

Hydrogeological Characterisation of Fountains East and Fountains West Karst Compartments

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ABSTRACT

The Fountains East and Fountains West groundwater compartments (by means of the Upper and Lower Fountain springs) have been supplying the City of Tshwane with water since the founding of Pretoria in 1855. These adjacent compartments, which are underlain by the Malmani dolomites of the Chuniespoort Group, are separated by the Pretoria syenite dyke and are bounded to the north by the rocks of the Pretoria Group (Timeball Hill Formation). The perennial, artesian springs that drain the compartments are situated within the Groenkloof Nature Reserve and currently supply the citizens of Pretoria with 46 ML/day of water. Inorganic chemistry data (2007-2012) as well as spring discharge volumes (2011-2012) for the Upper and Lower Fountain springs along with water levels obtained from DWA boreholes in the Fountains East and Fountains West compartments (1984-2013) and isotope data for both springs and numerous rainfall stations in the City of Tshwane (1979-2007) were used in order to aid in the characterization of the springs and the compartments to which they belong. This was done by means of statistical analysis, Piper diagrams, bar graphs and temporal plots. Interpretation of the water levels indicates that the Fountains East compartment generally has more shallow water levels while the Fountains West compartment has average water levels that are approximately 8.5m deeper and irrespective of the compartment, groundwater flow is generally from the south to the north in the karst aquifer. From the chemistry data the hydrochemical characteristics of both springs are found to be similar with the groundwater signature for both springs being $\text{Ca}(\text{Mg})\text{-HCO}_3$ which is indicative of fresh, recently recharged groundwater. Isotopically both springs are found to be depleted (as a result of the rainout effect) and may indicate that recharge of the compartments did not occur in the Pretoria area.

Contents

1. Introduction	1
1.1. Background.....	1
1.2. Project objectives.....	2
2. Literature review	4
2.1. Karst Aquifers	4
2.1.1. Formation of karst aquifers	4
2.1.2. Karst in South Africa	6
2.1.3. The importance of chert in South African karst systems.....	7
2.1.4. Existing conceptual karst models	8
2.1.5. Compartmentalisation of South African karst systems	12
2.1.6. Typical Parameters	15
2.1.7. Typical chemistry of the groundwater.....	16
2.2. Karst Springs	16
2.3. Stable Environmental Isotopes.....	18
2.3.1. Stable isotopes in precipitation	19
2.3.2. Stable isotopes in groundwater	20
3. Materials and Methods.....	22
3.1. Climate, Rainfall and Vegetation	22
3.2. Topography and Drainage	23
3.3. Geology	25
3.3.1. Lithostratigraphy	25
3.3.2. Structural Geology	28
3.4. Methodology	29
4. Data.....	31
4.1. Groundwater Levels.....	31
4.2. Spring Discharge	31
4.3. Inorganic Chemistry	31
4.4. Stable Isotopes	31
4.5. Borehole logs.....	32
5. Results.....	33
5.1. Groundwater Levels.....	33
5.2. Spring Discharge	39
5.3. Inorganic Chemistry	40
5.4. Stable Isotopes	49
5.5. Borehole logs and Conceptual Model	55

6. Discussion	58
6.1. Groundwater Levels.....	58
6.2. Spring Discharge	59
6.3. Inorganic Chemistry	60
6.4. Stable Isotopes	63
6.5. Local Geology and Conceptual Model.....	65
7. Conclusions	66
8. Way Forward	68
9. References	69
APPENDIX A: DWA water levels.....	73
APPENDIX B: Spring discharge volumes.....	74
APPENDIX C: Inorganic chemistry.....	75
APPENDIX D: Stable isotopes (springs)	76
APPENDIX E: Stable isotopes (rainfall)	77
APPENDIX F: NGA boreholes	78

Table of Figures

Figure 1. Locality map (Google Earth imagery superimposed onto a Digital Elevation model).....	3
Figure 2. Conceptual karst landscape (SANS 2009a).	9
Figure 3. The comprehensive karst system and associated processes (Ford and Williams 2007 in Karimi 2012).	10
Figure 4. General karst conceptual model (Holland 2007).	11
Figure 5. Schematic cross section through a typical dolomite aquifer (DWA, 2006).	12
Figure 6. Dolomite compartments in the City of Tshwane area (Holland for DWA, 2008). ..	14
Figure 7. Diagrams of the three different karst spring types (Ford and Williams, 2007).	18
Figure 8. Bar graph showing rainfall and evaporation for the Pretoria area.	22
Figure 9. Topography map showing the location of the Upper and Lower Fountain springs.	24
Figure 10. Geological map.	27
Figure 11. Time series of water levels for all DWA boreholes.....	36
Figure 12. Time series of water levels for all DWA boreholes except A2N0682.....	37
Figure 13. Correlation graph showing the relationship between the average water level and elevation for the DWA boreholes.....	37
Figure 14. Contour map: average water levels for the DWA boreholes and the Upper and Lower Fountain springs.....	38
Figure 15. Spring discharge volumes for the Upper and Lower Fountain springs as well as rainfall and evaporation data for the Pretoria area.....	40
Figure 16. Bar graph of major ion chemistry for the Upper and Lower Fountain Springs	42
Figure 17. Average concentrations of cations and anions of the Upper and Lower Fountain springs (percentage ion balance indicated above each).....	42
Figure 18. Piper diagram for the Upper and Lower Fountain springs.....	43
Figure 19. Time series: major cations (2008 – 2012).....	44
Figure 20. Time series: major anions (2008 – 2012).	45
Figure 21. Schoeller diagram for the Lower Fountain spring.	46
Figure 22. Schoeller diagram for the Upper Fountain spring.	46
Figure 23. Pie diagrams (meq/l).	48
Figure 24. ¹⁸ O time series graph.	50
Figure 25. ² H time series graph.....	50
Figure 26. Evaporation data for the rainfall station Rietvlei at Rietvlei Dam (1997-2007)....	51
Figure 27. ¹⁸ O time series graphs for rainfall stations in Pretoria.	52
Figure 28. ² H time series graphs for the rainfall stations in Pretoria.....	52
Figure 29. Relationship between ¹⁸ O and ² H in the rainfall in the Pretoria area and the Upper and Lower Fountain springs.	54
Figure 30. Conceptual model of the Upper and Lower Fountain springs.	57

Table of Tables

Table 1. The formations of the Malmani Subgroup and their chert contents (modified from DWA, 2006).	8
Table 2. Typical physical parameters of karst aquifers.	16
Table 3. Water level and elevation data for DWA boreholes in the Fountains East and Fountains West karst compartments.	35
Table 4. Statistical analysis of groundwater level data for boreholes in the study area.	36
Table 5. Spring discharge volumes for the Upper and Lower Fountain Springs.....	39
Table 6. Major ion chemistry data.	41
Table 7: Stable isotope concentrations for the Upper and Lower Fountain springs.	49
Table 8. Remaining ¹⁸ O and ² H data.	53
Table 9. Statistical analysis of NGA borehole logs.	55

1. Introduction

1.1. Background

The karst aquifer in the City of Tshwane municipal area has supplied Pretoria with water from the time of its founding in 1855 by means of two artesian springs. From the town's registration in 1855 until 1935 the sole source of water for the entire town was the artesian springs (known as the Upper and Lower Fountain springs today) and thereafter until present a mixture of the springs and borehole groundwater as well as surface water has been used (Haaroff et al., 2012). These perennial springs are situated within the Groenkloof Nature Reserve. The development of the approximately 8 km² nature reserve was initiated with the purpose of protecting the water that emerges from the springs. Today the springs together supply nearly 46 ML of water a day to the 2.4 million (2012) inhabitants of Pretoria, with people in the central business district (CBD) of the city receiving pure spring water that has only been treated by chlorination.

A locality map showing the two springs in relation to each other, the Central Business District of Pretoria as well as boundaries of the City of Tshwane is illustrated in Figure 1. The springs are situated approximately 500 m apart and less than 5 km away from the CBD of Pretoria.

The purpose of this project is to characterise the water emanating from the Upper Fountain (UF) and Lower Fountain (LF) springs which will aid in understanding the karst aquifer and the controls on groundwater system within and possibly between these compartments. Characterisation of these springs and the karst compartments to which they belong is an important and necessary undertaking for a number of reasons:

- A large number of people are dependent on the water emerging from both springs. It is therefore necessary for the water to be of sufficient and reliable (for future generations) quantity and quality.
- Characterisation of the springs will form a platform for the understanding and sustainable use of this valuable water resource and other dolomite springs in the country.
- The inherent vulnerability of karst systems (see Leyland, 2008) to pollution is an additional reason that the characterisation of the groundwater emerging from the karst springs is important. Tying in with the vulnerability of karst systems is the close proximity of the Upper and Lower Fountain springs and their respective karst compartments to the large metropolitan area of Pretoria (City of Tshwane). The only protection to the springs is provided by the Groenkloof Nature Reserve which only covers a small portion of the Fountains East and Fountains West karst compartments.
- By characterising the groundwater in the karst compartments it is also anticipated that a connection (or lack thereof) between the Fountains East and Fountains West karst aquifer compartments may be assessed. A connection between the compartments would imply that movement of water occurs through the Pretoria dyke which runs approximately north south between the two springs.

1.2. Project objectives

The project aims to:

- Characterise the discharge and chemistry of the Upper and Lower Fountain springs and therefore the karst compartments that supply them.
- Characterise the Fountains East and Fountains West karst compartments in terms of water levels and the differences or lack thereof between the two compartments
- Use isotope data to further characterise the Upper and Lower Fountain spring as well as to identify whether the recharge area of the springs lies within Pretoria.

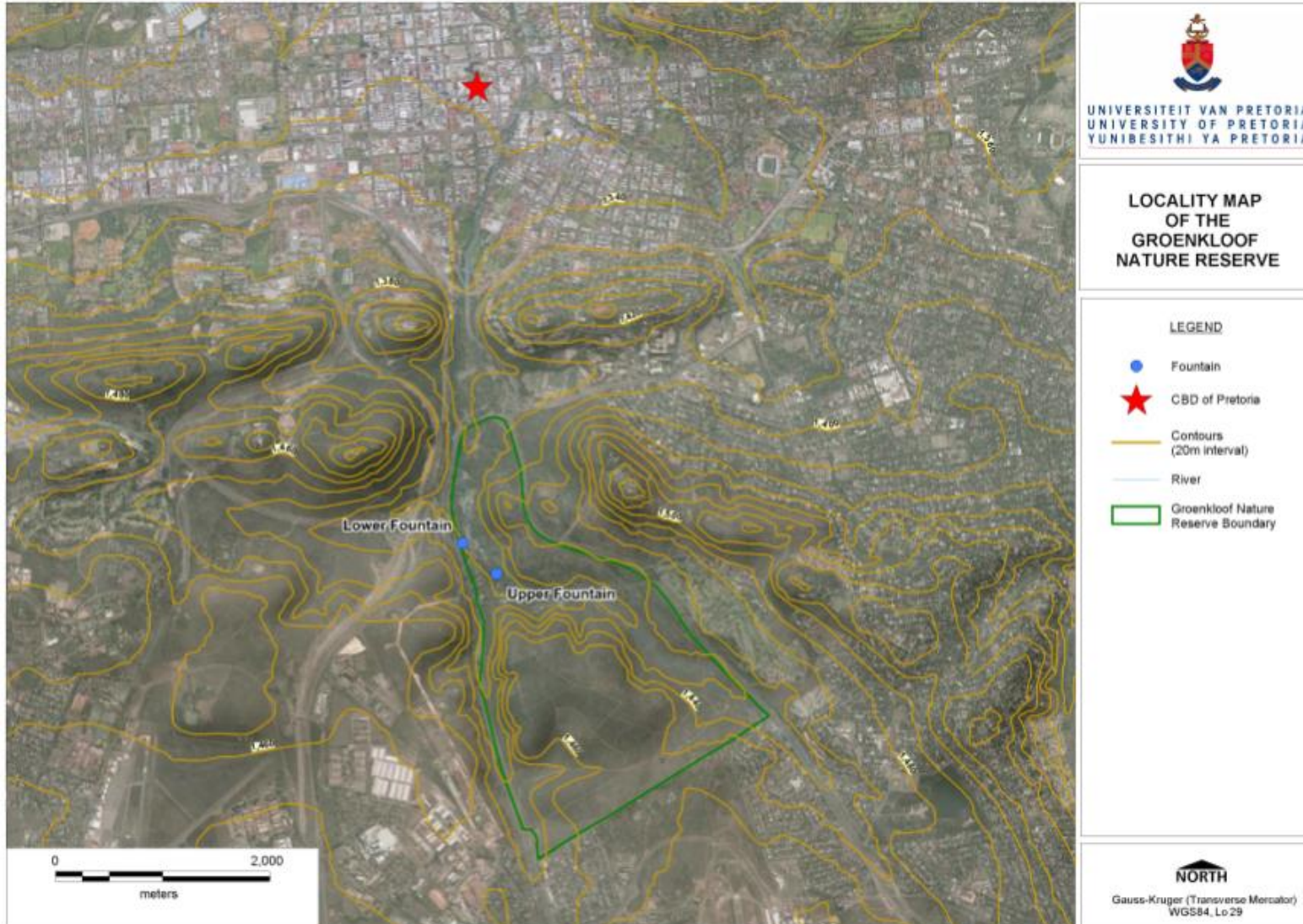


Figure 1. Locality map (Google Earth imagery superimposed onto a Digital Elevation model).

2. Literature review

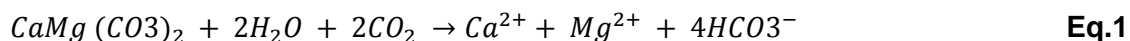
2.1. Karst Aquifers

2.1.1. Formation of karst aquifers

The karst in Gauteng is formed by the dissolution of the dolomite rocks of the Chuniespoort Group. The process of dolomite dissolution is called karstification. Karstification is a vital process in the dolomite because the karst aquifers form an essential source of water in South Africa and because dolomite bedrock is almost impermeable and has low porosity in the unweathered, unjointed state (Buttrick et al., 1993). Once karstification has taken place in the dolomite it develops highly permeable zones that have a very high storage capacity and is able to deliver high yielding boreholes (Leyland et al., 2008).

This process is controlled by many factors including the quantity of precipitation, the partial pressure of CO₂ and the hydraulic gradient between the top of the dolomite recharge area and the level of the springs or discharge points. The climate is also an important factor as it controls the CO₂ concentration as well as the temperature (Holland, 2007). Another important factor in the development of karst is the presence of structural features such as faults, dykes, fractures and bedding planes. These features are necessary as dolomite is basically impermeable in the unweathered state. They provide the pathways for water to move vertically through the dolomite to the water table and then for it to move horizontally towards a natural outlet. Large horizontal caverns are also more likely to form at or near highly weathered vertical fractures (DWAF, 2006).

As precipitation infiltrates into the soil CO₂ dissolves into the infiltrated water. This is because the partial pressure in the soil is higher than of the atmosphere as a result of root and microbial respiration (Carl et al., 1984). The resulting carbonic acid and groundwater mixture can have a pH of 5.7 and lower. As this mixture circulates through the dolomite rock it begins to dissolve the carbonate minerals (Holland, 2007) with this dissolution process of dolomite, called karstification, with a solubility product of $K = 10^{-16.9}$ (Appello and Posma, 2005). The solubility product describes a minerals affinity for dissolution. The solubility product of dolomite is low when compared with most salts as well as with calcite ($K = 10^{-8.48}$) but when it is compared with that of silicate minerals it is generally high. This indicates that dolomite may dissolve slower than salts and calcite but dissolves faster than silicate minerals (Eq. 1).



Once the carbonate minerals have been dissolved the flowing water removes the products and a hierarchical system of underground voids known as either the karst network or conduit system in the vadose and phreatic zone results (Leyland et al., 2008). The remnants of the karstification process that are generally washed through the fractures are made up of insoluble chert, iron oxides and manganese oxides. This mixture of insoluble material is called wad (weathered altered dolomite consisting of silica, iron and manganese

oxides and hydroxides) and is a dark, friable and highly porous material (DWAF, 2008 and Meyer, 2014).

According to Martini and Kavalieris (1976) there are three distinct stages in the process of dolomite dissolution. The first stage is called the incipient stage and it is characterised by a light brown discolouration on the dolomite and a mechanical weakening. The discolouration of the dolomite is associated with an increased concentration of Fe and Mn and a decrease in concentration of Ca and Mg along the boundaries of crystals.

The second stage is called the arenaceous stage of dissolution. In this stage the dolomite is reduced to a brown or speckled white and brown arenaceous/ granular material that is easily crushed or naturally eroded. On a microscopic level the crystals have heavy oxide coatings and high concentrations of quartz and dolomite are present. No calcite or aragonite is found in deeper caves which means that dissolution is congruent. An increase in porosity is also observed in this stage (Martini and Kavalieris, 1976).

In the third and final stage of dissolution called the cellular stage Martini and Kavalieris (1976) state that all carbonate is removed. This stage is only preserved in protected environments such as undisturbed caves. The whole fabric of the rock is composed of Fe and Mn oxides and hydroxides and silica. These minerals pseudomorph the original dolomite structure to form a dissolution product called wad in South Africa. The wad is fragile, highly compressible and has a low density of 0.4 g/cc. Up to 20% of its mass is made up by water and void spaces makes up 90% of the material.

The three stages of dissolution above lead to the formation of the karst network. There are many theories of how the pattern of this network develops. Scientists such as Drogue (1974), Grillo (1977) and Razack (1980) all state that the pattern of the karst network follows the pre-existing fracture pattern in the dolomite rock. According to them the larger conduits form along large faults and fractures while minor conduits form along fissures. On the other hand however karst hydrogeologists believe that the pattern of the karst network is different from the pattern of the pre-existing fractures. The "karst processes" are thought to select any geological discontinuity such as a bedding plane, joint or fracture and enlarge it until it is hydraulically continuous from surface to spring (Leyland et al., 2008).

At the end of the karstification process the conduit network that is left is highly permeable, highly heterogeneous and can conduct large quantities of water as the conduits in these networks are generally a number of meters wide and can be kilometres long. This is because karst aquifers have a double or triple porosity which is composed of solution voids, fractures and the intergranular pores in the rock matrix (this portion is very low in unweathered dolomite but may be higher in the weathered dolomite). The pores in the rock matrix generally form the storage in the aquifer while the conduits and fractures act as the drains in the aquifer (Leyland et al., 2008).

