions. The resulting spring flow fluctuated

in response to the rainfall fluctuations that occur within the future forecasted rainfall. An average steady-state spring flow rate of 20.6 ML/d was simulated for the Oberholzer Spring. As with the Venterspost and Bank Compartments, the recovery scenarios resulting spring flow was significantly lower than pre-mining measured flow of 54.1 ML/d.

7.5 Conclusion

During the recovery scenarios, each of the SVF water budgets fluctuated according to the main inputs and outputs with seasonal rainfall variability affecting recovery and spring flow the most. Although the IGF scenarios did affect the rewatering process, the IGF that occurred between compartments was significantly lower than the larger inflow volumes into each compartment. This led to a faster vertical rise in the groundwater level rather than the horizontal IGF flow ($R > IGF$) as depicted by Schrader *et al.* (2014c) in Figure 41, Section 5.2.6.

As a result, each of the rewatering scenarios supported the notion of Schrader *et al.* (2014c) that the groundwater levels within all the dewatered dolomitic compartments would restore to near pre-mining levels prompting spring flow to occur. A mega-compartment, where water levels would not fully recover within the dewatered compartment rendering most of the springs dry was less likely to occur during the recovery scenarios, Table 34.

Most of the compartments recovered within the estimated periods by Dennis & Dennis (2017) and Usher & Scott (2001). To a lesser extent, the resulting spring flow within most of the compartments were largely underestimated compared to the original spring flow values as well as those by Swart, *et al.*, (2003a) and Schrader, *et al.*, (2014c). Whilst keeping in mind that the estimations and calculations by both Swart, *et al.*, (2003a) and Schrader, *et al.*, (2014c) were based on pre-mining aquifer conditions as well as the remark by Schrader, *et al.*, (2014c) that they ignored all effects by IGF. Therefore in part these differences could contribute to the fact that the IGF will be more prominent in post-mining times based on the obtained calibration parameters within Chapter 6, resulting in lower spring flow volumes across the FWR.

A number of shortcomings were observed within the created recovery SVF-model. The largest of these included the generalisation and grouping of the parameters as a result of absence or general lack in data availability. Assumptions mentioned in Section 6.1.2 and Section 7.3 led to oversimplification of the study area. This in conjunction with the unfavourable use of different calibration boreholes to recovery scenario boreholes due to a lack in data availability could also explain the overall under-estimations that occurred.

Regardless of the volumetric outcomes, overall the SVF water budget model simulated groundwater level recovery together with net spring flows characteristic of karst spring systems as seen in Figure 10, Section 3.2.1.

Table 34: Predicted recovery results after mining activities have stopped in the FWR.

		Compartment			
		Gemsbokfontein	Venterspost	Bank	Oberholzer
Rewatering estimates (years)	Usher & Scott (1) (2001) (2)	-	0.42	18.3	15.2
		54	25	30	11
	Van Biljon (3) (2016)	6.5	-	-	-
	Dennis & (4) Dennis (2017)	15	-	-	-
	(5)	35	-	-	-
Recovery Scenario		18	25	57	30
Pre-mining ave. spring flow		8.64	20.9	49.1	54.1
Predicted spring flow volumes (MI/d)	Wolmarans (6) (1984)	0	0	0	0
	Swart, <i>et al.</i> , (2003a)	8.64	20.9	49.1	54.1
	Schrader, <i>et al.</i> , (7) (2014c)	-	20	43.6	53.3
	Van Biljon (3) (2016)	12.8	-	-	-
	Dennis & (4) Dennis (2017)	12 to 13	-	-	-
	(5)	12	-	-	-
Recovery Scenario		11.4	9.8	31.4	20.6
(1) - based on mine records		(4) - based on SVF model		(6) - based on mega compartment	
(2) - based on spring flows		(5) - based on numerical model		(7) - regardless of IGF	
(3) - as mentioned by De Klerk (2019)					

CHAPTER 8

CONCLUSIONS AND RECOMMENDATIONS

8.1 Conclusions

The FWR has undergone various transformations over millions of years, including the latest transformations caused by man-made activities. These natural and man-made effects have caused the Malmani Dolomites to develop complex structures and features resulting in karst aquifers with great uncertainties with respect to its hydrogeological parameters. The complexity of the FWR ranges from well-developed micro-fissures and preferential groundwater flow pathways to regional sub-divided dolomitic compartments. These dolomitic compartments divided by impermeable dykes not only function individually but also directly affect each other. This resulted as mining activities pierced through the Syenite dykes, leading to these dolomitic compartments becoming hydrologically connected. The constant dewatering within the FWR has left the landscape unstable with accelerated sinkhole formation and the subsequent lowered groundwater table caused the dolomitic springs to dry up.

The factors associated with the karst aquifers have caused great uncertainty regarding the estimation of effective recharge and other aquifer parameters in the FWR. Throughout the years there have been several studies and techniques implemented which aimed at understanding these complex systems to not only comprehend their current workings but also their foreseeable future, a future where mine related dewatering has ceased allowing these dolomitic aquifers to recover. Therefore, modelling of these dolomitic compartments is essential in estimating the rewatering process, spring reactivation as well as determining the interactions between each of the dewatered compartments.