2.1.2. Karst in South Africa

There have been at least four periods of karstification in 2.2 billion year history of dolomite in South Africa. All of the periods coincide with periods when weathering was pronounced and deposition has suspended (Martini and Kavalieris, 1976).

The first of these periods is the pre-Pretoria Group. This period occurred before 1.95 Ga and is evidenced by the 'Giant Chert' which is a chert breccia that has a dark siliceous matrix. Other evidence to support this karstification period can be seen in the karst features such as filled paleosinkholes and cave passages and collapsed chambers.

According to Martini and Kavalieris (1976) the next period of karstification occurred pre-Waterberg which is approximately 1.7 Ga. Evidence for this period can be seen where dissolution cavities in the dolomite are filled with red sandstone (these have been observed near the border of Botswana).

The third period is the pre-Karoo period and occurred before the deposition of the Dwyka Group during the Upper Carboniferous-Lower Permian. During this period the dolomite was exposed to erosion for a long time (in the order of 1000 million years). The evidence for this period is seen very prominently in the Pretoria area with chert breccia filled sinkholes, kaolinitic clay as well as carbonaceous sediments. Other evidence includes some bleached iron poor varve-like deposits which are expected in the colder climate of the pre-Karoo times and thick pockets of coal lying on the Malmani dolomites (Martini and Kavalieris, 1976).

The fourth and final period is that of the Tertiary to present. During this period many orogenies were occurring over most of South Africa. This means that the rocks of the Malmani Subgroup as well as some other deeply buried rocks were exposed on surface to erosion. These orogenies are associated with different erosion surfaces and will be discussed further below.

A single erosion cycle from the rifting and breakup of Gondwanaland to the early Miocene was the result of the first uplift during this period. By the early Miocene a gently dipping pedepane called the African surface was formed at elevations between 500 and 600 m. The majority of erosion took place at the beginning of this period. Between the Miocene and the Pliocene there was another cycle of uplift of approximately 150-300 m that caused the continent to tilt slightly. The pedepane formed during this short lived and modest uplift cycle is called the Post Africa 1. At the end of the Pliocene a second major uplift occurred that lifted the eastern interior up to 900 m (and less in the west) and caused the down cutting of many rivers (Trollip, 2006).

After completion of the dissolution and the karstification process, four karst morphological types remained in South Africa, viz.: plateau type, escarpment type, Bushveld type and Vaal river type.

The escarpment type of karst will be elaborated on as it is found between Pretoria and Krugersdorp. This karst type was formed when an erosion cycle penetrated inland from

the east coast to the plateau causing a jagged topography and a dendritic network of valleys. In this area karst morphology is poorly developed on surface except for a few lapies and collapse sinkholes. Underground morphology is however well developed with numerous caves being present. Some of these caves are associated with the previous plateau type but many are developing today in the escarpment type topography (Martini and Kavalieris, 1976).

A full study of the karst regions in South Africa can be found in hydrogeology of the karst belt compiled by Meyer (2014). This study includes the description of the physical environment, geology, karst development and the hydrogeology of the different karst compartments in South Africa. Previous investigations that were conducted in the area are also incorporated in the study including reports compiled by DWA. A section specifically on the hydrogeology of the karst in the Pretoria area can also be found.

2.1.3. The importance of chert in South African karst systems

A feature of the South African karst with regard to the groundwater bearing capacity is a result of the chert content of the different formations of the Malmani dolomite (DWAF, 2006). The Oaktree, Lyttelton and middle Monte Christo Formations are chert poor and therefore do not have large amounts of water as these formations have not undergone as much dissolution. The Eccles, lower Monte Christo and the Upper Monte Christo Formations are, however, chert rich (Kafri and Foster, 1989). The formations which are rich in chert can have a SiO₂ content of up to 60 percent (Foster, 1989).

The difference in solubility between the chert and dolomite is the reason that the chert rich formations being able to bear more groundwater than the chert poor formations. Chert is much more resistant to weathering than dolomite. This means that once the dolomite has dissolved in the chert rich layers (such as the Eccles and Monte Christo Formations) the chert is still in place and forms a skeletal structure which holds up the remaining dolomite and allows for the formation of caverns and cavities in which the water can flow (DWAF, 2006).

Table 1. The formations of the Malmani Subgroup and their chert contents (modified from DWA, 2006).

Formation	Thickness (m)	Description
Eccles	~380	Chert rich, dark coloured dolomite with stromatolites and oolitic bands. Chert contents increases to the top
Lyttelton	~150	Chert-free, dark coloured dolomite with large stromatolitic mounds
Upper Monte Christo	~258	Chert-rich dolomite
Middle Monte Christo	~162	Chert-rich dolomite
Lower Monte Christo	~275	Chert-rich dolomite
Oaktree	~200	Chert-poor dolomite with interlayered carbon-rich shale towards the base

2.1.4. Existing conceptual karst models

As one of the products of this undertaking is a conceptual model of the Upper and Lower Fountain springs and the two compartments to which they belong, it is therefore necessary to elaborate on the existing models accepted for the South African karst aquifers. There are a number of conceptual models of the karst aquifers in South Africa which have been created by various authors over the years. These models explain the karstification of dolomite and the movement groundwater in the karst features for different localities around South Africa. Many of these models are based in Gauteng as karst features are well developed in this area. Four of these models have been chosen to illustrate the conceptual understanding of the karst aquifer in the City of Tshwane area.

The first conceptual model is taken from SANS 1936-1 2009 (Figure 2). This model shows that the process of dissolution of the dolomite is generally vertically controlled, meaning that the unweathered dolomite bedrock is overlain by slightly weathered jointed dolomite bedrock and this is overlain by totally weathered low strength residual material. The residual material is made up of mainly wad (manganese oxide), chert and iron oxides (Figure 2).

This model also illustrates the formation of solution cavities in the weathered dolomite bedrock as well the subsidence of the low strength residuum into a sinkhole. Indicated as well is the formation of grykes which are deeply incised openings where dissolution has occurred along joints from the surface down and pinnacles which are isolated pieces of dolomite bedrock within the weathered residuum (Trollip, 2006).

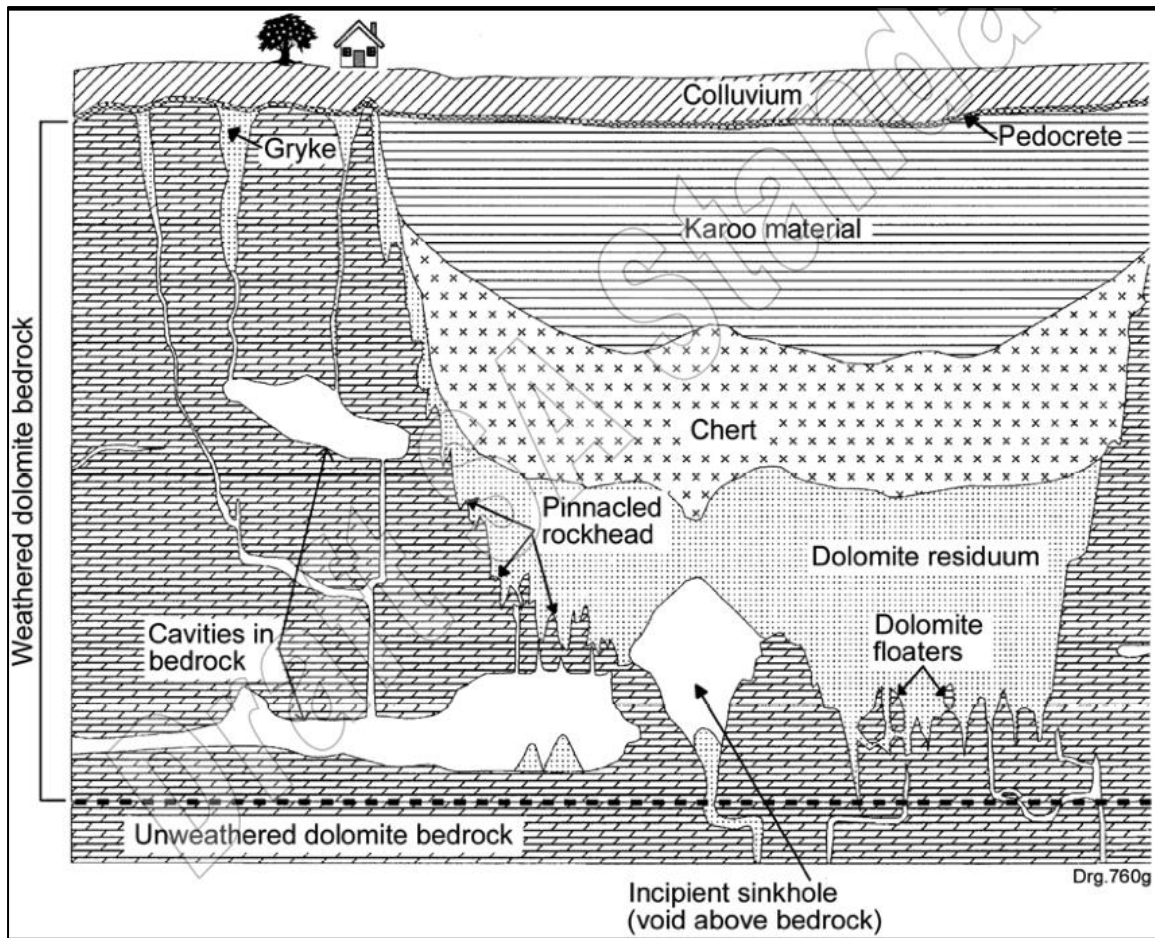


Figure 2. Conceptual karst landscape (SANS 2009a).

Figure 3 which is taken from Karimi (2012), illustrates the typical processes that occur in a karst system which has been affected by geothermal/metamorphic processes. The system moves from a highland (made up of igneous rock) where the dolomite is being actively weathered and eroded to form a karst landscape to a low lying lagoon area where deposition is forming new dolomite rock. The black arrows show the direction of allogenic and groundwater movement in the system and the clear arrows indicate the movement of the mixing of geothermal water from deeper sources and from the igneous country rock. This model also illustrates some features developed in a karst landscape such as karen, dolines and polje (Figure 3).

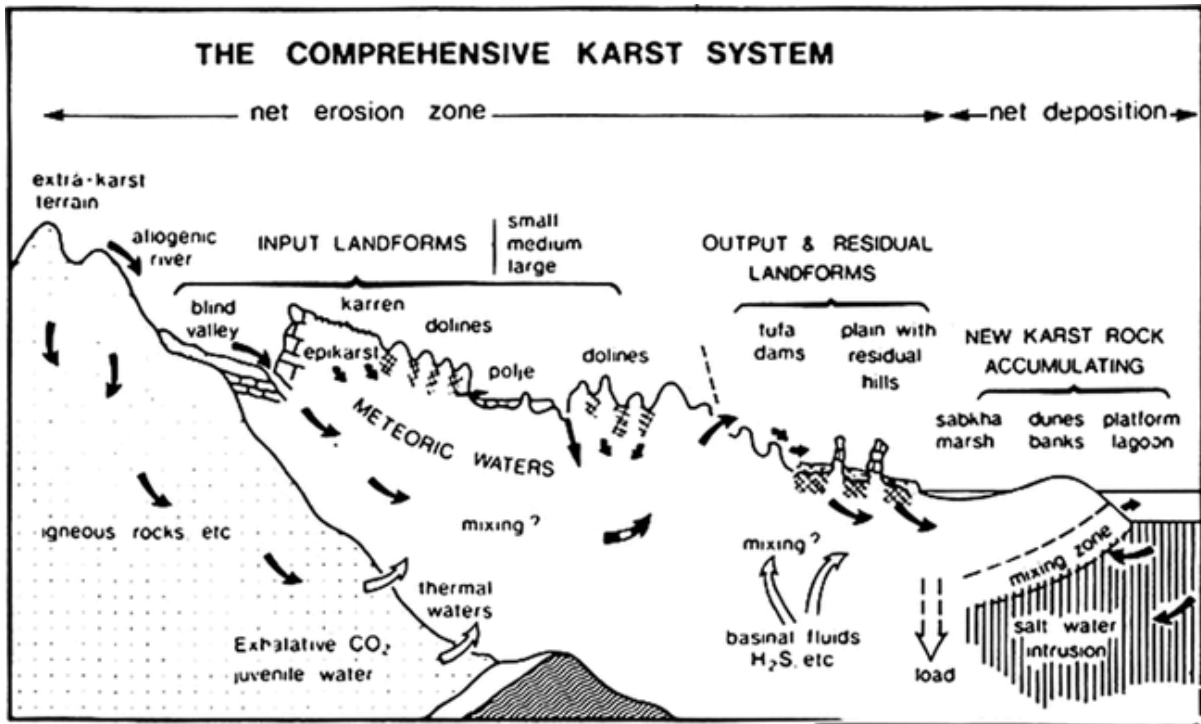


Figure 3. The comprehensive karst system and associated processes (Ford and Williams 2007 in Karimi 2012).

Holland (2007) used a model to illustrate the basic hydraulic behaviour (recharge, flow and discharge), structure and essential features (such as depressions and solution cavities) in a karst landscape. This model shows the different sources of recharge that a karst landscape is subject to such as allogenic recharge (surface water streams that originate in non-carbonate parts of the basin and drain into the carbonate aquifer via swallow holes, sinkholes or dolines), autogenic recharge (precipitation that enters the carbonate aquifer directly from the overlying soil) and recharge from perched catchments which occurs when precipitation from a perched aquifer overflows into the vadose zone of the karst aquifer vertical shafts or solutionally widened fractures (White, 2002). This model also illustrates the triple porosity of a karst system which is made up of the solutional voids, the rock matrix (pore spaces) and the fractures. This model also shows that recharge may be diffuse (when entering the system through the epikarst zone) or it may be concentrated (when infiltration through closed depressions). This is shown in Figure 4.

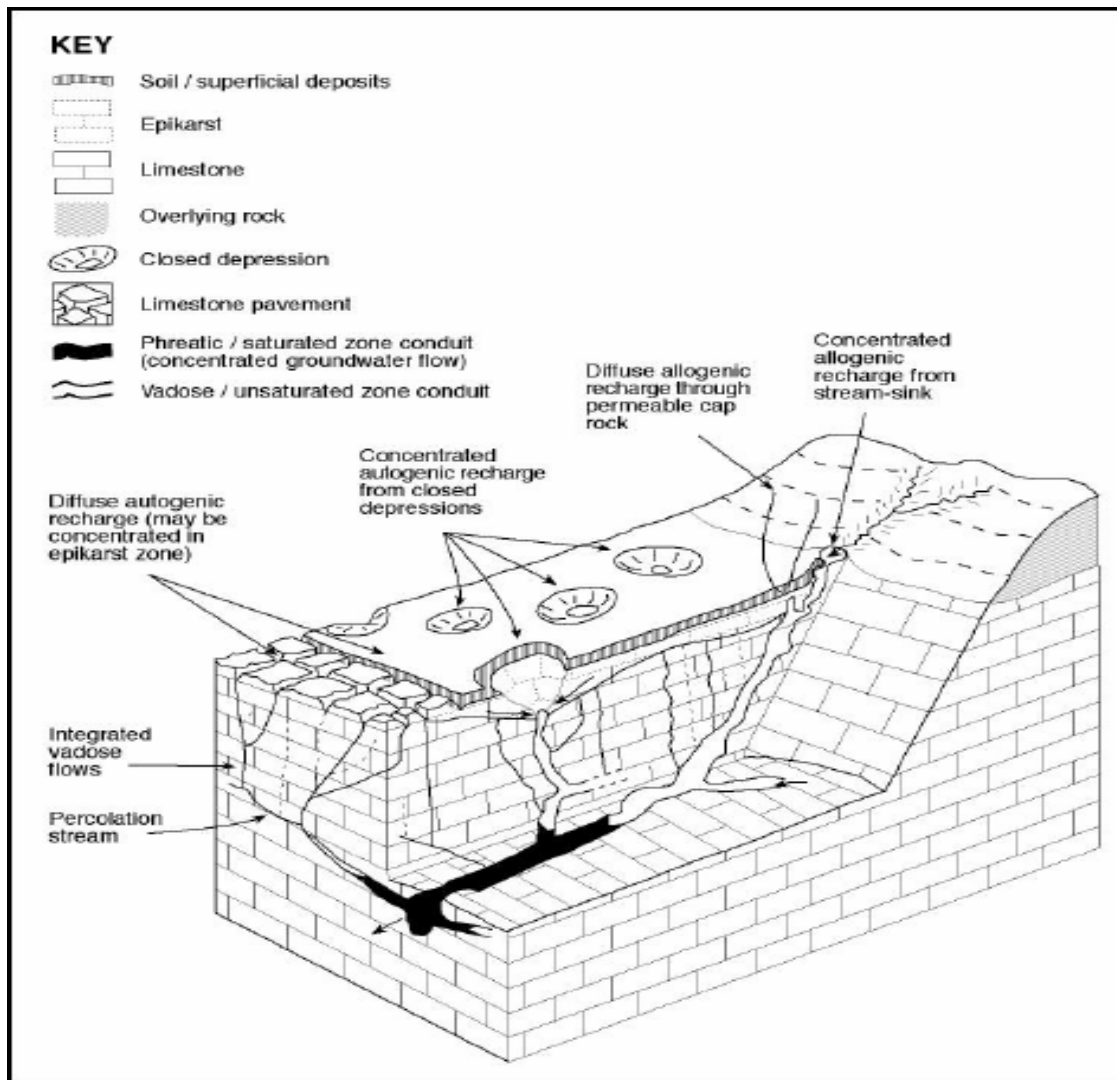


Figure 4. General karst conceptual model (Holland 2007).

The final karst conceptual model (Figure 5) is taken from DWAF (2006) and illustrates an unspoiled karst system from which very little water is abstracted. The arrows indicate the movement of water from the recharge area through the cavities in the dolomite rock. Abstraction is occurring on the left hand side of the model from a pumping borehole. This causes a localised cone of depression around the pump. Also illustrated are dykes (that act as impermeable boundaries) which compartmentalise the South African karst systems. A typical South African scenario of a spring caused by groundwater damming against an impermeable dyke can also be seen.

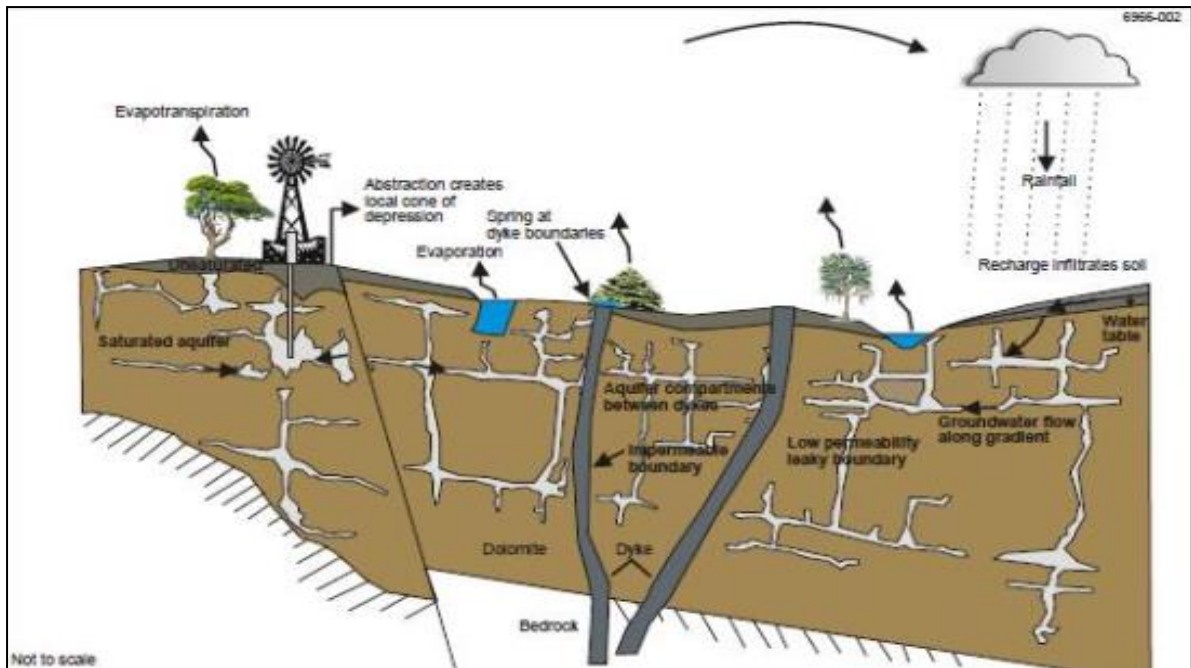


Figure 5. Schematic cross section through a typical dolomite aquifer (DWAf, 2006).

2.1.5. Compartmentalisation of South African karst systems

When dolerite or syenite dykes intrude into the dolomite they form barriers to the flow of groundwater and depending on the permeability (or rather the impermeability) of the dykes they may have a damming effect on the groundwater (DWAf, 2006). According to Trollip (2006) the karstified Malmani dolomites in the area of the City of Tshwane were probably compartmentalised in a similar manner by intruding dykes and/or sills.

According to DWAf (2006) if these barriers are effective the compartments can fill with groundwater as time goes on and the movement of this groundwater can only happen via near surface flow or spring flow. The near surface flow will only occur in the upper few meters of weathered dolerite while the spring flow occurs at the lowest elevation. It is for this reason that water levels between compartments may vary by 10s of meters.

Groundwater movement may occur between different compartments as a result of fractured or weathered dykes or the near surface weathered zone. This does not however mean that the dolomite rocks of the Malmani Subgroup can be treated as one large water resource, seeing that compartmentalisation from more substantial dykes may inhibit the flow of groundwater from one compartment to another (DWAf, 2008). These compartments therefore form separate aquifer units to a greater or lesser extent. The karst aquifer compartments, also, do not necessarily coincide with quaternary catchment borders.

There is some debate on the subdivisions of the of the dolomite compartments in the Tshwane area as the linear structures such as dykes which form the boundaries of the compartments are not always continuous and even when they are groundwater levels may not differ considerably. It also may not be possible in some cases to assess the extent to

which these linear structures prevent or allow the movement of water from one compartment to the next (DWAF, 2008). For this study the map compiled by Holland for DWA in 2007 (Figure 6) was used as it is based on the most recent water level observations and other relevant data.

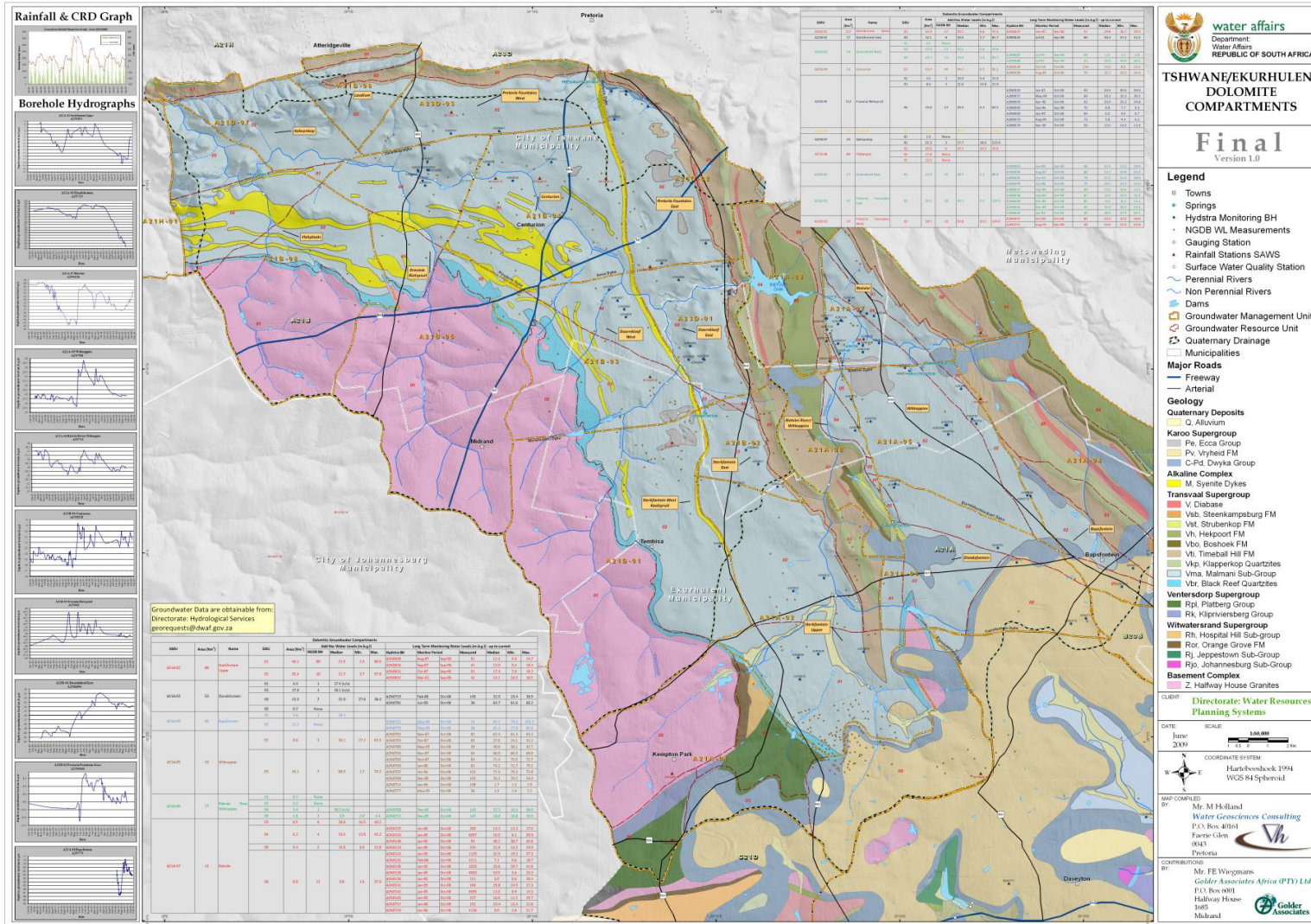


Figure 6. Dolomite compartments in the City of Tshwane area (Holland for DWA, 2008).

There are many different compartments formed by the dykes that have divided the Malmani dolomites within the City of Tshwane area (Haupt, 1985). Examples of these compartments (taken from Holland, 2007) are Aalwynkop, Bapsfontein, Centurion, Doornkloof East, Doornkloof West, Elandsfontein, Erasmia Rietspruit, Laudium, Pretoria Fountains East, Pretoria Fountains West, Rietvlei, Rietvlei River Witkoppies, Sterkfontein East, Sterkfontein Upper, Sterkfontein West Kaalspruit, Vlakpaats and Witkoppies.

The Pretoria Fountains West karst aquifer compartment in which the Lower Fountain spring is situated is bound to the east by an unnamed dyke, to the north by the clastic sedimentary rocks of the Pretoria Group, to the South by the Irene Dyke and to the East by the Pretoria dyke. To the north and east the rocks of the Pretoria Group form the boundary of the Fountain East compartment while to the South the boundary is formed by the Irene dyke and to the West the aquifer compartment boundary is formed by the Pretoria dyke.

To the west the Fountains West compartment is bordered by the Laudium compartment and to the south by the Erasmia Rietspruit and the Centurion compartments. The Fountains East compartment is bordered to the west by the Fountains West and Centurion compartments and to the south by the Doornkloof East compartment.

2.1.6. Typical Parameters

A number of attempts aimed to quantify the hydraulic parameters of the Malmani dolomite karst aquifer in the Gauteng area. Some of the parameters stated by a few authors have been summarised in Table 2. These values are for the karstified dolomite not the unweathered dolomite which is almost impermeable with an effective porosity between 0 and 0.3 percent (Frommurze, 1937).

The value for effective porosity taken from Mulder (1988) is for the leached zone in the dolomite while the storage value is the total storage per year for all the dolomite (2750 km²) in the Gauteng area. DWAF (2006) has given the storage as a total storage in all the South African dolomite.

With respect to the storage of the City of Tshwane karst aquifer there are three main models describing where the groundwater is stored in the aquifer (Leyland, 2008). The first model states that groundwater is stored in the matrix of the dolomite rock and it assumes that the saturated zone is hydraulically continuous. It also assumes that the saturated zone has a low permeability and the aquifer is thought to be drained by conduits.

The second model states that groundwater is stored in karstic voids. These voids are thought to develop around the conduits and are connected to these conduits by high water head losses. This model assumes that there are no exchanges between the matrix and the conduits. The third model states that there is no groundwater storage in the phreatic zone. This model assumes that the conduits drain groundwater from the unsaturated zone and that the epikarst zone plays an important role in storage in the system. Generally depending on the rock texture and geological history one of these models is prevalent.

Table 2. Typical physical parameters of karst aquifers.

Author	Effective porosity	Transmissivity	Storage	Storativity	Storage coefficient
Buttrick and Van Rooy (1993)	2-14%	10-30000m ² /day	0.0005-9.5%		
Foster (1989)	1-3.4%	10-30000m ² /day			
Mulder	20%	10-30000m ² /day	12000 million m ³	0.0001-0.1	
DWAF (2006)			5000 million m ³		
DWAF and WRC(1995)					0.1

Karst aquifers generally have a high storage and, in the absence of groundwater abstraction, natural water level fluctuations in these aquifers are thought to be small and generally less than 5m (DWA, 2008). The high storage in karst aquifers is a result of dissolution voids. These voids are also the reason for water levels in dolomite being typically flat as opposed to other aquifers where water levels follow the topography regionally (Trollip, 2006).

2.1.7. Typical chemistry of the groundwater

According to the Groundwater Resources of the Republic of Southern Africa sheet 2 (1995) the typical groundwater type for the karst aquifer in the City of Tshwane area is Ca-HCO₃ with the dominant cations being calcium and magnesium and the dominant anion being bicarbonate. According to DWAF (2006) the average total dissolved solids (TDS) of the dolomite groundwater in the area is between 200 and 300mg/l (with 87 percent of these salts being calcium and magnesium bicarbonates, 1.6 percent being calcium sulphate and 4.9 percent being chloride). This chemical signature of groundwater is a result of the dissolution of the rock which it has passed through and is therefore representative of composition of the rock. The Ca(Mg)-HCO₃ signature of the dolomite groundwater is therefore representative of the dolomite rock which it has flowed through.

2.2. Karst Springs

A great number of the world's largest springs originate in karst environments. South Africa is no exception with karst springs being a common occurrence in the Malmani dolomites (DWAF, 2006). These karst springs which are the surface expression of the water table form a discharge point for the dolomite groundwater.

Surface fluvial processes become dominant at the point that a spring emerges and the spring controls the depth of the water table at this output point. It is for these reasons as

well as that the spring determines the head in the aquifer that springs exert a significant amount of control on karst systems. Karst springs are themselves however controlled by a number of factors including geomorphological events, sea level fluctuations and valley aggradation (Ford and Williams, 2007).

Another factor that influences the control a spring will have on the karst aquifer is the type of spring. If the function of hydrogeological control is considered there are three distinctive types of karst springs. The first type is free draining springs, consisting of karst rocks sloping into a valley with the karst groundwater draining freely into the valley. The second type of karst springs are dammed springs. This is the most common type of karst spring and is the product of a barrier (such as a dyke or change in lithology) lying in the flow path of the karst groundwater. Dammed karst springs may be impounded, aggraded or coastal. Confined springs are the third type of karst springs. These springs form an artesian system and are a result of low permeability rocks overlying the karst (Ford and Williams, 2007).

Karst springs may also be classified according to the temporal nature of the outflow (perennial, non-perennial or episodic) or the origin of the groundwater (emergent, resurgent or exsurgent).

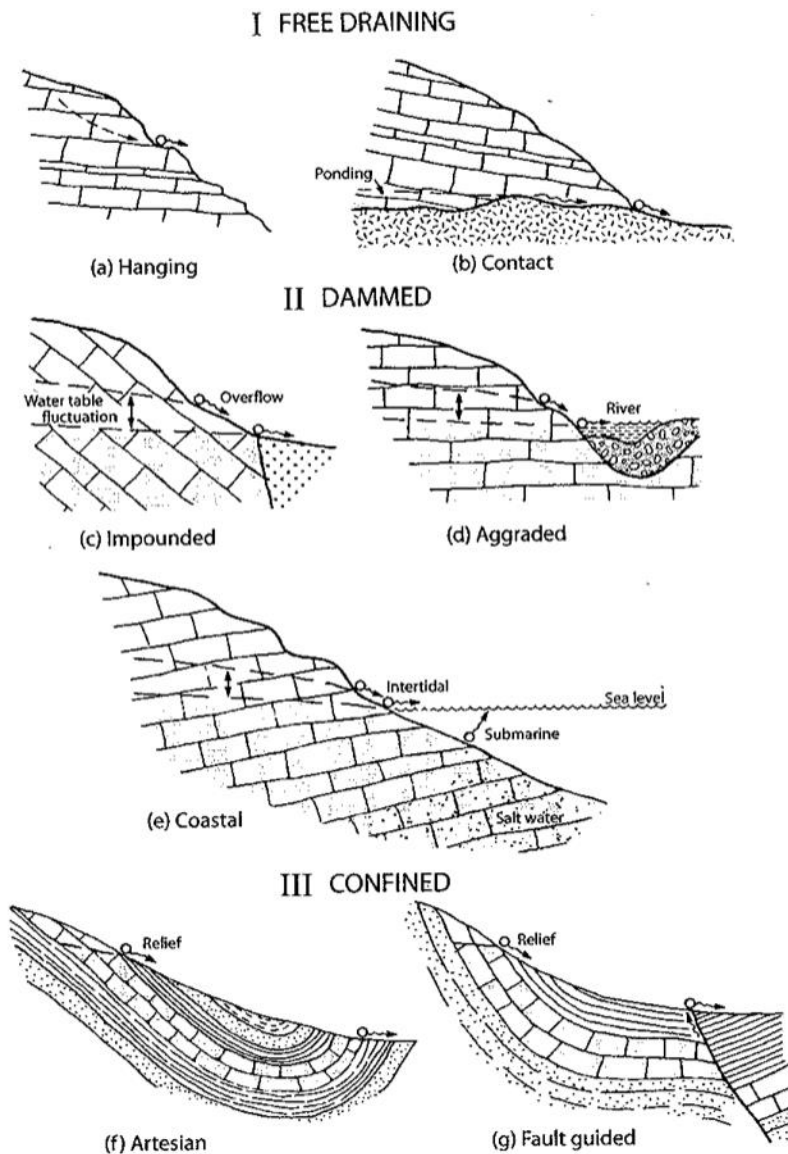


Figure 7. Diagrams of the three different karst spring types (Ford and Williams, 2007).

2.3. Stable Environmental Isotopes

Isotope geochemistry is a discipline that is important in a number of environmental study areas. It provides a dating method, tracers, information on the rate of movement and a manner by which certain chemical processes can be fingerprinted (Porcelli and Baskaran, 2011). Generally elements have more than one isotope as a result of the variation in the number of neutrons that may form a stable nucleus. There are 75 elements that form isotopes that can be useful in environmental studies. These elements can be subdivided into six groups, namely, atmospheric (short lived nuclides), cosmogenic nuclides in solids, decay series nuclides, anthropogenic isotopes, radiogenic isotopes and stable isotopes.

For the purposes of this project the stable isotopes of oxygen ($^{18}\text{O}/^{16}\text{O}$) and hydrogen ($^2\text{H}/^1\text{H}$) will be utilised. The heavier isotope of hydrogen, ^2H , is called deuterium. The

isotopes of oxygen and hydrogen above are believed to date from the formation of the earth and are present in measurable concentrations in all natural waters (Dincer and Davis, 1982).

Oxygen, which is the most abundant element on earth, occurs in thermally stable solid, liquid and gaseous compounds. The isotopes of this element have been studied greatly. The three stable oxygen isotopes arranged from largest to smallest are: ^{16}O (99.763%) $> ^{17}\text{O}$ (0.0375%) $> ^{18}\text{O}$ (0.1995%). The two stable isotopes of hydrogen have one of the largest mass differences between isotopes and are found in almost every environment on earth including the mantle. 99.9844% and 0.0156% are the average of ^1H and deuterium respectively (Hoefs, 1980).