Detailed research was conducted on the FWR to understand the geological characteristics that led to the formation of the mass dolomitic rock and the subsequent change thereof into the karst system we know today. From previous studies it was noted that each sub-divided dolomitic compartment functioned individually, essentially evolving through karstification and sinkhole development based on their regional activities. This led to each compartment having unique aquifer parameters including (but not limited to) recharge, storage and specific yield. Although extensive investigations were done within the FWR due to the ongoing mining activities, most of this data is not accessible, leading to a lack of data available to the general public.

Based on its lumped parameters approach, the SVF method was selected and used for mathematically modelling the dewatered compartments of the FWR in order to determine whether a water budget model could be used to simulate groundwater level recovery and spring reactivation. Taking into account a list of assumptions, in total six SVF water budget equations were created and calibrated based on the current aquifer conditions. These compartments included the dewatered Gemsbokfontein, Venterspost, Bank and Oberholzer Compartments as well as their direct upstream and downstream neighbouring compartments: Zuurbekom and Turffontein. Each of these water budgets was linked to each other's inflows and outflows and calibrated against observed water levels within the monitoring boreholes to establish each compartment's water budget components and parameters.

The implemented SVF water budget equations enabled easy changes within the calibration parameters for rapid calibration. Showcasing its adaptiveness by allowing simplification within the data inputs to offset any lack in input data. From the calibration results and sensitivity analysis, it was clear that effective recharge % was most sensitive to change, followed by the grouped leakage inflows and the IGF outflows. The estimated effective recharge for the dolomitic compartments varied between 6.0% and 9.3%, with higher values for recharge and specific yield values obtained for the compartments that were most affected by the dewatering activities. This suggests progressive karstification within these dewatered compartments directly affects the aquifer parameters. Each of the six compartments was calibrated with satisfactory results, obtaining Pearson Correlation values between the simulated and observed water levels rarely below 0.94 and NCS values upwards of 0.78. The calculated RMSE averaged around 0.073m with the only significant outlier the calibration of the Turffontein Compartment with an obtained RMSE of 0.15m due to the additional complexity of the compartments natural water balance.

The adaptiveness of the SVF equation was proved once again during the recovery scenario. In order to showcase that a water budget model can simulate groundwater recovery and spring flow. The resulting water budgets for each of the four dewatered dolomitic compartments were adapted to simulate groundwater level recovery through the pre-determined inputs and outputs. With the main focus being the exclusion of outflows by dewatering. Once a compartment's groundwater level fully recovered to the spring ground elevation level, the net outflow governed by an assigned conductance was characterised as spring discharge.

The recovery process included the use of forecasted future rainfall data for a 100-year period, within this forecasted period dry and wet spells occurred at random with major events every 10 years. The recovery phase started simultaneously within all the dewatered compartments with each recovery from an assumed dewatered water level based on the obtained monitoring

borehole data. Each of the compartment's groundwater tables rewatered to their respective spring elevation levels at different rates. Recovery lasted between 18 years for the Gembokfontein Compartment and 57 years for the Bank Compartment, additionally the Venterspost Compartment and Oberholzer Compartment recovery within 25 and 30 years respectively. The recovery time depended on several local and regional features including the characterised effective recharge for each compartment as well as the different IGF volumes driven by the difference in head between neighbouring compartments.

Full recovery as opposed to the formation of a mega-compartment was mainly due to the vertical recharge occurring within each compartment being more profound than the horizontal IGF leakages. Once groundwater levels within these dolomitic compartments reached the ground elevation level for each of their associated springs, the springs reactivated. The resulting spring flows, characteristic to that of karst springs fluctuated accounting the seasonal variability within the forecasted rainfall. For the Gembokfontein Compartment, the resulting simulated spring flow was within the predicted range with an average steady-state flow volume of 11.4 Ml/d above that of the pre-mining volume of 8.6 Ml/d. The spring flow for the remaining compartments were well below that of previous estimations as well as the pre-mining volumes: 9.8 Ml/d for the Venterspost Compartment as opposed to 20.9 Ml/d; 31.4 Ml/d for the Bank Compartment as opposed to 49.1 Ml/d and 20.6 Ml/d for the Oberholzer Compartment as opposed to 54.1 Ml/d. This noticeable difference could either be due to the model's oversimplification that led to underestimations or indicate that the increased IGF caused by the mining activities within the FWR subsequently led to more uniform spring flow across the valley.

The SVF model performed as expected, each water balance simulated the additional net-inflows to raise the groundwater levels simulating aquifer recovery. As a tool for obtaining first-order estimates the SVF water balance method demonstrated to be a fast and simplistic method for the estimation of groundwater level recovery and spring reactivation using limited data. This however can also be seen as the SVF-method's biggest downfall, as a lack in data availability can lead to oversimplicity that comes with a range of limitations and assumptions.

The study proved that a simplistic water budget such as the Saturated Volume Fluctuation method (SVF) can be utilised to estimate groundwater level recovery within a post-mining rewatering scenario using limited data with a set of assumptions and lumping parameters together. But, in order to obtain accurate predictions as to the foreseeable future of the FWR, more relevant and accurate data needs to be obtained in order to more accurately model and calibrate the study area.

8.2 Recommendations

From this study, the following recommendations can be made:

1. The study faced many uncertainties regarding the dolomitic aquifer's parameters, with parameter lumping and assumptions enabling successful modelling. The availability of good quality, accurate data across the study posed many challenges. Future research should make use of the most recent data available or conduct hydrogeological investigations to further determine area characteristics.
2. A more heterogenic approach can be implemented for each dolomitic compartment where aquifer parameters are set to change according to spatial variability with several sub-water balance equations set up to represent different zones within a single compartment.
3. The FWR is a complex system that not only consists of hydrogeological aspects. Therefore, a fully integrated and distributed model for the area should be developed that includes the interactions between the hydrological conditions, hydrogeological conditions and all mining-related impacts.

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APPENDIX A: SVF CALIBRATION MODELS

ZUURBEKOM "a" LAG 2months										
Date	STUDY AREA			RECHARGE		INFLOW	OUTFLOWS			ΔH _{ab} (m)
	Observed BC00024 (mamsl)	Study Area (km ²)	Specific Yield	Monthly Rainfall (mm/month)	Effective Recharge (MI/d)	La Leakage (MI/d)	Rand water Pumping	Qab Outflows (MI/d)	Simulated Water Level	
	1587	100.23	0.0270		9.3%	2.9	3.6	21.5	1575.55	1552.59
Oct-03	1575.32	100230000	0.0270	11	3	2.9	3.6	21.5	1575.34	22.76
Nov-03	1575.12	100230000	0.0270	9	3	2.9	3.6	21.5	1575.13	22.54
Dec-03	1575.06	100230000	0.0270	45	14	2.9	3.6	21.5	1575.03	22.45
Jan-04	1575.01	100230000	0.0270	99	14	2.9	3.6	21.5	1574.94	22.35
Feb-04	1574.95	100230000	0.0270	80	31	2.9	3.6	21.5	1575.04	22.45
Mar-04	1575.15	100230000	0.0270	155	25	2.9	3.6	21.5	1575.07	22.48
Apr-04	1575.17	100230000	0.0270	148	48	2.9	3.6	21.5	1575.35	22.77
May-04	1575.62	100230000	0.0270	55	46	2.9	3.6	21.5	1575.62	23.03
Jun-04	1575.57	100230000	0.0270	42	17	2.9	3.6	21.5	1575.56	22.97
Jul-04	1575.52	100230000	0.0270	0	13	2.9	3.6	21.5	1575.46	22.87
Aug-04	1575.16	100230000	0.0270	3	0	2.9	3.6	21.5	1575.21	22.62
Average	1575.24	100230000	0.0270	69.6	19.4	2.94	3.6	21.5	1575.25	22.66

GEMSBOKFONTEIN "b" LAG 2months												
Date	STUDY AREA			RECHARGE		INFLOWS		OUTFLOWS		SIMULATION		
	Observed BC00052 (mamsl)	Study area (km ²)	Specific yield	Monthly Rainfall (mm/month)	Effective Recharge (MI/d)	Qab Inflows (MI/d)	Lb Leakage (MI/d)	Qbc Outflows (MI/d)	Mine Outflows (MI/d)	Simulated Water Level	ΔH _{bc} (m)	Ground level ΔH (m)
	1585.4	84.95	0.10		8.7%	21.50	26.95	1.10	77.0	1552.90	1550.18	1585.4
Oct-03	1552.84	84950000	0.10	11	3	21.5	27.0	1.1	77.0	1552.80	2.6	32.6
Nov-03	1552.72	84950000	0.10	9	2	21.5	27.0	1.1	77.0	1552.71	2.5	32.7
Dec-03	1552.66	84950000	0.10	45	11	21.5	27.0	1.1	77.0	1552.64	2.5	32.8
Jan-04	1552.52	84950000	0.10	99	24	21.5	27.0	1.1	77.0	1552.62	2.4	32.8
Feb-04	1552.59	84950000	0.10	80	20	21.5	27.0	1.1	77.0	1552.59	2.4	32.8
Mar-04	1552.57	84950000	0.10	155	38	21.5	27.0	1.1	77.0	1552.62	2.4	32.8
Apr-04	1552.62	84950000	0.10	148	36	21.5	27.0	1.1	77.0	1552.64	2.5	32.8
May-04	1552.60	84950000	0.10	55	14	21.5	27.0	1.1	77.0	1552.59	2.4	32.8
Jun-04	1552.53	84950000	0.10	42	10	21.5	27.0	1.1	77.0	1552.52	2.3	32.9
Jul-04	1552.44	84950000	0.10	0	0	21.5	27.0	1.1	77.0	1552.41	2.2	33.0
Aug-04	1552.35	84950000	0.10	3	1	21.5	27.0	1.1	77.0	1552.31	2.1	33.1
Average	1552.59	84950000	0.10	69.6	14.5	21.50	26.95	1.1	77.0	1552.59	2.4	32.8