These isotopes undergo a process called isotope fractionation which is due to variations in the distribution of isotopes between different phases or chemical species. The slight variations in isotope concentrations caused by fractionation may allow a deduction to be made on the climate at the time that the rainfall infiltrated into the ground or even the origin of the water (Appelo and Postma, 2010).

Minor changes in stable isotope concentrations can be detected currently using a double inlet, double collector mass spectrometer. The ratio between these isotope concentrations (with the least abundant isotope over the most abundant) relative to a standard is generally represented by the delta (δ) notation (Eq.2) For water the commonly accepted standard against which the ratio is expressed is the Vienna Standard Mean Ocean Water (VSMOW). An increase in δ when using Eq. 2 this notation indicates an increase in the heavier (less abundant) isotope (Appelo and Postma, 2010 and Hiscock, 2009).

$$\delta_{\text{sample}} = \frac{R_{\text{sample}} - R_{\text{standard}}}{R_{\text{standard}}} \times 100 \quad \text{Eq. 2}$$

The isotopes of hydrogen and oxygen in water undergo the same physical processes during fractionation in nature. It is therefore observed that the fractionation of ^{18}O parallels that of deuterium in most cases (Hoefs, 1980).

2.3.1. Stable isotopes in precipitation

The isotopic fractionation in rainwater can be described by the Rayleigh process. The change in isotopic concentrations during condensation is an example of this process. During condensation (from water vapour) water that is more isotopically heavy is produced as a result of the heavier H_2^{18}O molecule condensing more readily than the lighter H_2^{16}O molecule. This would mean that the condensate that is left behind is increasingly enriched in the heavier H_2^{18}O molecule (Appelo and Postma, 2010).

As a result of the proportionality between the $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$ fractionations as well as their proportionality with changing temperatures a cross-plot of the global distribution of $\delta^2\text{H}$ against $\delta^{18}\text{O}$ generally yields a straight line. The straight line relationship between $\delta^2\text{H}$ and $\delta^{18}\text{O}$ is called the World Meteoric Water Line (WMWL) or Global Meteoric Water Line (GMWL) and is described by Eq. 3 (Hiscock, 2009).

$$\delta 2H = 8\delta 18O + 1$$

Eq. 3

There are a number of effects which affect the isotopic composition of precipitation. These effects include the latitude effect, the rainout effect (Rayleigh distillation), the temperature effect, the altitude effect and the amount effect.

The rainout effect involves the depletion of a vapour reservoir (clouds) of stable isotopes as precipitation occurs. The depletion of stable isotopes in the vapour reservoir corresponds to a decrease in condensation temperature (Hiscock, 2009). Due to the strong relationship between temperature and ^{18}O concentrations precipitation tends to be more depleted in this isotope at higher altitudes. The amount effect results in a negative correlation between the quantity of precipitation and the concentration of heavy isotopes. This implies that heavier and more intense rainfalls lead to precipitation which is more depleted (Dansgaard, 1964).

The position of a site in relation to the source of the evaporating water also plays an important role in the isotopic composition of meteoric water. This is called the continental effect and is a result of temperature extremes and topographic effects. The continental effect leads to precipitation and therefore meteoric water that is increasingly depleted in ^{18}O and ^2H as the vapour mass moves away from the original evaporating water mass (Hiscock, 2009).

2.3.2. Stable isotopes in groundwater

As discussed above the different isotopic effects (including seasonal, latitude, altitude, amount and continental) affect the concentrations of isotopes in precipitation. This in turn has an influence on the isotopic composition of the groundwater. These isotopes also appear to be unaffected by interactions with aquifer material and may therefore be regarded as the perfect tracer for groundwater studies (Burger, 2010).

Evaporation is one process that may influence the stable isotope composition of precipitation once it has infiltrated into the unsaturated zone but before it has recharged the aquifer. Higher evaporation rates generally lead to a more enriched isotopic signature. The enrichment of infiltrating water by evaporation is more pronounced in arid areas.

Once the precipitation has been recharged into the aquifer and has been removed from any surface processes (evaporation, fractionation and condensation) the isotope ratios are conservative and cannot be affected by any factors except mixing and increasing temperatures (geothermal processes). Direct infiltration in temperate and humid climates therefore leads to groundwater with a similar isotopic composition to the precipitation (Hoefs, 1980, Gat, 1996 and Burger, 2010).

As a result of the various isotopic effects, isotope fractionation and the fact that precipitation and groundwater generally have a similar isotopic composition there is a large array of applications for the oxygen and hydrogen isotopes in groundwater studies (Baskaran, 2011). Examples of the uses of oxygen and hydrogen isotopes in hydrogeology include (i) identifying the recharge sources of groundwater, (ii) determining mixing patterns

of groundwaters with different isotope compositions, (iii) detecting the interaction between groundwater and seawater or lake water and (iv) paleo-climate reconstruction (Hoefs, 1980, Hiscock, 2009 and Gat, 1996).

Stable isotopes have also been used in the highly variable karst groundwater environments of Europe and Northern Africa where greatly differing discharge volume and complicated tectonic activity make it difficult to establish the actual extent of a drainage area. In karst groundwater systems, stable isotopes have been used to identify groundwater flow paths and connections between sinkholes and springs, identifying flow velocities in various karst aquifers and ascertaining the source of groundwater recharge (Fritz et al., 1980).

3. Materials and Methods

3.1. Climate, Rainfall and Vegetation

The Upper and Lower Fountain springs are both situated within the same climatic region. The climate in this area is known to be subtropical with hot, wet summers and cool, dry winters. Most of the rainfall takes place as sudden thunderstorms caused by convection and the average rainfall in the area is between 700 and 1000mm a year. The annual evapotranspiration in the area is approximately 1600 mm (Kafri and Foster, 1989).

A closer look at the climate in the Pretoria area is illustrated in Figure 2. This bar graph shows the average monthly rainfall (77 years) and evaporation (78 years) for the Department of Water Affairs (DWA) rainfall station at Rietvlei at Rietvlei Dam. The mean annual precipitation in the area is approximately 705 mm with the highest rainfall in January (124.41 mm) and the lowest rainfall in July (4.49 mm). According to the DWA data and the bar graph the evaporation is higher than the rainfall throughout the year in the Pretoria area. The average temperature in the City of Tshwane is between 18 and 20 degrees Celsius (DWAF, 2003).

Grassland and sparse Bushveld shrubbery and trees are the main types of natural vegetation found throughout the area.

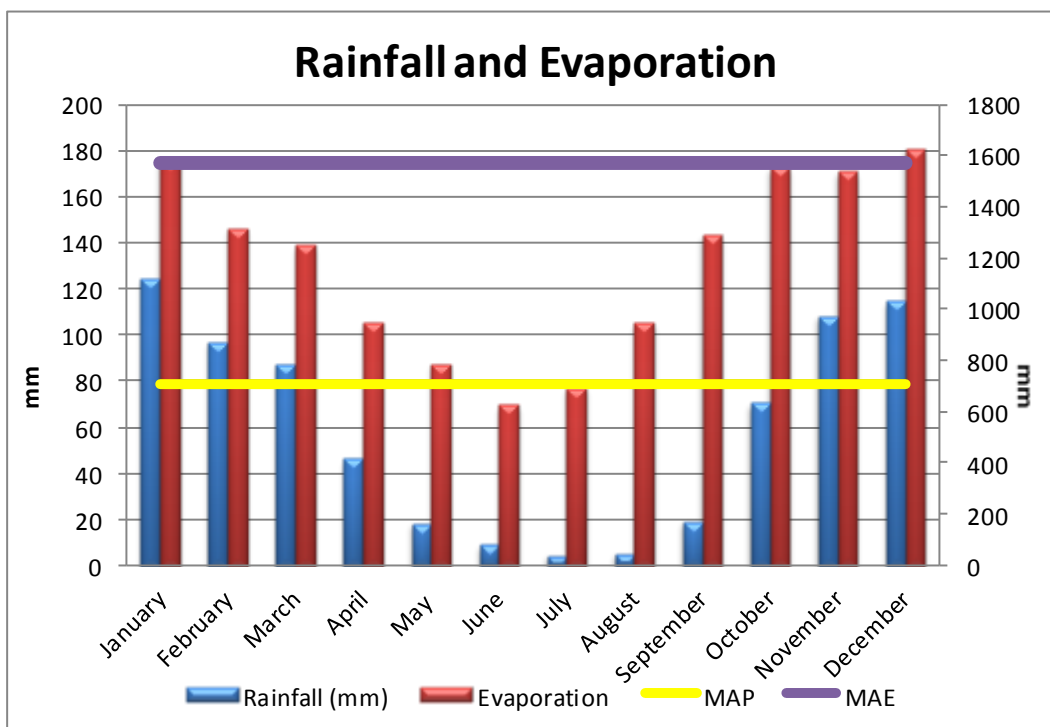


Figure 8. Bar graph showing rainfall and evaporation for the Pretoria area.

3.2. Topography and Drainage

The UF and LF springs fall within Water Management Area WMA 3 which is the Crocodile West and Marico area, now reclassified to form part of the greater Limpopo WMA 1. This catchment is made up of parts of the North West Province, Limpopo and northern Gauteng. The main rivers in this Water Management Area are the Crocodile River and the Marico River, both of which confluence with the Limpopo River. The Molopo River is another important river in this Water Management Area as it is one of the tributaries of the Orange River. Approximately 75 percent of the total runoff in the WMA flows into the Crocodile River, approximately 20 percent flows into the Marico River and approximately 5 percent flows into the Molopo River (DWA, 2002).

The historical WMA 3 is divided into six tertiary drainage regions or sub catchments. The City of Tshwane falls into two of these sub catchments namely the Upper Crocodile and the Apies/Pienaars subcatchments. The subcatchments are further divided into Quaternary Drainage Regions or sub areas of which the City of Tshwane belongs to 10 (DWA 2008 and DWA 2003). These are A23A (Apies-Pienaars: Roodeplaat), A21B-H (Upper Crocodile: Hartebeespoort), A23D and E (Apies-Pienaars: Upstream of Klipvoor).

There are few surface streams on the dolomite rocks which is typical of karst terrains. The runoff is also insignificant in the area which means that most of the precipitation percolates into the phreatic zone. The drainage density on the Malmani dolomites is therefore much lower than that of other groups at 0.58 to 0.60 km/km² compared to 0.84-0.91 km/km² on the Ventersdorp and Pretoria Groups (Kafri and Foster, 1989). The main rivers found in the City of Tshwane area are the Pienaars River and the Apies River (DWA 2002).

There are no natural lakes or large wetlands in the City of Tshwane area but there are a number of existing manmade water structures. The existing water infrastructure in the area includes dams and reservoirs. Dams found within the City of Tshwane boundaries are Rietvlei Dam, Roodeplaat Dam and the Bon Accord Dam (DWA, 2002).

Regionally the topography is subdued and there is a general decrease in elevation from the south to the north. The clastic sedimentary rocks generally form ridges with the east to west striking Daspoort, Timeball Hill and Magaliesberg ranges forming the main topographic features in the area (Vegter, 1990). A topography map overlain on the DEM for the area can be seen in Figure 9. Also displayed on the map are the Upper and Lower Fountain springs and the CDB of Pretoria.

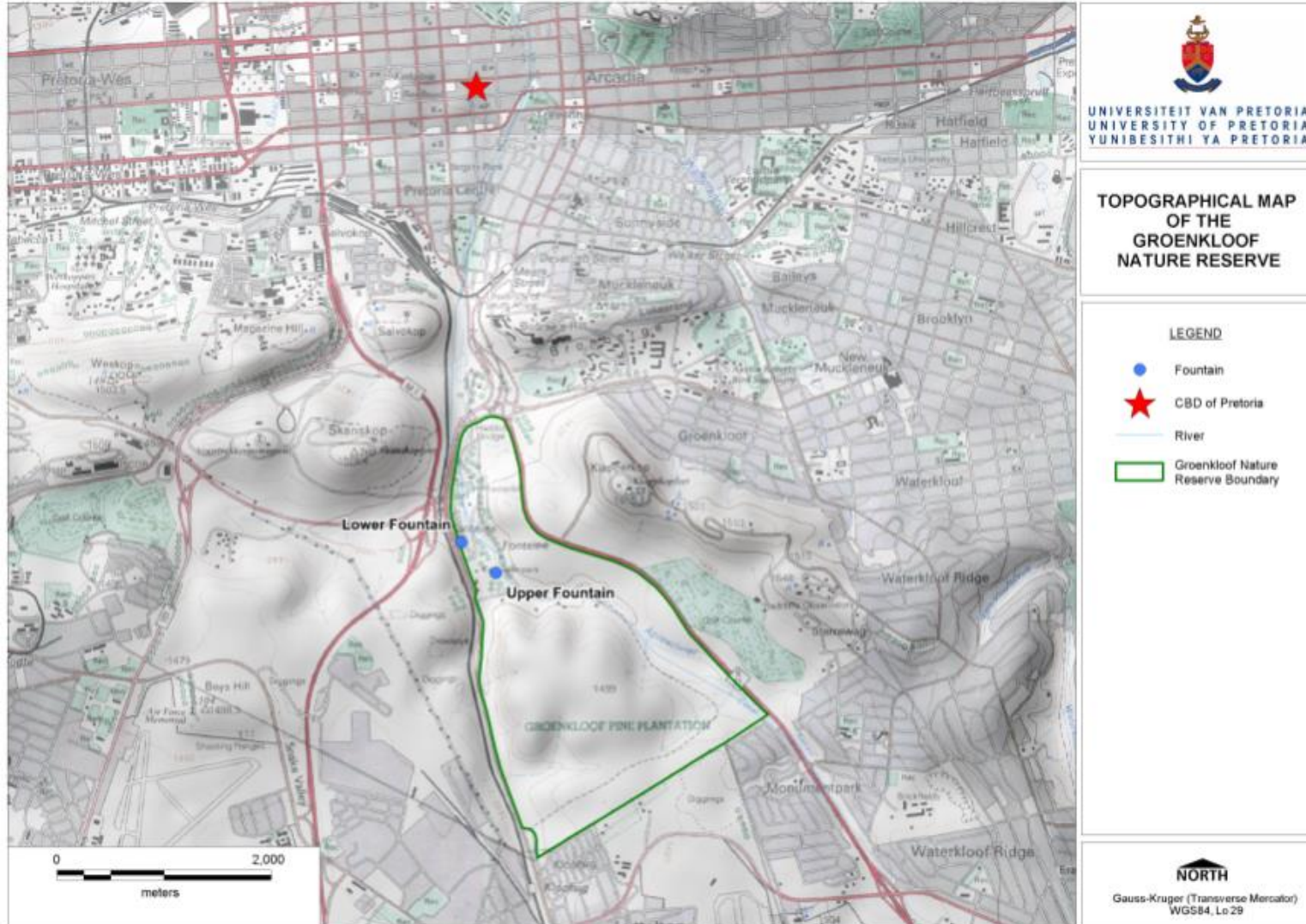


Figure 9. Topography map showing the location of the Upper and Lower Fountain springs.

3.3. Geology

3.3.1. Lithostratigraphy

From the 1:250000 geology map (2528 Pretoria) it is apparent that both the Upper and Lower Fountain springs are situated on the rocks of the Malmani Subgroup which forms part of the Chuniespoort Group. A portion of this geological map can be seen in Figure 10 with the positions of the springs as well as the Central Business District of Pretoria indicated. These dolomitic rocks were laid down in the Transvaal Basin during the Vaalian Era. The Malmani Subgroup is dated at between 2.5 and 2.6 Ga. It is comprised of five formations and is up to 2km thick (Eriksson et al., 2006).

The first formation in the Malmani sequence is the Oaktree Formation. This unit is between 10 and 200m thick and shows a transition from siliciclastic sediments to platform carbonates. Carbonaceous shales, stromatolitic dolomites and some quartzites make up the Oaktree Formation (Eriksson et al., 2006). The dating of a tuff layer in the Oaktree Formation at 2.585 Ga. has lead authors to suggest that there was a significant break between the deposition of the Black Reef Formation and the Oaktree Formation.

The second unit in this sequence is the Monte Christo Formation. This formation is between 300 and 500m thick and moves from an erosive breccia at its base to stromatolitic and oolitic platform dolomite towards the top. Shales, quartzites and stromatolitic dolomites make up the third unit which is called the Lyttelton Formation. The Lyttelton Formation is between 100 and 200m (Eriksson et al., 2006).

Overlying the Lyttelton Formation is the Eccles Formation. This formation can be up to 600m thick and consists of cherty dolomites and includes a series of erosion breccias. These erosion breccias have been mineralised locally by the remobilisation fluids of the Bushveld Igneous Complex and are therefore auriferous (Eriksson et al., 2006). Following the Eccles Formation is an erosion breccia and then the fifth and final unit in the sequence called the Frisco Formation. This formation is up to 400m thick and consists of mainly stromatolitic dolomites at the bottom and moves towards shale rich dolomites towards the top. These shale rich deposits represent the deepening deposition environment in the Transvaal Basin (Eriksson et al., 2006).

The different type of dolomite in these formations influences the amount of karstification and therefore the amount of water in the aquifer. The more chert rich layers such as the Eccles Formation have a large amount of karstification and contain more groundwater (Holland, 2007). It is in these dolomitic formations that Pretoria's fountains are found.

The dolomitic rocks of the Chuniespoort Group are bound to the north by the rocks of the Pretoria Group. The Upper and Lower Fountain springs are situated near the boundary between the two lithologies. The Pretoria Group consists of principally mudrocks and quartzitic sandstones. There are also major basaltic-andesitic lavas, minor conglomerates and diamictites as well as carbonates. All of these rock types have undergone low grade

metamorphism. Seven of the nine formations that make up the Pretoria group can be found in the Pretoria area (Eriksson, 2006). These formations together reach a thickness of up to 6km.

The Rooihoogte Formation forms the base of the Pretoria group. It overlies the palaeokarst topography of the Chuniespoort Group. In the Pretoria area the Rooihoogte Formation consists of well-developed chert breccias (Eriksson et al., 2006).

Thick shales and minor sandstones make up the second formation in the Pretoria group called the Timeball Hill formation. The origins of this formation are attributed to have a fluvio-deltaic basin fill sedimentation system according to Eriksson (1973). It consists of black shales, pyroclastics, mudstones, fine grained sandstones, quartzite as well as diamictites and wackes in the Pretoria area.

The Hekpoort Formation is the third formation. It consists of andesitic- basaltic lavas which were erupted subaerially predominantly as fissure eruptions. These rocks have a thickness of between 50 and 800 m (Eriksson et al., 2006).

In the Pretoria area the fourth formation is the Strubenkop Formation. This formation is comprised of mudstones and alternating siltstones as well as minor immature fine grained sandstones (Eriksson et al., 2006). This formation generally upward coarsening and is probably of lacustrine in origin according to Eriksson et al. (1991).

Overlying the Strubenkop Formation unconformably is the fifth formation in the Pretoria area, the Daspoort Formation. This formation consists of immature sandstones, pebbly arenites, conglomerates and mudrocks. The Daspoort Formation marks the transition between fluvial conditions and the beginning of a major marine transgression (Eriksson et al., 2006).

The Silverton Formation is the sixth formation of the Pretoria Group in the Tshwane area. This formation which is thought to reflect the advance of an epeiric sea consists of high alumina shales as well as high CaO-MgO-MnO shales (Eriksson et al., 2006).

Finally the Magaliesberg Formation overlies the Silverton Formation as the seventh formation in the area. According to Eriksson et al. (1995) the dominant rock type in this formation is immature sandstone which was laid down in an environment with a combination of braided deltas and high energy tidal channels. Therefore this formation would represent the withdrawal of the Silverton Epeiric Sea from the Kaapvaal Craton.

Situated to the north of the rocks of the Pretoria Group (and still within the boundaries of the City of Tshwane) is the Bushveld Igneous complex. The contact aureole of this mafic igneous intrusion has led to low grade metamorphism of the Pretoria Group rocks (Cawthorn et al., 2006). This contact aureole has its outer limits running through the city of Pretoria. The metamorphism caused by the intrusion of the BIC caused the arenaceous rocks of the Pretoria Group such as the Magaliesberg and Daspoort formations to be recrystallised to quartzites (Cawthorn et al., 2006).

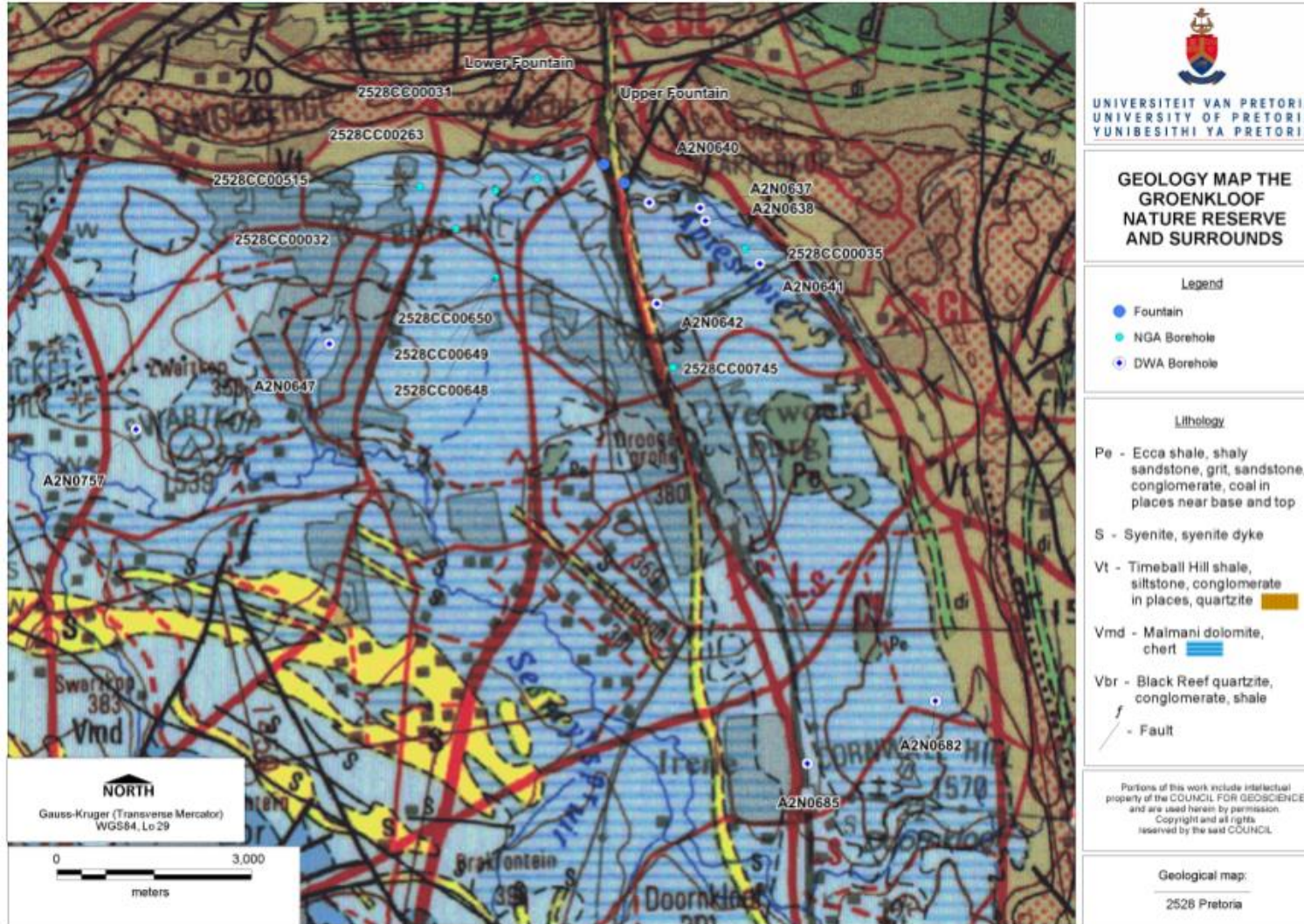


Figure 10. Geological map.

3.3.2. Structural Geology

A large number of dykes, sills and faults are present within the area of the Upper and Lower Fountain springs (City of Tshwane). These structural features vary in size from microscopic to large regional structures and they often play an important role in the movement of groundwater within the different lithologies.

According to Trollip (2006) the dykes and sills in the area are related to the Pilanesberg Alkaline Province. These rocks were emplaced between 1.45 Ga and 1.2 Ga and are largely under saturated with silica (Verwoerd, 2006).

The Pilansberg dyke swarm is quite extensive, ranging from Botswana to the Vaal River (approximately 350km). The predominant rock types of the dykes are dolerite, syenite and nepheline (most of the dykes are composite consisting of both the older fine grained dolerite and the younger syenite in the centre). An example of a dyke from this swarm (composed of syenite) is the Pretoria dyke which cuts through Fountains Valley between the Upper and Lower Fountain spring (Verwoerd, 2006). A younger set of dykes which also forms part of the Pilanesberg Alkaline Complex is the East Rand dyke swarm which is dated at approximately 1.1 Ga. These rocks are normally magnetised compared to the rocks of the Pilanesberg dyke swarm which are reversely magnetised (Foster, 1989 and Verwoerd, 2006).

According to Foster (1989) there have been three main emplacements of dykes in the history of dolomite in South Africa. The first was the Pilanesberg dyke swarm, the second was the East Rand dyke swarm (both of which have been discussed above) and the third was the Post Karoo emplacement. He also stated that sills emplaced in the dolomite were associated with the Bushveld Igneous Complex and a Post Karoo emplacement. The Post Karoo dykes and sills do not however affect the City of Tshwane area greatly as it lies outside the main Karoo basin while the dolerite emplacements were most common (Duncan and Marsh, 2006).

A large number of dolerite dykes and silicified faults that cross cut the Tshwane karst aquifer (Holland, 2007). These dykes are much less permeable than the surrounding dolomite and therefore divide the aquifer into compartments which are discussed further in Section 2.1.5. The approximate trends of the dykes are N-S and E-W while the faults have four major trends (NNW-SSE, NE-SW, SW-SE and NNE-SSW).

The dykes which cause compartmentalisation of the aquifer also control the occurrence of springs or “fountains”. These springs usually occur at the lowest topographically point on the upstream side of a dyke (DWAF, 2006).

Numerous faults can be seen regionally from the 1:500000 geological map (2528 East Rand). These faults cut through both the clastic sedimentary rocks of the Pretoria Group and the chemical sedimentary rocks of the Malmani Subgroup but the majority of faults appear to be within the rocks of the Pretoria Group. There is one fault located in close proximity to the springs. This fault extends from the rocks of the Pretoria Group to the north

towards the Upper Fountain spring. The fault terminates at the Pretoria dyke and has a NE-SW strike.

3.4. Methodology

In order to properly interpret the results of the data acquired it is necessary to construct a number of different graphs and diagrams. These graphs and diagrams need to be explained briefly so that the results and discussions are clearly understood. The graphical methods used allow an overview of a large amount of data at a glance as well as aiding in the identification of trends in the data (Appelo and Postma, 2010).

The first type of diagram that is used numerous times is scatter plots or x-y plots. Most of the scatter plots in this report are time series graphs with the date on the X axis and the concentration (Section 5.3), water levels (Section 5.1), spring discharge volumes (Section 5.2) on the Y-axis. A scatter plot is also used in Section 5.4 to interpret the isotope data.

Another graphical method that is used in this manuscript/study is a bar graph. A bar graph consists of vertical columns with the length of the columns being proportional to the meq/l of the constituent if the graph is being used for the interpretation of chemistry data (see section 5.3) or in a unit of length (e.g. mm) if the graph is being used to interpret rainfall data as in section 3.1. Along the X-axis is the name of the constituent (chemistry) or the month (rainfall).

Pie diagrams are also used in the interpretation of the chemistry data. These circular diagrams represent the proportions of the different constituents as meq/l as subdivisions in the total area. If these diagrams are drawn to scale their radius can be used to represent the concentrations of total dissolved solids of each sample. Pie diagrams are typically used as symbols on maps to show the spatial variations in the hydrochemistry of an area (Hem, 1985). By using the above bar graphs and pie diagrams groundwater with different signatures can speedily be distinguished.

A tri-linear diagram that can be used to represent a large amount of data in an uncluttered and easily interpreted manner is a Piper diagram. The Piper diagram consists of two triangles, one showing the proportions of cations and one showing the proportions anions, both with the units of meq/l. the cation triangle has 100 % Ca on the left corner and 100 % Mg at the top of the triangle while 100 % Na + K falls at the furthest left hand corner of the triangle. 100 % HCO₃ can be found in the furthest right corner of the anion triangle and 100 % Cl is represent on the left hand corner while 100 % SO₄ falls at the top most corner of the anion triangle (Appelo and Postma, 2010).

By extending the two data points in the cation and anion triangles parallel to the outer boundaries until they intersect in the upper diamond diagram the relative chemical composition of a sample can be represented by a single point. If the sample plots in the left hand quadrant of the diamond the water type of the sample is said to be calcium/magnesium

bicarbonate (fresh recently recharged groundwater) and if the point plots in the bottom quadrant the sample has a sodium/potassium bicarbonate water type (deeper fresh groundwater). Waters that plot in the top or right quadrant may have a more varied water type and can either be calcium/ magnesium sulphate or calcium/ magnesium chloride (groundwater that has been affected by contamination) or sodium/ potassium sulphate or sodium/potassium chloride (groundwater that has undergone ion exchange) respectively. The water types derived from the diamond diagram generally form the background for the descriptive terminology of the chemistry of groundwater (Appelo and Postma, 2010 and Hem, 1985).

Schoeller diagrams are the final type of graphical method used to interpret the chemistry data. This diagram represents the concentrations of anions and cations in meq/l as six equally spaced logarithmic scales so that when the constituent concentrations are plotted on the diagram the points are connected by straight lines. The concentrations of the major ions from a sample are shown on this diagram as well as the differences between the various samples that were analysed. The semi-logarithmic scale used in the construction of a Schoeller diagram results in the ratio of the ions in two analysed samples being equal if a line joining two points in one sample is parallel to a line joining the same two ions in another sample (Bowen, 1986).

Piper, pie, and Schoeller diagrams as well as time series graphs of the chemistry data were all constructed using EnviroInsite 7 which is a groundwater analysis tool created by HydroAnalysis, Inc.

4. Data

4.1. Groundwater Levels

Groundwater level data for boreholes in the Fountains East and Fountain West karst aquifer compartments were acquired from the Department of Water Affairs (DWA) and can be found in Appendix A. These boreholes are used by DWA for water level monitoring in the karst aquifer. Altogether 7 DWA boreholes were found within the Fountain East (A2N0637, AN02638, AN02640, AN02641 and AN02642) and Fountains West compartments (AN02647 and AN02757) and water level data was available for 6 of these boreholes. The water level data that are for a 29 year period between 1984 and 2013 for most of the boreholes. Approximately 120 water level measurements were taken during this time period.

4.2. Spring Discharge

Spring discharge data for a number of dolomite springs within the City of Tshwane including the Rietvlei springs, the Sterkfontein springs and the Upper and Lower Fountain springs were supplied by the City of Tshwane. Only spring discharge data for the Upper and Lower Fountain springs were used for this project (Appendix B) and are for the period between July 2007 and September 2012 and include 15 discharge measurements each for the Upper and Lower Fountain springs.

4.3. Inorganic Chemistry

The major ion chemistry data were supplied by the City of Tshwane. The data is for both the Upper and Lower Fountain springs and cover a time period from January 2007 to May 2012 (Appendix C). Data for other springs (Grootfontein, Sterkfontein and Varsfontein) and boreholes (Rietvlei BH1, Rietvlei BH2, Rietvlei BH3, Rietvlei BH4, Rietvlei BH5, Rietvlei BH6 and Valhalla borehole) were also supplied by the City of Tshwane but this data were not utilised as it is not within the scope of this project. Approximately 80 samples were analysed during this period. The constituents analysed include but are not limited to Al, NH₄, Ca, Cl, Cr, F, Mg, NO₃, K, Na and SO₄. Samples were also analysed for total alkalinity, electrical conductivity and pH.

4.4. Stable Isotopes

Isotope data were supplied by Mr Siep Talma and Dr Eddie van Wyk. This isotope data are for the Upper and Lower Fountain springs and the rainfall at numerous rainfall stations in and around Pretoria (Appendix D). The data includes but is not limited to ¹⁸O and ²H data for the rainfall stations and ¹⁸O, ²H, ¹⁴C, ³H and Rn data for the Upper and Lower Fountain springs as well as numerous other karst springs in the Gauteng area (examples of these springs include Elandsfontein, Grootfontein, Sterkfontein and van Reenen spring). Only the

heavy stable isotopes (^{18}O and ^2H) were used for this project as both oxygen and hydrogen are elements fundamental in the water molecule, and as these two stable isotopes have been studied extensively and can be used for a number of different purposes in groundwater studies.

The stations from which the isotope data for rainfall were obtained include Brooklyn (1967 to 1972), CSIR (1972 to 1989), FORUM (1958 to 2001), Lynwood Road (1971 to 1972) and Menlo Park (1996 to 2005). Approximately 250 ^2H measurements and 450 ^{18}O measurements were taken at the rainfall stations altogether for the period between 1958 and 2005. The isotope data for the rainfall stations can be found in Appendix E. For the Upper Fountain spring isotope data were collected from 1970 until 2007 and for the Upper Fountain spring isotope data were collected between 1991 and 2007. There were 10 ^{18}O measurements and 3 ^2H measurements for the Lower Fountain spring taken during this period while there were 7 ^{18}O and 3 ^2H measurements taken for the Upper Fountain spring.

4.5. Borehole logs

Geological logs for boreholes situated within the Fountains East and Fountains West aquifer compartments (less than 3.5 km away from the Upper and Lower Fountain springs) were acquired from the DWA's National Groundwater Archive (NGA). The borehole logs for all the boreholes within the two compartments can be found in Appendix F. This data were used to assess the local geology within approximately 3km of the springs and to assist in the construction of the conceptual model.

5. Results

Results addressed in this section is based on data documented in §4 and have been manipulated minimally with duplicates being removed and appropriated data being grouped together.

5.1. Groundwater Levels

Table 3 summarises the groundwater level data that was acquired from the DWA. This data was supplied in the units metres below ground level (mbgl) but the data has been converted to metres above mean sea level (mamsl) using elevation data for each borehole that was also acquired from DWA. The water levels were converted to these units to illustrate relative differences between adjacent points and for illustrative scale on graphs- in such an instance, elevation accuracy is no longer an issue but relative accuracy is important. The water levels for each year have been averaged from the numerous measurements that were taken each year. Blank cells indicate that no measurements were taken at that particular borehole for the year.

Possible reasons for the lack of data in some years include restricted access or the boreholes being destroyed (which may, for instance, be the case at borehole A2N0682 where no data is available after 2001). It should also be noted that water level data are only available from one borehole in the Fountains West compartment and this may not be adequate for the interpretation of water levels in the compartment. Another consideration that should be kept in mind is that the water levels represented here are not the “natural” water levels in the compartments as the dolomites have been used for anthropogenic activities for more than 100 years and the recharge and discharge mechanisms have therefore been changed significantly.

Using the complete water level data set a statistical analysis of the water levels were done (Table 4). The statistical analysis was done using the data in mbgl and includes the minimum, maximum, average and median water levels for each borehole throughout the monitored period. Also analysed statistically were the changes in water level throughout the period (average change and maximum change).

The time series graph of the water level data for all the DWA boreholes from the end of 1984 until the middle of 2013 can be seen in Figure 11. This time series graph shows the changing water levels for the boreholes A2N0637, A2N0638, A2N0640, A2N0641, A2N0642, A2N0682 (Fountains East compartment) and A2N0647 (Fountains West compartment). The large change in water level in borehole A2N0682 diminishes the visual effect of the water level changes in the other DWA boreholes. This borehole was therefore removed in Figure 12 and a clear graphical representation can be seen of the changes in water levels of the remaining boreholes.

Figure 13 is a correlation graph of the average water levels for the 29 year period between 1984 and 2013 for each DWA borehole in relation to the elevation of each

borehole. The coefficient of determination of the correlation graph is 0.784 and the gradient of the trend line is 0.19.

Average water levels for the period between 1984 and 2013 in the units mamsl for all the DWA boreholes as well as the Upper and Lower Fountain springs (at which the water level was taken as surface elevation) have been compiled into the contour map in Figure 14. The average water levels at each borehole were used to compile the contour map so that a general idea of the groundwater flow direction in the compartments and the gradient (flat or steep) can be assessed. The water levels range from 1375.04 to 1401.04 mamsl and the contours indicate a change in water level of one metre. The lowest heads (approximately 1375 mamsl) are represented by a dark blue colour and move gradually towards a reddish colour at the water levels with higher elevations (approximately 1401 mamsl).