VENTERSPOST "c" LAG 2months												
Date	STUDY AREA			RECHARGE		INFLOWS		OUTFLOWS		SIMULATION		
	Observed BH002459 (mamsl)	Study (km ²)	Specific yield	Monthly Rainfall (mm/month)	Effective Recharge (MI/d)	Qbc Inflows (MI/d)	Lc Leakage (MI/d)	Qcd Outflows (MI/d)	Mine Outflows (MI/d)	Simulated Water Level	ΔH _{bc} (m)	Ground level ΔH (m)
	1586.9	54.4	0.0650		7.9%	1.1	7.0	2.2	28.0	1550.86	1562.62	1586.90
Oct-03	1550.69	54400000	0.0650	11	2	1.1	7.0	2.2	28.0	1550.69	-11.9	36.2
Nov-03	1550.63	54400000	0.0650	9	1	1.1	7.0	2.2	28.0	1550.51	-12.1	36.4
Dec-03	1550.54	54400000	0.0650	45	6	1.1	7.0	2.2	28.0	1550.38	-12.2	36.5
Jan-04	1550.47	54400000	0.0650	99	14	1.1	7.0	2.2	28.0	1550.31	-12.3	36.6
Feb-04	1550.32	54400000	0.0650	80	11	1.1	7.0	2.2	28.0	1550.22	-12.4	36.7
Mar-04	1550.23	54400000	0.0650	155	22	1.1	7.0	2.2	28.0	1550.22	-12.4	36.7
Apr-04	1550.24	54400000	0.0650	148	21	1.1	7.0	2.2	28.0	1550.21	-12.4	36.7
May-04	1549.99	54400000	0.0650	55	8	1.1	7.0	2.2	28.0	1550.09	-12.5	36.8
Jun-04	1549.94	54400000	0.0650	42	6	1.1	7.0	2.2	28.0	1549.96	-12.7	36.9
Jul-04	1549.77	54400000	0.0650	0	0	1.1	7.0	2.2	28.0	1549.77	-12.8	37.1
Aug-04	1549.62	54400000	0.0650	3	0	1.1	7.0	2.2	28.0	1549.59	-13.0	37.3
Average	1550.22	54400000	0.0650	69.6	10.0	1.10	7.00	2.2	28.0	1550.18	12.4	36.7

BANK"d" LAG 2months												
Date	STUDY AREA			RECHARGE		INFLOWS		OUTFLOWS		SIMULATION		
	Observed BH001218A (mamsl)	Study (km2)	Specific	Monthly Rainfall (mm/month)	Effective Recharge (ML/d)	Qcd Inflows (ML/d)	Ld Leakage (ML/d)	Qde Outflows (ML/d)	Mine Outflows (ML/d)	Simulated Water Level	ΔHbc (m)	Ground level ΔH (m)
	1576.4	156.7	0.0630		8.5%	2.2	7	6.156	60.0	1563.24	1437.12	1576.40
Oct-03	1563.17	156700000	0.0630	11	5	2.2	7.0	6.2	60.0	1563.08	126.0	13.3
Nov-03	1563.04	156700000	0.0630	9	4	2.2	7.0	6.2	60.0	1562.92	125.8	13.5
Dec-03	1562.95	156700000	0.0630	45	20	2.2	7.0	6.2	60.0	1562.81	125.7	13.6
Jan-04	1562.90	156700000	0.0630	99	44	2.2	7.0	6.2	60.0	1562.77	125.7	13.6
Feb-04	1562.70	156700000	0.0630	80	36	2.2	7.0	6.2	60.0	1562.70	125.6	13.7
Mar-04	1562.69	156700000	0.0630	155	69	2.2	7.0	6.2	60.0	1562.74	125.6	13.7
Apr-04	1562.74	156700000	0.0630	148	65	2.2	7.0	6.2	60.0	1562.76	125.6	13.6
May-04	1562.57	156700000	0.0630	55	25	2.2	7.0	6.2	60.0	1562.67	125.5	13.7
Jun-04	1562.50	156700000	0.0630	42	18	2.2	7.0	6.2	60.0	1562.55	125.4	13.9
Jul-04	1562.38	156700000	0.0630	0	0	2.2	7.0	6.2	60.0	1562.38	125.3	14.0
Aug-04	1562.21	156700000	0.0630	3	1	2.2	7.0	6.2	60.0	1562.21	125.1	14.2
Average	1562.71	156700000	0.0630	69.6	30.9	2.15	7.00	6.2	60.0	1562.62	125.6	13.7