Table 3. Water level and elevation data for DWA boreholes in the Fountains East and Fountains West karst compartments.

Year	A2N0637			A2N0638			A2N0640			A2N0641			A2N0642			A2N0647			A2N0682		
	Elevation (mamsl)	WL (mbgl)	WL (mamsl)	Elevation (mamsl)	WL (mbgl)	WL (mamsl)	Elevation (mamsl)	WL (mbgl)	WL (mamsl)	Elevation (mamsl)	WL (mbgl)	WL (mamsl)	Elevation (mamsl)	WL (mbgl)	WL (mamsl)	Elevation (mamsl)	WL (mbgl)	WL (mamsl)	Elevation (mamsl)	WL (mbgl)	WL (mamsl)
1984		25.5	1374.5		35	1375		10	1380		53	1377					57	1373			
1985														61	1379						
1986		24.76	1375.24											60.15	1379.85		58.41	1371.59		135.74	1374.26
1987		24.53	1375.47											60.95	1379.05						
1988		24.86	1375.14		34.99	1375.01		10.26	1379.74		53.14	1376.86					58.15	1371.85		61.7986	1448.2
1989		24.67	1375.33											62.7	1377.3		58.51	1371.49		130.19	1379.81
1990		24.53	1375.47		34.99	1375.01		10.26	1379.74		53.1	1376.9		60.35	1379.65		58.03	1371.97		137	1373
1991																				130.05	1379.95
1992																				131.76	1378.24
1993																	53.16	1376.84			
1994																	58.4	1371.6		117.667	1392.33
1995		32.055	1367.95		35.04	1374.96		10.29	1379.71								58.585	1371.42			
1996		31.35	1368.65		34.66	1375.34		10.3967	1379.6								54.7367	1375.26		131.22	1378.78
1997	1			1	32.86	1377.14	1	8.385	1381.62	1			1			1			1	130.845	1379.16
1998	4	20.45	1379.55	4	32.74	1377.26	3	8.5375	1381.46	4			4			4	53.5667	1376.43	5	114.405	1395.6
1999	0	22.625	1377.38	1	33.8	1376.2	9	8.9825	1381.02	3			4			3			1	128.91	1381.09
2000	0	21.75	1378.25	0	33.04	1376.96	0	8.51	1381.49	0	50.2733	1379.73	0			0	53.875	1376.13	0	128.89	1381.11
2001		21.52	1378.48		32.92	1377.08		8.61667	1381.38		50.475	1379.53					54.145	1375.86		128.87	1381.13
2002		23.13	1376.87		33.7567	1376.24		9.3375	1380.66		51.175	1378.83					54.2225	1375.78			
2003		23.75	1376.25		34.096	1375.9		9.69444	1380.31		51.827	1378.17					55.52	1374.48			
2004		23.57	1376.43		34.332	1375.67		9.83444	1380.17		52.2289	1377.77		53.08	1386.92		55.49	1374.51			
2005		23.513	1376.49		34.306	1375.69		9.815	1380.19		52.124	1377.88					55.143	1374.86			
2006		22.6855	1377.31		33.9245	1376.08		9.50333	1380.5		51.745	1378.26		47.726	1392.27		54.8682	1375.13			
2007		23.1164	1376.88		34.0455	1375.95		9.60273	1380.4		51.81	1378.19		47.9485	1392.05		55.5427	1374.46			
2008		22.092	1377.91		33.581	1376.42		9.143	1380.86		51.2122	1378.79		57.465	1382.54		54.968	1375.03			
2009		22.0555	1377.94		33.4164	1376.58		9.00364	1381		50.92	1379.08		58.1409	1381.86		54.775	1375.23			
2010		20.9391	1379.06		33.06	1376.94		8.72091	1381.28		50.2592	1379.74		57.72	1382.28		54.456	1375.54			
2011		19.8809	1380.12		32.5336	1377.47		8.25636	1381.74		49.5082	1380.49		57.0809	1382.92		53.3927	1376.61			
2012		20.753	1379.25		32.9322	1377.07		8.58	1381.42		50.056	1379.94		56.882	1383.12		54.119	1375.88			
2013		21.9178	1378.08		33.4056	1376.59		9.03222	1380.97		50.7567	1379.24		57.0978	1382.9		54.3678	1375.63			
	Average	22.4232	1377.58	Average	33.6095	1376.39	Average	9.16883	1380.83	Average	51.098	1378.9	Average	55.7619	1384.24	Average	54.952	1375.05	Average	109.527	1400.47

Table 4. Statistical analysis of groundwater level data for boreholes in the study area.

Station	Depth to groundwater (mbgl)				Change		Record length	
	Min	Max	Average	Median	Average Δ	Max Δ	Start	End
A2N0637	16.90	32.08	22.39	22.35	0.03	10.97	Aug-84	Sep-13
A2N0638	31.78	35.08	33.64	33.62	0.02	2.06	Sep-84	Sep-13
A2N0640	7.56	11.11	9.19	9.13	0.01	2.69	Dec-84	Sep-13
A2N0641	48.73	53.34	51.08	51.09	0.02	2.75	Dec-84	Sep-13
A2N0642	47.30	62.70	55.70	57.14	0.05	7.27	Jan-85	Sep-13
A2N0682	5.60	141.15	108.48	128.90	0.27	130.14	Jun-86	May-01
A2N0647	52.81	58.80	54.99	54.81	0.02	4.91	Oct-84	Sep-13

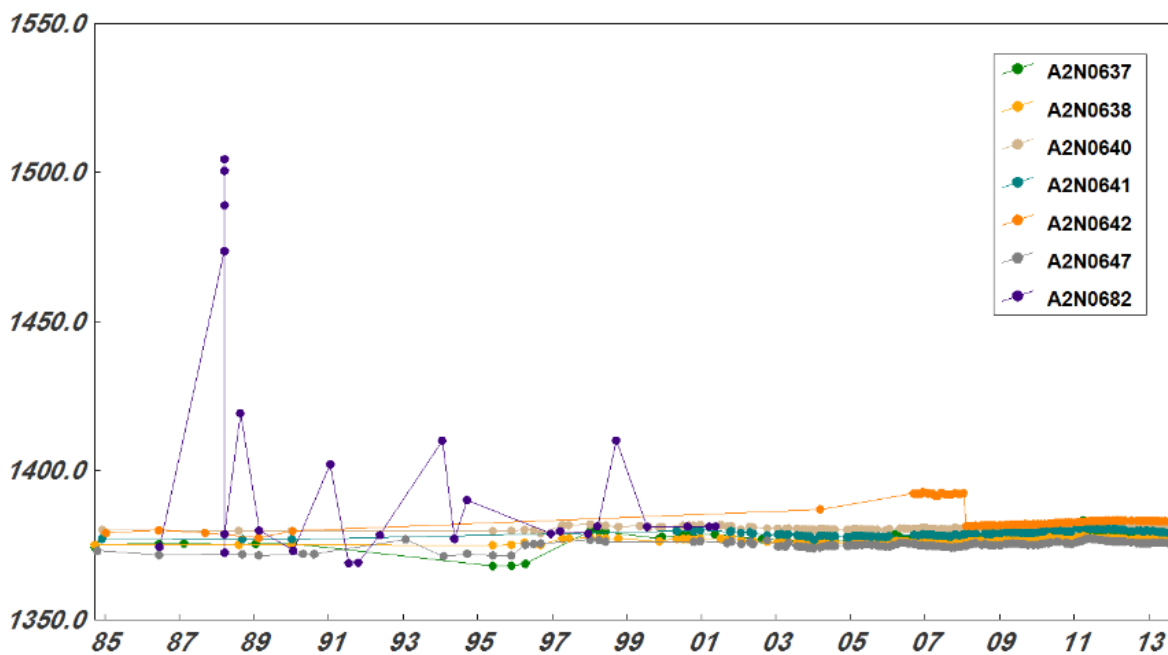


Figure 11. Time series of water levels for all DWA boreholes.

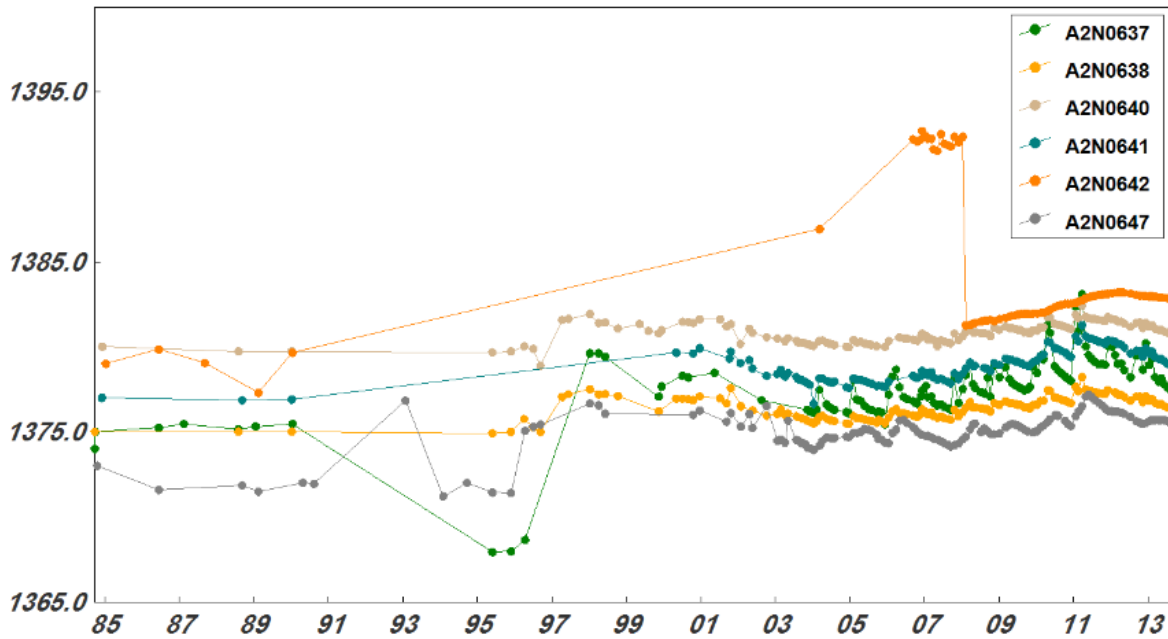


Figure 12. Time series of water levels for all DWA boreholes except A2N0682.

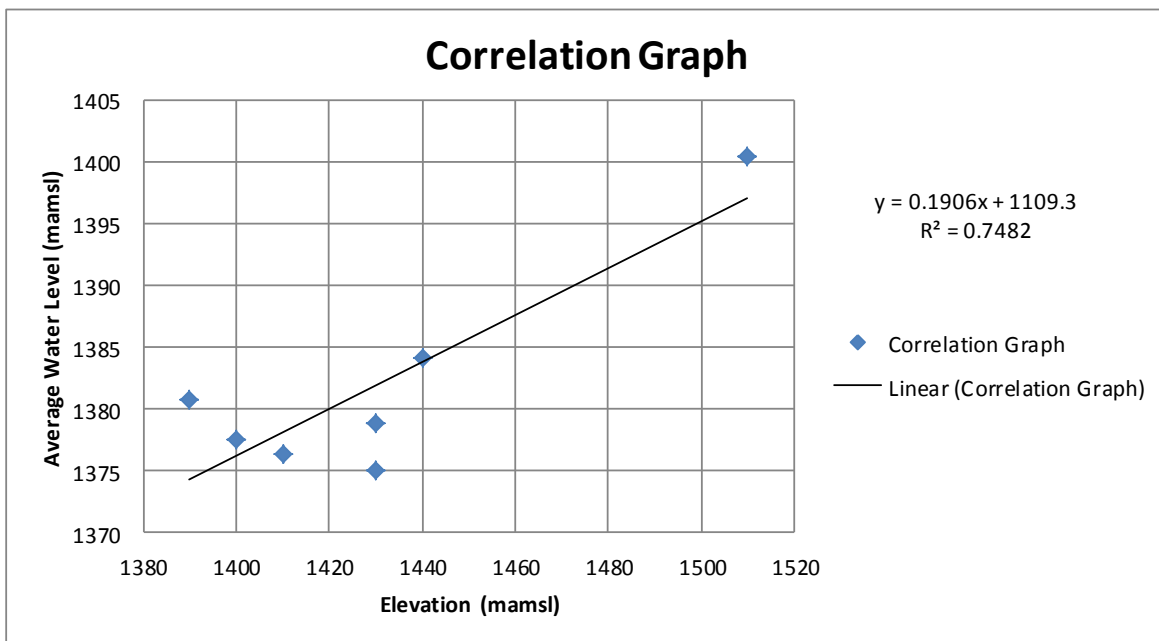


Figure 13. Correlation graph showing the relationship between the average water level and elevation for the DWA boreholes.

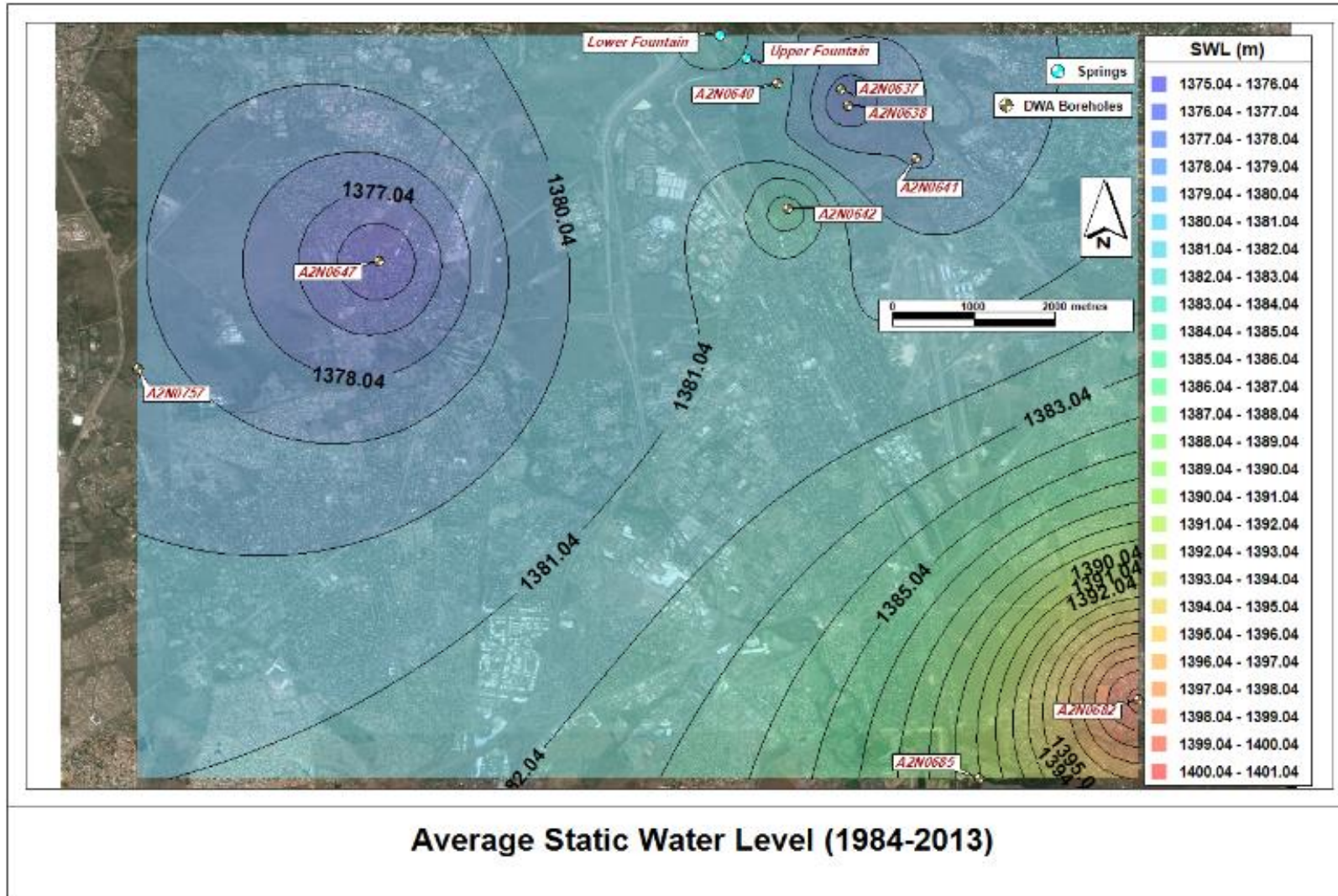


Figure 14. Contour map: average water levels for the DWA boreholes and the Upper and Lower Fountain springs.

5.2. Spring Discharge

Spring discharge volumes for the Upper and Lower Fountain springs in Ml/day can be seen in Table 5. This data is for the period from July 2011 until September 2012 and also includes the total discharge volume and the average discharge volume for each spring.

Table 5. Spring discharge volumes for the Upper and Lower Fountain Springs.

Month	Upper Fountain (Ml/day)	Lower Fountain (Ml/day)
July 2011	23.27	58.69
August 2011	23.21	22.46
September 2011	22.88	45.88
October 2011	22.51	64.40
November 2011	19.09	42.12
December 2011	23.72	15.50
January 2012	19.62	13.11
February 2012	26.38	18.07
March 2012	22.59	15.68
April 2012	22.57	15.48
May 2012	22.21	15.32
June 2012	21.79	14.98
July 2012	21.46	15.07
August 2012	21.13	14.95
September 2012	21.81	15.51
TOTAL	334.24	387.22
AVERAGE	22.28	25.81

Spring discharge volumes for the springs as well as the rainfall and evaporation data for the Pretoria area (Rietvlei at Rietvlei Dam) are shown in Figure 15. This is for a one-year period with higher discharge following the winter months for the Lower Fountain spring while the discharge from the Upper Fountain spring remains fairly constant throughout the year. The rainfall and evaporation data are added to the graph to establish if a relationship exists between the rainfall (recharge) and the discharge from the springs, exists.

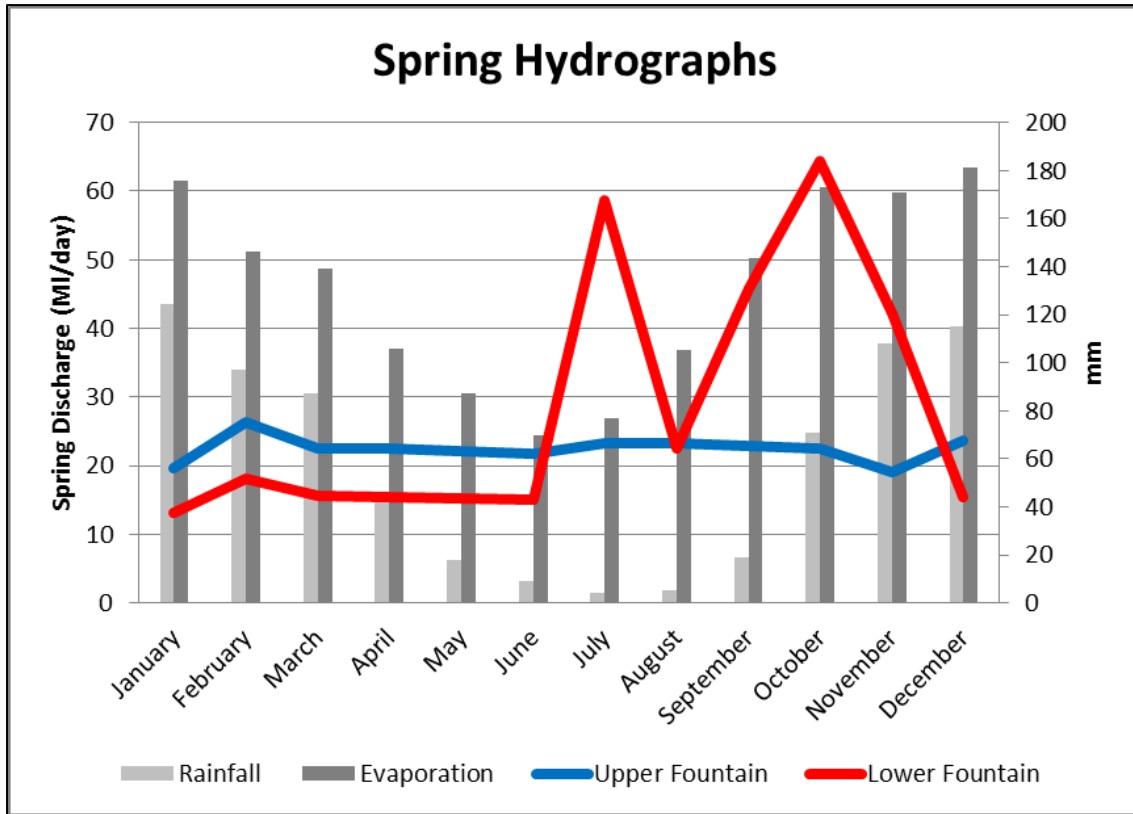


Figure 15. Spring discharge volumes for the Upper and Lower Fountain springs as well as rainfall and evaporation data for the Pretoria area.

5.3. Inorganic Chemistry

Summarised major ion chemistry data for the Upper and Lower Fountain springs for the 5 year period can be seen in Table 6. The average TDS and cation anion balance for each spring can also be seen in this table, the data in Table 6 were used to compile a number of graphs and diagrams in order to assess the spatial and temporal differences between the two springs.

Table 6. Major ion chemistry data.

	Average constituent concentrations	
	Upper Fountain (mg/l)	Lower Fountain (mg/l)
Ca	41.99	46.22
Mg	24.51	26.60
Na	6.08	11.52
K	0.77	1.05
HCO ₃	294.69	280.29
Cl	7.69	23.92
SO ₄	7.88	16.88
NO ₃	1.80	4.65
TDS	270.65	318.29
Cat An Balance (%)	7.33	8.99

The bar graphs in Figure 16 illustrate the average major ion chemistry for each spring. The graphs show that the constituents with the highest concentrations are HCO₃ (approximately 5 meq/l), Ca (approximately 2.25 meq/l), Mg (approximately 2 meq/l), Na (approximately 0.5 meq/l) and Cl (approximately 0.5 meq/l).

Also constructed using the average inorganic chemistry data for the springs are the bar graphs in Figure 17. The average cation concentrations are represented by the red upper portion of the columns and the anion concentrations are represented by the blue lower portions. These graphs allow for the differences between the cations and anions in the each spring and the differences between the two springs to be easily distinguished.

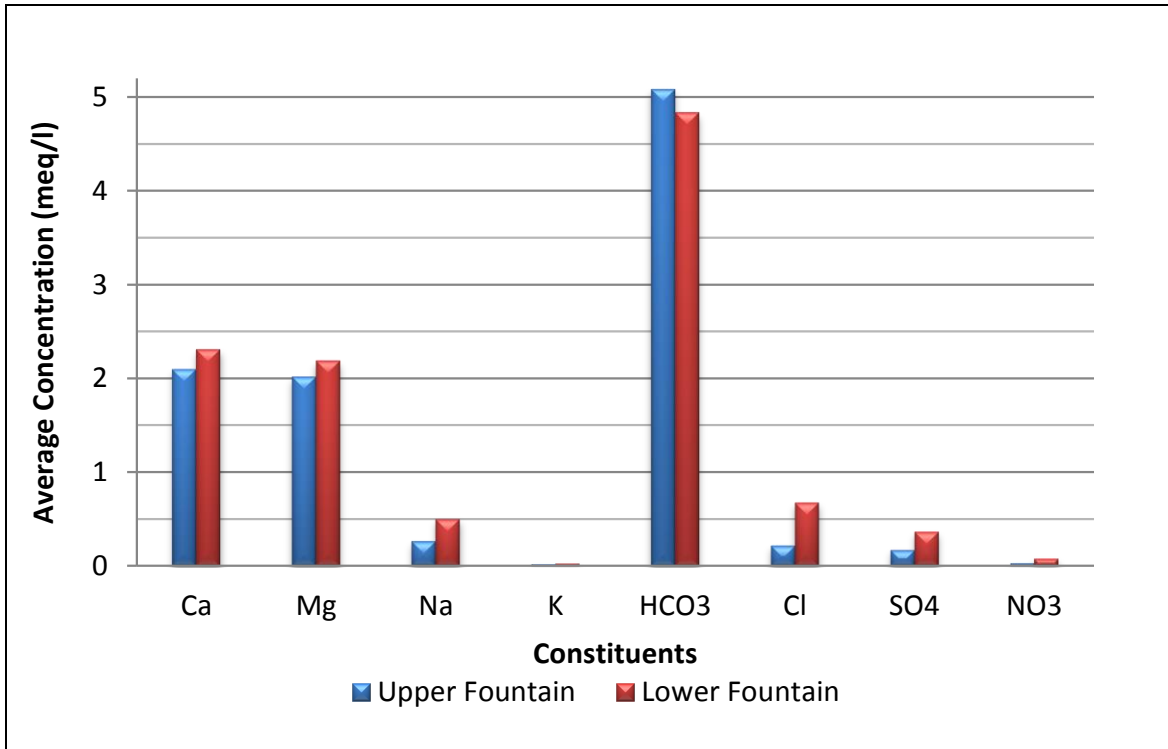


Figure 16. Bar graph of major ion chemistry for the Upper and Lower Fountain Springs

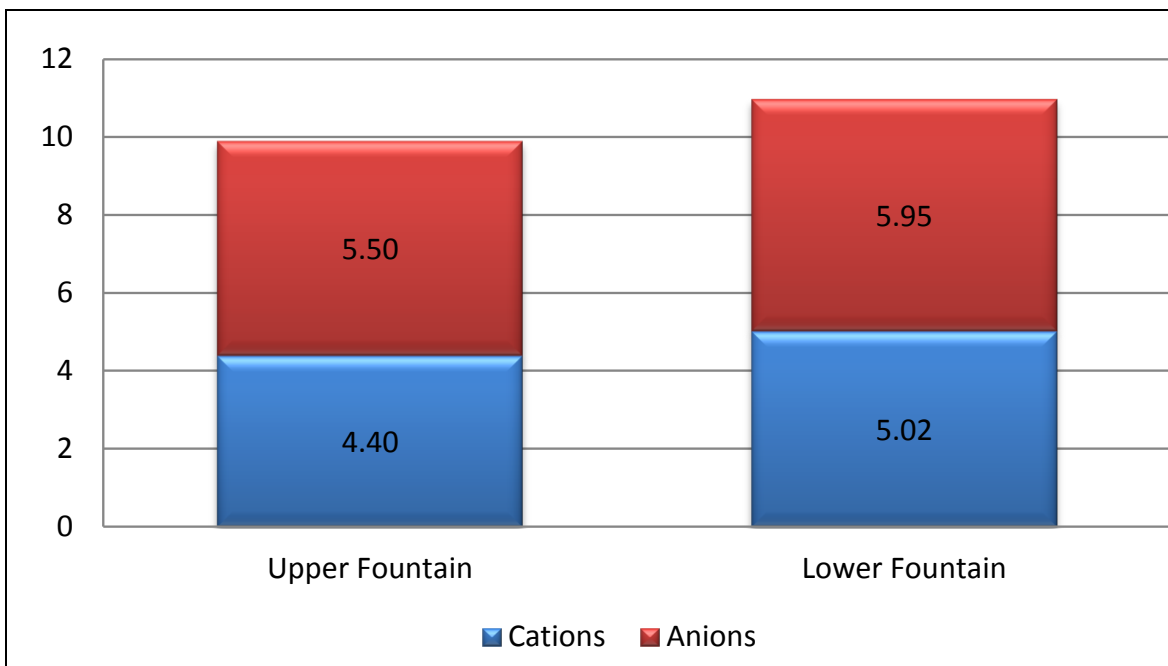


Figure 17. Average concentrations of cations and anions of the Upper and Lower Fountain springs (percentage ion balance indicated above each).

The average major ion chemistry data were also used to assess the hydrochemical water type of the springs. This was done by plotting the major ion data on a tri-linear diagram

called a Piper diagram. Both springs plot in a similar position in the left hand quadrant of the Piper diagram.

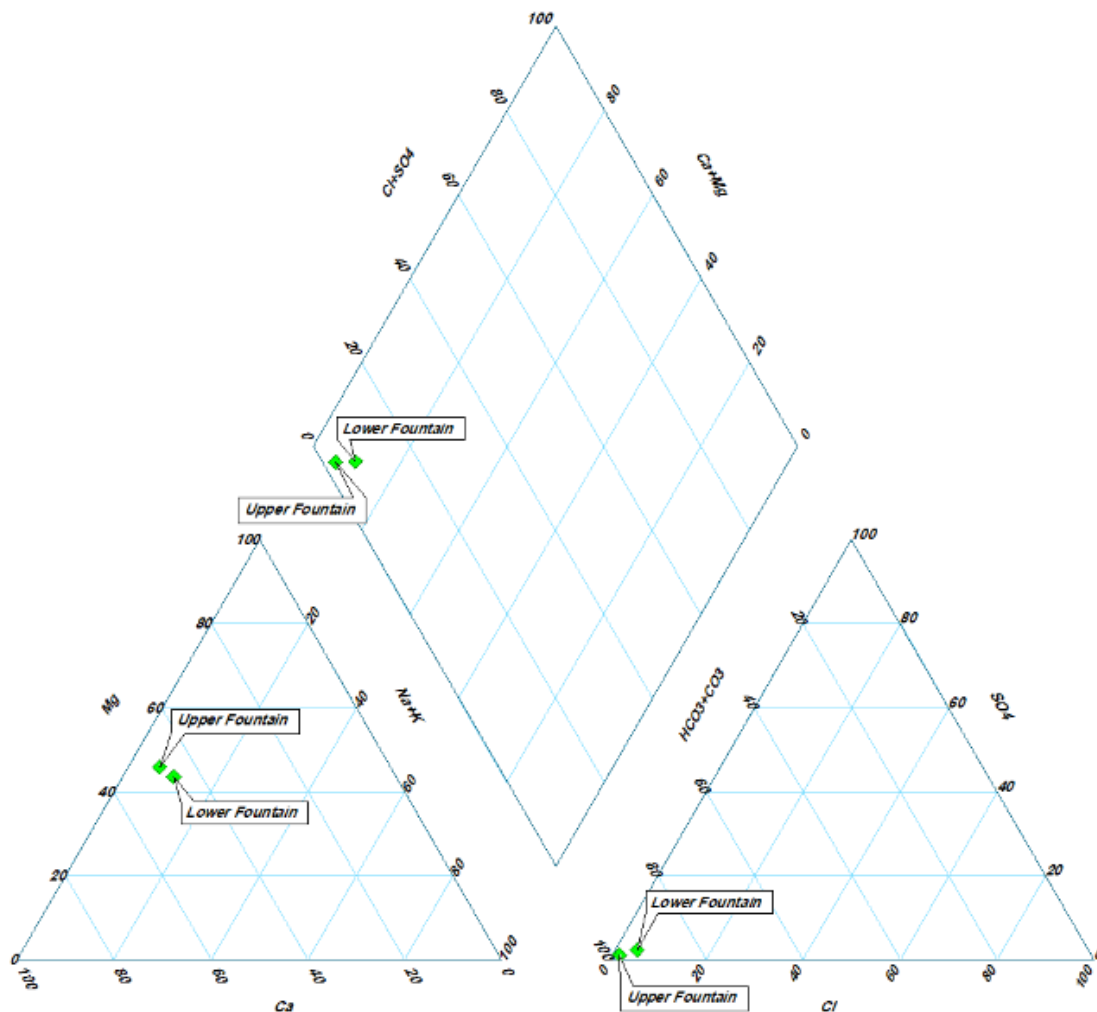


Figure 18. Piper diagram for the Upper and Lower Fountain springs.

In order to assess the temporal variations in the major chemistry at each spring time series graphs of the major cations (Figure 19) and major anions (Figure 20) were compiled.

The major cation time series graph shows the concentrations of Ca (yellow), Mg (red), Na (blue) and K (green) in meq/l for the 5 year period between 2007 and 2012 for both the Upper and Lower Fountain springs. A general order of decreasing dominance in major cations for both springs can be distinguished from the trend analysis: $Ca > Na > Mg > K$.

Figure 19 shows the trend in major anions from 2007 until 2012 for the Upper and Lower Fountain springs. Illustrated on the graph are the trend of HCO_3 (blue), Cl (red), SO_4 (yellow) and NO_3 (green). The trends in Figure 20 indicate that the range of dominating anions is as follows: $HCO_3 > Cl > SO_4 > NO_3$.

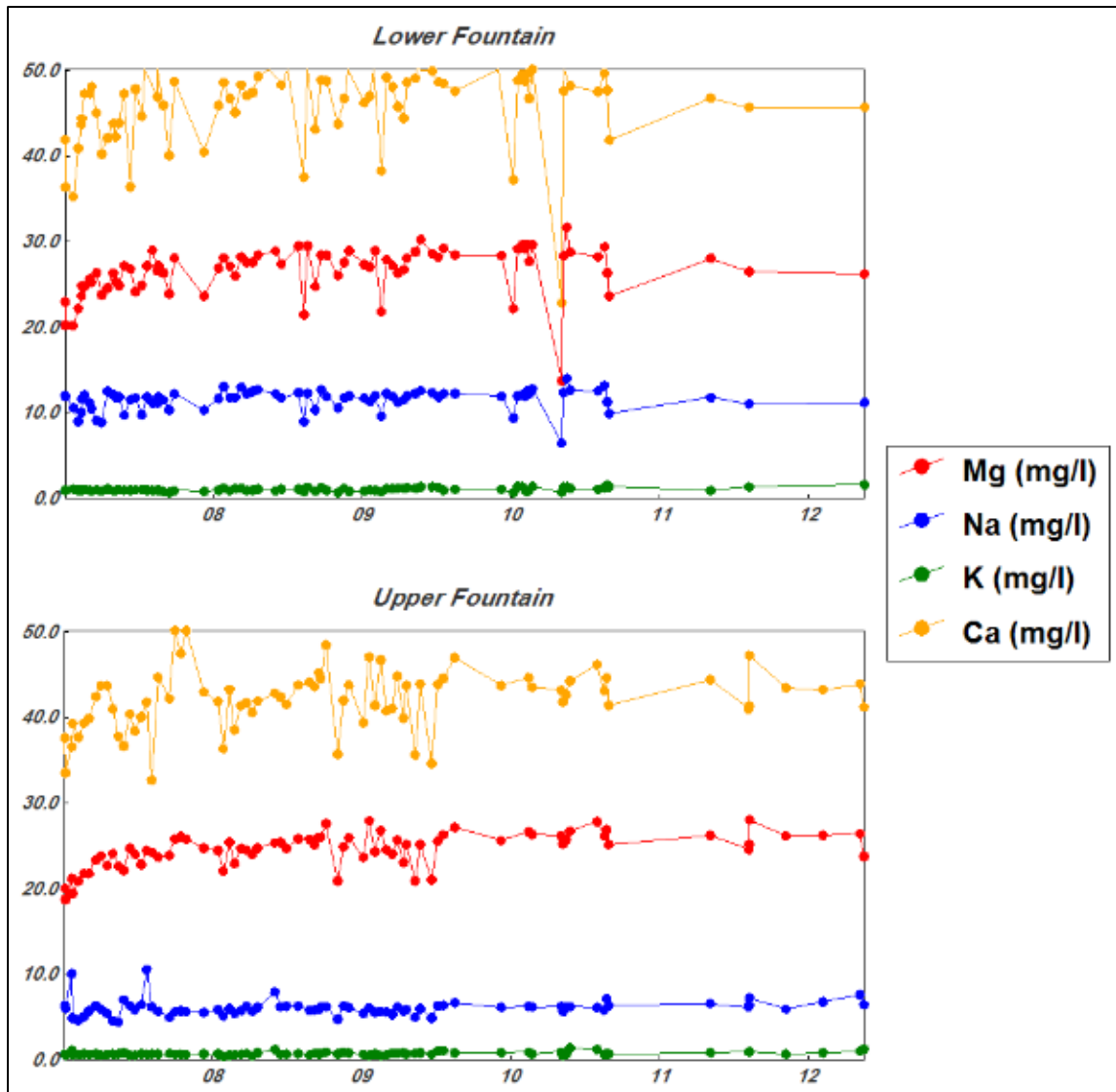


Figure 19. Time series: major cations (2008 – 2012).