OBERHOLZER"e" LAG 3months												
Date	STUDY AREA			RECHARGE		INFLOWS		OUTFLOWS		SIMULATION		
	Observed BH001712 (mamsl)	Study (km2)	Specific	Monthly Rainfall (mm/month)	Effective Recharge (ML/d)	Qde Inflows (ML/d)	Le Leakage (ML/d)	Qef Outflows (ML/d)	Mine Outflows (ML/d)	Simulated Water Level	ΔHbc (m)	Ground level ΔH (m)
	1509.8	153.8	0.0520		8.2%	6.2	20	16.1	50.0	1437.00	1479.80	1509.80
Oct-03	1436.96	153800000	0.0520	18	8	6.2	20.0	16.1	50.0	1436.88	-42.9	72.9
Nov-03	1436.87	153800000	0.0520	17	7	6.2	20.0	16.1	50.0	1436.76	-43.0	73.0
Dec-03	1436.85	153800000	0.0520	214	90	6.2	20.0	16.1	50.0	1436.94	-42.9	72.9
Jan-04	1436.80	153800000	0.0520	48	20	6.2	20.0	16.1	50.0	1436.87	-42.9	72.9
Feb-04	1436.82	153800000	0.0520	120	50	6.2	20.0	16.1	50.0	1436.91	-42.9	72.9
Mar-04	1437.18	153800000	0.0520	315	132	6.2	20.0	16.1	50.0	1437.25	-42.5	72.5
Apr-04	1437.18	153800000	0.0520	114	48	6.2	20.0	16.1	50.0	1437.28	-42.5	72.5
May-04	1437.37	153800000	0.0520	134	56	6.2	20.0	16.1	50.0	1437.34	-42.5	72.5
Jun-04	1437.33	153800000	0.0520	56	24	6.2	20.0	16.1	50.0	1437.28	-42.5	72.5
Jul-04	1437.21	153800000	0.0520	7	3	6.2	20.0	16.1	50.0	1437.14	-42.7	72.7
Aug-04	1436.99	153800000	0.0520	22	9	6.2	20.0	16.1	50.0	1437.03	-42.8	72.8
Average	1437.08	153800000	0.0520	114.3	48.0	6.16	20.00	16.1	50.0	1437.12	-42.7	72.7

TURFFONTEIN "f" LAG 4months												
Date	STUDY AREA			RECHARGE		INFLOWS		OUTFLOWS			SIM	
	Observed BH001712 (mamsl)	Study Area (km2)	Specific Yield	Monthly Rainfall (mm/month)	Effective Recharge (ML/d)	Qef Inflows (ML/d)	Lf Leakage ML/d	Outflows (ML/d)	GMB Spring (ML/d)	Boskop Spring (ML/d)	Turffontein Springs (ML/d)	Water Level (mamsl)
	1408.8	704.5	0.0300		6.0%	16.1	3	28.6	52.7	45.0	17.4	1479.38
Oct-03	1479.34	704500000	0.0300	17	24	16.1	3.0	28.6	52.7	45.0	17.4	1479.24
Nov-03	1479.20	704500000	0.0300	214	302	16.1	3.0	28.6	52.7	45.0	17.4	1479.49
Dec-03	1479.16	704500000	0.0300	48	68	16.1	3.0	28.6	52.7	45.0	17.4	1479.41
Jan-04	1479.13	704500000	0.0300	120	170	16.1	3.0	28.6	52.7	45.0	17.4	1479.47
Feb-04	1479.66	704500000	0.0300	315	444	16.1	3.0	28.6	52.7	45.0	17.4	1479.93
Mar-04	1479.89	704500000	0.0300	114	160	16.1	3.0	28.6	52.7	45.0	17.4	1479.98
Apr-04	1479.93	704500000	0.0300	134	189	16.1	3.0	28.6	52.7	45.0	17.4	1480.07
May-04	1479.84	704500000	0.0300	56	79	16.1	3.0	28.6	52.7	45.0	17.4	1480.00
Jun-04	1479.81	704500000	0.0300	7	10	16.1	3.0	28.6	52.7	45.0	17.4	1479.84
Jul-04	1479.41	704500000	0.0300	22	31	16.1	3.0	28.6	52.7	45.0	17.4	1479.71
Aug-04	1479.31	704500000	0.0300	0	0	16.1	3.0	28.6	52.7	45.0	17.4	1479.53
Average	1479.60	704500000	0.0300	101.8	143.9	16.10	3.00	28.6	52.7	45.0	17.4	1479.80

APPENDIX B: SVF RECOVERY SCENARIOS

GEMSBOKFONTEIN "B" LAG 3														
Date	ZUURBEKOM		STUDYAREA		RECHARGE		INFLOWS			OUTFLOWS			RECOVERY	
	Water level BC00024 (mamsl)	"BC" ΔH (m)	Study (km2)	Specific	Monthly Rainfall (mm/month)	Effective Recharge (ML/d)	C "AB"	Q "AB" Inflows (ML/d)	L "B" Leakage (ML/d)	C "BC"	Q"BC" Outflows (ML/d)	spring (ML/d)	Water level (mamsl)	"BC" ΔH (m)
	1575.2		84.95	0.1000		8.7%	0.9487	note	0.82	0.4566		1562.0	1539.80	
2050/01/01	1575.24	35.44	84950000	0.1000	55	13	0.95	33.6	18.1	0.46	9.2	0.0	1540.00	20.1
2050/02/01	1575.24	35.24	84950000	0.1000	108	27	0.95	33.4	17.9	0.46	9.2	0.0	1540.24	20.2
2050/03/01	1575.24	35.00	84950000	0.1000	159	39	0.95	33.2	17.6	0.46	9.3	0.0	1540.53	20.4
2050/04/01	1575.24	34.72	84950000	0.1000	300	74	0.95	32.9	17.3	0.46	9.4	0.0	1540.93	20.5
2050/05/01	1575.24	34.31	84950000	0.1000	163	40	0.95	32.5	17.1	0.46	9.4	0.0	1541.22	20.6
2050/06/01	1575.24	34.03	84950000	0.1000	85	21	0.95	32.3	16.9	0.46	9.4	0.0	1541.43	20.7
2050/07/01	1575.24	33.81	84950000	0.1000	99	24	0.95	32.1	16.7	0.46	9.5	0.0	1541.65	20.8
2050/08/01	1575.24	33.59	84950000	0.1000	41	10	0.95	31.9	16.6	0.46	9.6	0.0	1541.83	21.0
2050/09/01	1575.24	33.41	84950000	0.1000	0	0	0.95	31.7	16.5	0.46	9.6	0.0	1541.96	21.1
2050/10/01	1575.24	33.28	84950000	0.1000	0	0	0.95	31.6	16.3	0.46	9.7	0.0	1542.10	21.3
2050/11/01	1575.24	33.14	84950000	0.1000	0	0	0.95	31.4	16.2	0.46	9.8	0.0	1542.23	21.5
2050/12/01	1575.24	33.01	84950000	0.1000	0	0	0.95	31.3	16.1	0.46	9.9	0.0	1542.36	21.6
5th year of simulation														
2055/12/01	1575.24	24.26	84950000	0.1000	49	12	0.95	23.0	9.0	0.46	12.5	0.0	1551.09	27.4
10th year of simulation														
2060/12/01	1575.24	18.12	84950000	0.1000	0	0	0.95	17.2	4.0	0.46	13.3	0.0	1557.15	29.1
15th year of simulation														
2065/12/01	1575.24	14.37	84950000	0.1000	47	12	0.95	13.6	0.9	0.46	13.2	0.0	1560.92	29.0
20th year of simulation														
2070/12/01	1575.24	11.42	84950000	0.1000	0	0	0.95	10.8	-1.5	0.46	12.6	2.7	1563.80	27.6
25th year of simulation														
2075/12/01	1575.24	10.48	84950000	0.1000	48	12	0.95	9.9	-2.3	0.46	11.6	6.3	1564.76	25.4
30th year of simulation														
2080/12/01	1575.24	9.76	84950000	0.1000	0	0	0.95	9.3	-2.8	0.46	10.6	9.8	1565.43	23.2
35th year of simulation														
2085/12/01	1575.24	9.89	84950000	0.1000	50	12	0.95	9.4	-2.8	0.46	10.1	9.2	1565.35	22.1
40th year of simulation														
2090/12/01	1575.24	9.52	84950000	0.1000	0	0	0.95	9.0	-3.0	0.46	9.7	11.2	1565.67	21.3
45th year of simulation														
2095/12/01	1575.24	9.78	84950000	0.1000	50	12	0.95	9.3	-2.8	0.46	9.6	9.9	1565.46	21.0
50th year of simulation														
2100/12/01	1575.24	9.45	84950000	0.1000	0	0	0.95	9.0	-3.1	0.46	9.4	11.6	1565.74	20.5
55th year of simulation														
2105/12/01	1575.24	9.74	84950000	0.1000	48	12	0.95	9.2	-2.9	0.46	9.3	10.1	1565.50	20.3
60th year of simulation														
2110/12/01	1575.24	9.40	84950000	0.1000	0	0	0.95	8.9	-3.1	0.46	9.1	11.9	1565.79	19.9
65th year of simulation														
2115/12/01	1575.24	9.69	84950000	0.1000	50	12	0.95	9.2	-2.9	0.46	9.0	10.3	1565.55	19.8
70th year of simulation														
2120/12/01	1575.24	9.38	84950000	0.1000	0	0	0.95	8.9	-3.1	0.46	8.9	12.0	1565.80	19.6
75th year of simulation														
2125/12/01	1575.24	9.70	84950000	0.1000	47	12	0.95	9.2	-2.9	0.46	9.0	10.3	1565.54	19.7
80th year of simulation														
2130/12/01	1575.24	9.38	84950000	0.1000	0	0	0.95	8.9	-3.1	0.46	8.9	12.1	1565.81	19.5
85th year of simulation														
2135/12/01	1575.24	9.69	84950000	0.1000	50	12	0.95	9.2	-2.9	0.46	9.0	10.3	1565.55	19.7
90th year of simulation														
2140/12/01	1575.24	9.39	84950000	0.1000	0	0	0.95	8.9	-3.1	0.46	8.9	12.0	1565.80	19.5
95th year of simulation														
2145/12/01	1575.24	9.68	84950000	0.1000	48	12	0.95	9.2	-2.9	0.46	9.0	10.4	1565.55	19.7
99th year of simulation														
2149/05/01	1575.24	9.50	84950000	0.1000	126	31	0.95	9.0	-3.1	0.46	8.9	11.7	1565.80	19.6

VENTERSPOST "C" LAG 3												
STUDY AREA			RECHARGE		INFLOWS			OUTFLOWS			RECOVERY	
Date	Study (km2)	Specific	Monthly Rainfall (mm/month)	Effective Recharge (MI/d)	C"BC"	Q "BC" Inflows (MI/d)	L "C" Leakage (MI/d)	C"CD"	Q"CD" Outflows (MI/d)	spring (MI/d)	Water level BH0001636 (mamsl)	ΔH (m)
	54.4	0.0650		7.9%	0.4566	note	0.19	0.1728	+	1539.00	1519.75	
2050/01/01	54400000	0.0650	55	8	0.46	9.2	3.7	0.173	17.1	0.0	1519.78	99.2
2050/02/01	54400000	0.0650	108	15	0.46	9.2	3.6	0.173	17.1	0.0	1519.88	99.1
2050/03/01	54400000	0.0650	159	23	0.46	9.3	3.6	0.173	17.1	0.0	1520.03	99.0
2050/04/01	54400000	0.0650	300	43	0.46	9.4	3.6	0.173	17.1	0.0	1520.36	98.8
2050/05/01	54400000	0.0650	163	23	0.46	9.4	3.5	0.173	17.1	0.0	1520.53	98.7
2050/06/01	54400000	0.0650	85	12	0.46	9.4	3.5	0.173	17.0	0.0	1520.59	98.6
2050/07/01	54400000	0.0650	99	14	0.46	9.5	3.5	0.173	17.0	0.0	1520.68	98.5
2050/08/01	54400000	0.0650	41	6	0.46	9.6	3.5	0.173	17.0	0.0	1520.70	98.4
2050/09/01	54400000	0.0650	0	0	0.46	9.6	3.5	0.173	17.0	0.0	1520.66	98.3
2050/10/01	54400000	0.0650	0	0	0.46	9.7	3.5	0.173	17.0	0.0	1520.63	98.2
2050/11/01	54400000	0.0650	0	0	0.46	9.8	3.5	0.173	17.0	0.0	1520.60	98.2
2050/12/01	54400000	0.0650	0	0	0.46	9.9	3.5	0.173	16.9	0.0	1520.57	98.1
5th year of simulation												
2055/12/01	54400000	0.0650	49	7	0.46	12.5	2.9	0.173	16.2	0.0	1523.66	93.6
10th year of simulation												
2060/12/01	54400000	0.0650	0	0	0.46	13.3	2.1	0.173	15.6	0.0	1528.05	90.0
15th year of simulation												
2065/12/01	54400000	0.0650	47	7	0.46	13.2	1.3	0.173	15.0	0.0	1531.96	86.6
20th year of simulation												
2070/12/01	54400000	0.0650	0	0	0.46	12.6	0.5	0.173	14.3	0.0	1536.18	83.0
25th year of simulation												
2075/12/01	54400000	0.0650	48	7	0.46	11.6	-0.1	0.173	13.7	0.0	1539.38	79.1
30th year of simulation												
2080/12/01	54400000	0.0650	0	0	0.46	10.6	-0.6	0.173	12.9	2.0	1542.24	74.4
35th year of simulation												
2085/12/01	54400000	0.0650	50	7	0.46	10.1	-0.8	0.173	11.9	3.5	1543.28	68.6
40th year of simulation												
2090/12/01	54400000	0.0650	0	0	0.46	9.7	-1.0	0.173	10.8	5.5	1544.34	62.5
45th year of simulation												
2095/12/01	54400000	0.0650	50	7	0.46	9.6	-1.0	0.173	9.7	5.8	1544.50	56.4
50th year of simulation												
2100/12/01	54400000	0.0650	0	0	0.46	9.4	-1.2	0.173	8.7	7.5	1545.24	50.4
55th year of simulation												
2105/12/01	54400000	0.0650	48	7	0.46	9.3	-1.2	0.173	7.7	7.4	1545.23	44.6
60th year of simulation												
2110/12/01	54400000	0.0650	0	0	0.46	9.1	-1.3	0.173	6.8	9.2	1545.92	39.6
65th year of simulation												
2115/12/01	54400000	0.0650	50	7	0.46	9.0	-1.3	0.173	6.4	8.7	1545.77	37.2
70th year of simulation												
2120/12/01	54400000	0.0650	0	0	0.46	8.9	-1.4	0.173	6.3	10.0	1546.20	36.5
75th year of simulation												
2125/12/01	54400000	0.0650	47	7	0.46	9.0	-1.3	0.173	6.3	9.0	1545.86	36.4
80th year of simulation												
2130/12/01	54400000	0.0650	0	0	0.46	8.9	-1.4	0.173	6.3	10.1	1546.24	36.3
85th year of simulation												
2135/12/01	54400000	0.0650	50	7	0.46	9.0	-1.3	0.173	6.3	9.0	1545.88	36.4
90th year of simulation												
2140/12/01	54400000	0.0650	0	0	0.46	8.9	-1.4	0.173	6.3	10.1	1546.23	36.3
95th year of simulation												
2145/12/01	54400000	0.0650	48	7	0.46	9.0	-1.3	0.173	6.3	9.0	1545.88	36.4
99th year of simulation												
2149/05/01	54400000	0.0650	126	18	0.46	8.9	-1.4	0.173	6.3	9.8	1546.23	36.4

BANK"D" LAG3												
STUDY AREA			RECHARGE		INFLOWS			OUTFLOWS			SIMULATION	
Date	Study (km2)	Specific	Monthly Rainfall (mm/month)	Effective Recharge (MI/d)	C"CD"	Q"CD" Inflows (MI/d)	L "D" Leakage (MI/d)	C"DE"	Q"DE" Outflows (MI/d)	spring (MI/d)	Water level BH (mamsl)	ΔH (m)
	156.7	0.0630		8.5%	0.1728	note	0.511	0.0490	-	1502.0	1420.60	
2050/01/01	1.57E+08	0.0630	55	24	0.173	17.13	41.5	0.05	0.2	0.0	1420.73	4.1
2050/02/01	1.57E+08	0.0630	108	48	0.173	17.12	41.4	0.05	0.2	0.0	1420.92	4.0
2050/03/01	1.57E+08	0.0630	159	71	0.173	17.10	41.3	0.05	0.2	0.0	1421.19	4.1
2050/04/01	1.57E+08	0.0630	300	133	0.173	17.08	41.0	0.05	0.2	0.0	1421.64	4.0
2050/05/01	1.57E+08	0.0630	163	72	0.173	17.06	40.9	0.05	0.2	0.0	1421.92	4.1
2050/06/01	1.57E+08	0.0630	85	38	0.173	17.04	40.8	0.05	0.2	0.0	1422.08	4.2
2050/07/01	1.57E+08	0.0630	99	44	0.173	17.02	40.7	0.05	0.2	0.0	1422.27	4.1
2050/08/01	1.57E+08	0.0630	41	18	0.173	17.01	40.7	0.05	0.2	0.0	1422.37	4.1
2050/09/01	1.57E+08	0.0630	0	0	0.173	16.99	40.6	0.05	0.2	0.0	1422.42	4.1
2050/10/01	1.57E+08	0.0630	0	0	0.173	16.98	40.6	0.05	0.2	0.0	1422.47	4.0
2050/11/01	1.57E+08	0.0630	0	0	0.173	16.96	40.6	0.05	0.2	0.0	1422.53	3.9
2050/12/01	1.57E+08	0.0630	0	0	0.173	16.95	40.5	0.05	0.2	0.0	1422.58	3.8
5th year of simulation												
2055/12/01	1.57E+08	0.0630	49	22	0.173	16.18	36.7	0.05	0.0	0.0	1430.11	-1.0
10th year of simulation												
2060/12/01	1.57E+08	0.0630	0	0	0.173	15.55	32.6	0.05	-0.2	0.0	1438.10	-3.6
15th year of simulation												
2065/12/01	1.57E+08	0.0630	47	21	0.173	14.96	28.9	0.05	-0.3	0.0	1445.44	-5.4
20th year of simulation												
2070/12/01	1.57E+08	0.0630	0	0	0.173	14.34	24.9	0.05	-0.3	0.0	1453.27	-5.4
25th year of simulation												
2075/12/01	1.57E+08	0.0630	48	21	0.173	13.66	21.2	0.05	-0.2	0.0	1460.39	-5.0
30th year of simulation												
2080/12/01	1.57E+08	0.0630	0	0	0.173	12.85	17.4	0.05	-0.2	0.0	1467.95	-3.4
35th year of simulation												
2085/12/01	1.57E+08	0.0630	50	22	0.173	11.86	13.9	0.05	-0.1	0.0	1474.75	-1.2
40th year of simulation												
2090/12/01	1.57E+08	0.0630	0	0	0.173	10.81	10.3	0.05	0.2	0.0	1481.90	3.9
45th year of simulation												
2095/12/01	1.57E+08	0.0630	50	22	0.173	9.74	7.0	0.05	0.5	0.0	1488.23	9.2
50th year of simulation												
2100/12/01	1.57E+08	0.0630	0	0	0.173	8.71	3.6	0.05	0.8	0.0	1494.90	15.8
55th year of simulation												
2105/12/01	1.57E+08	0.0630	48	21	0.173	7.71	0.7	0.05	1.0	0.0	1500.72	21.3
60th year of simulation												
2110/12/01	1.57E+08	0.0630	0	0	0.173	6.84	-2.3	0.05	1.3	10.0	1506.42	27.1
65th year of simulation												
2115/12/01	1.57E+08	0.0630	50	22	0.173	6.43	-3.4	0.05	1.4	22.1	1508.58	29.1
70th year of simulation												
2120/12/01	1.57E+08	0.0630	0	0	0.173	6.31	-3.9	0.05	1.5	30.3	1509.67	30.3
75th year of simulation												
2125/12/01	1.57E+08	0.0630	47	21	0.173	6.30	-3.8	0.05	1.5	28.2	1509.43	29.9
80th year of simulation												
2130/12/01	1.57E+08	0.0630	0	0	0.173	6.28	-4.0	0.05	1.5	32.1	1509.88	30.5
85th year of simulation												
2135/12/01	1.57E+08	0.0630	50	22	0.173	6.29	-3.8	0.05	1.5	28.6	1509.48	29.9
90th year of simulation												
2140/12/01	1.57E+08	0.0630	0	0	0.173	6.28	-4.0	0.05	1.5	32.0	1509.88	30.5
95th year of simulation												
2145/12/01	1.57E+08	0.0630	48	21	0.173	6.29	-3.8	0.05	1.5	28.7	1509.49	30.0
99th year of simulation												
2149/05/01	1.57E+08	0.0630	126	56	0.173	6.28	-4.0	0.05	1.5	31.3	1509.87	29.8