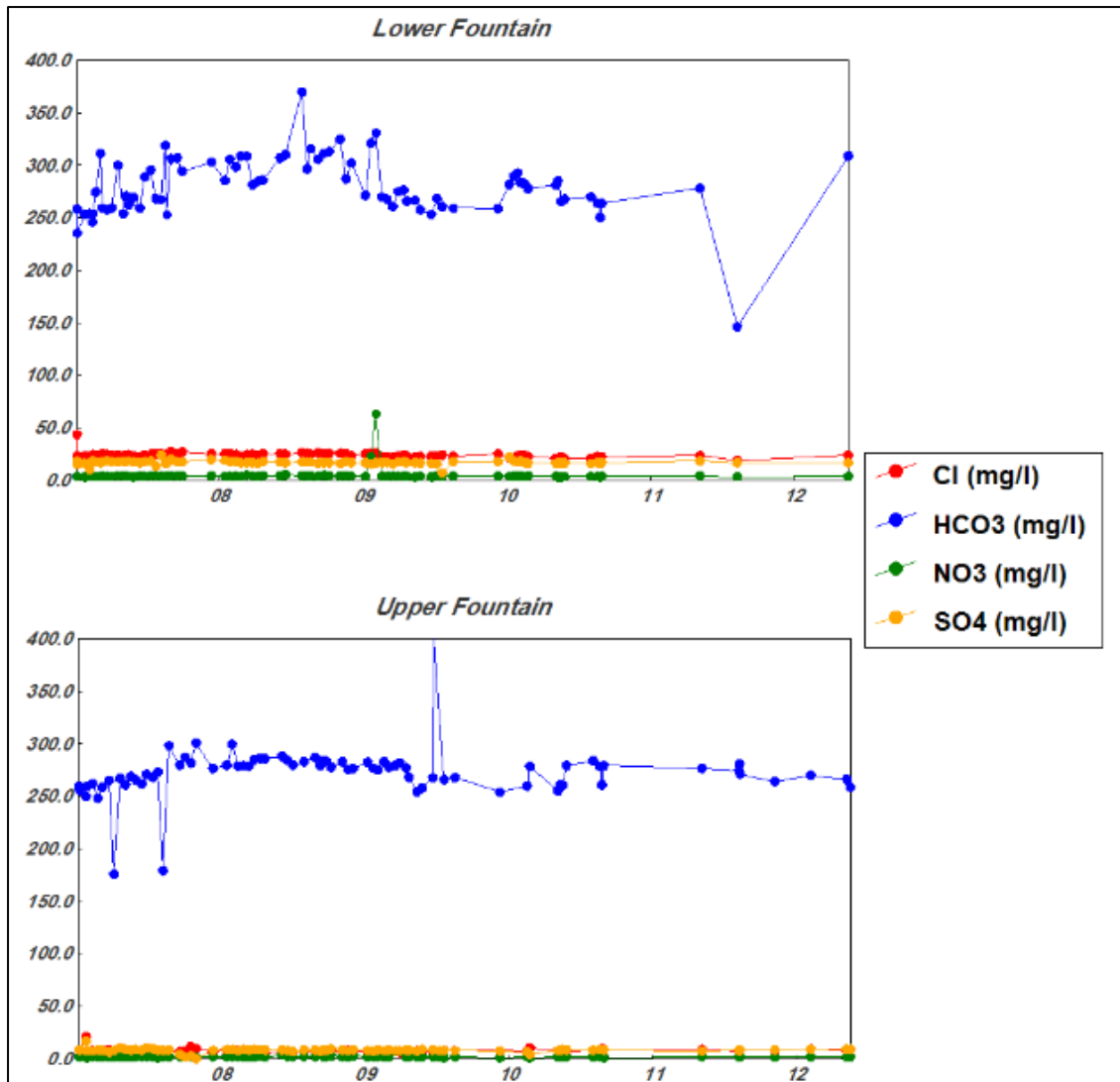


Figure 20. Time series: major anions (2008 – 2012).

Schoeller diagrams are also used to illustrate the temporal trend in the major ion chemistry as per Figure 21 (Lower Fountain spring) and Figure 22 (Upper Fountain spring). The average annual concentrations for each of the constituents in meq/l were used in the Schoeller diagrams to show the change in chemistry of the springs from year to year. The diagrams show the chemistry data from 2007 (darkest coloured graph) to 2012 (lightest coloured graph).

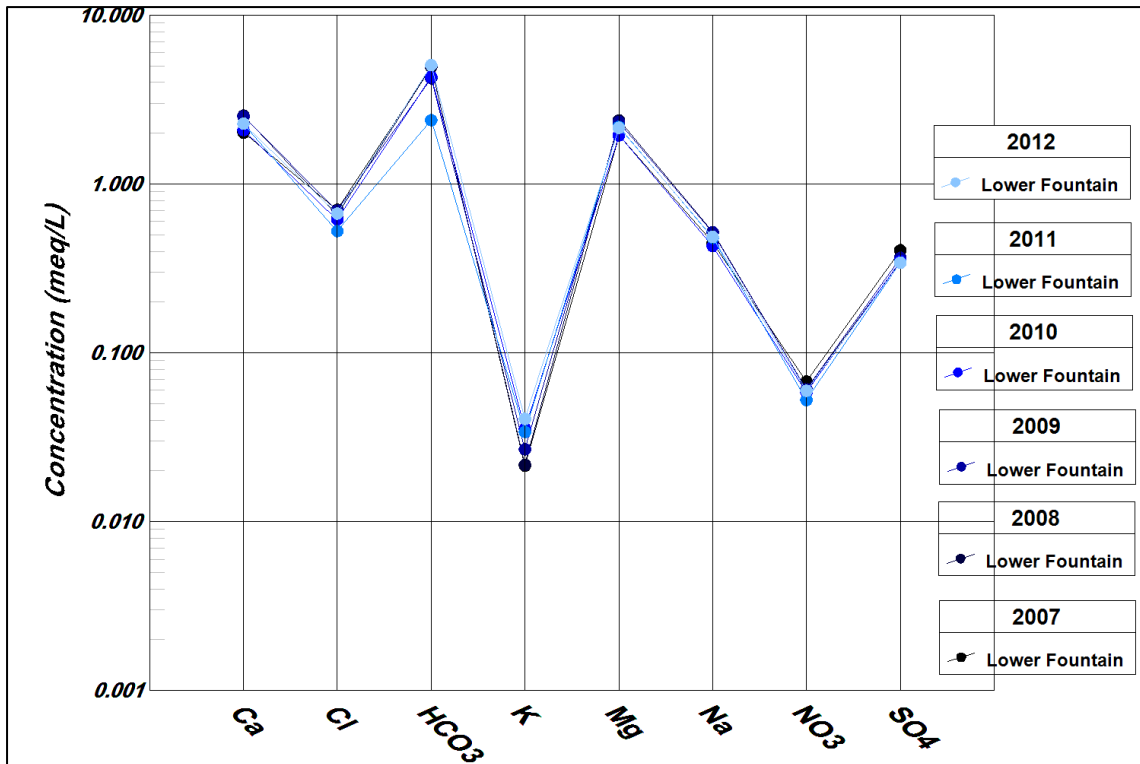


Figure 21. Schoeller diagram for the Lower Fountain spring.

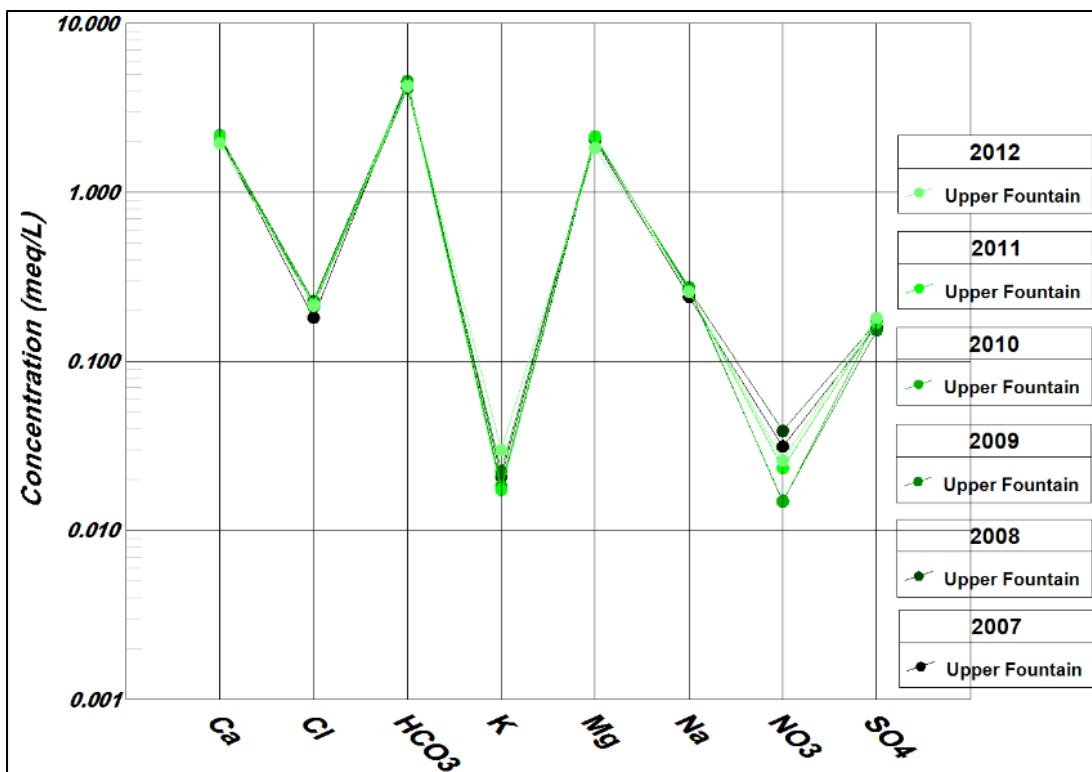
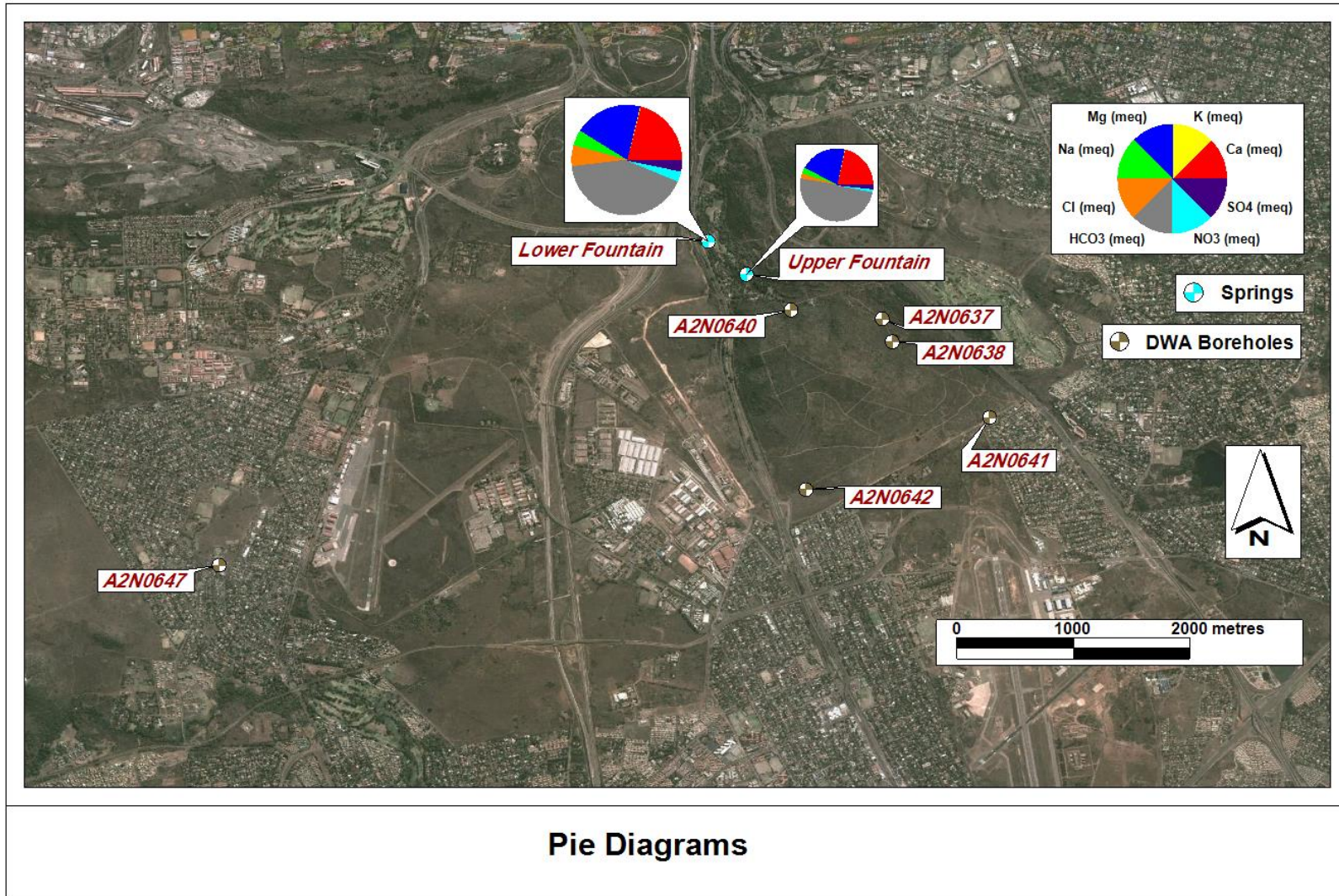


Figure 22. Schoeller diagram for the Upper Fountain spring.

The spatial variation in the major ion chemistry between the Upper and Lower Fountain springs was shown using the circular graphs in Figure 23 called pie diagrams. The pie diagrams (which are plotted on Google Earth imagery overlain on the DEM for the area) show the proportions of Cl, Ca, SO₄, NO₃, Na, Mg, K and HCO₃ relative to each other for each spring. The radii of the circles are governed by the concentration of the TDS. This means that a larger radius indicates a higher TDS, which in turn, is indicative of the degree of mineralisation of the groundwater.



Pie Diagrams

Figure 23. Pie diagrams (meq/l).

5.4. Stable Isotopes

The heavy stable isotope data (^2H and ^{18}O) for the Upper and Lower Fountain springs are shown in Table 7 and are presented in delta notation. The table shows that ^{18}O data is available for both springs from the start until the time of cessation of isotope sampling while deuterium data is only intermittently available during this period.

Table 7: Stable isotope concentrations for the Upper and Lower Fountain springs.

Name	Type	Year	^{18}O	^2H
PRETORIA FOUNTAINS LOWER	Spring	1979	-4.14	
	Spring	1984	-3.82	
	Spring	1991	-3.82	
	Spring	1992	-3.70	
	Spring	1992	-3.77	
	Spring	1993	-3.74	
	Spring	1994	-4.23	
	Spring	1997	-3.92	-23.50
	Spring	1999	-3.58	-18.60
	Spring	2007	-3.20	-18.70
	PRETORIA FOUNTAINS UPPER	Spring	1991	-4.47
Spring		1992	-4.23	
Spring		1992	-4.54	
Spring		1993	-4.30	
Spring		1994	-3.48	
Spring		1997	-4.48	-26.60
Spring		2007	-3.32	-20.81

From the data in Table 7 time series graphs for the stable isotopes in the Upper and Lower Fountain springs were constructed (Figure 24 for $\delta^{18}\text{O}$ and Figure 25 for $\delta^2\text{H}$). As there is a gap in the $\delta^{18}\text{O}$ and $\delta^2\text{H}$ data for the Upper Fountain spring between 1997 and 2007 a dashed line was used to represent this time period. Evaporation data for the period between 1997 and 2007 are presented in Figure 26. This graph was constructed in order to assess the relationship between the heavy stable isotope concentrations and evaporation.

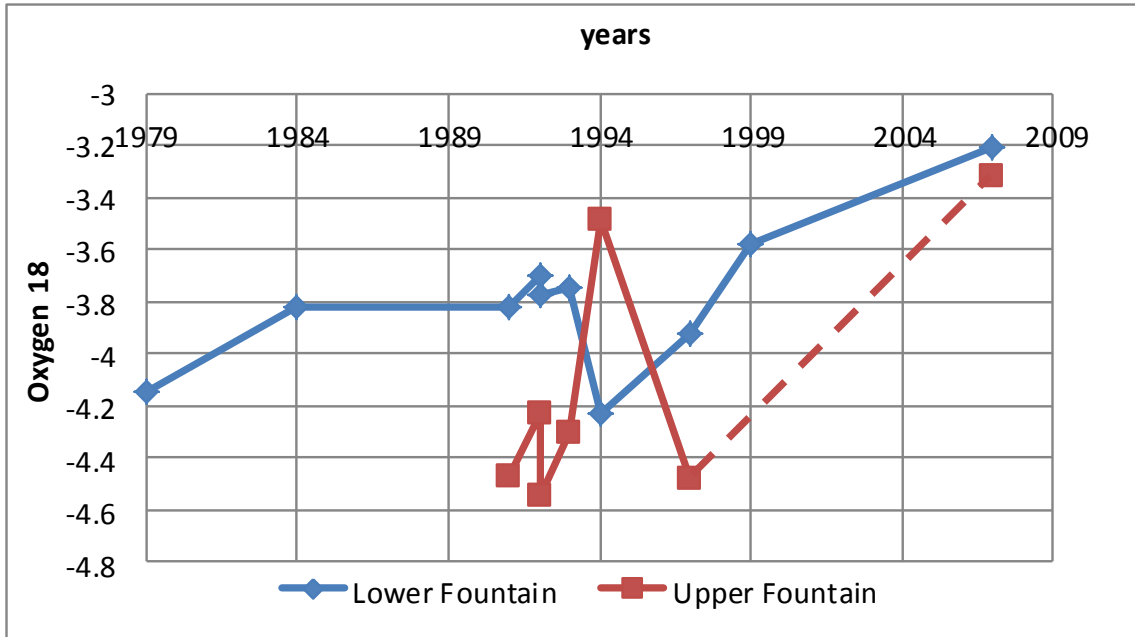


Figure 24. ^{18}O time series graph.

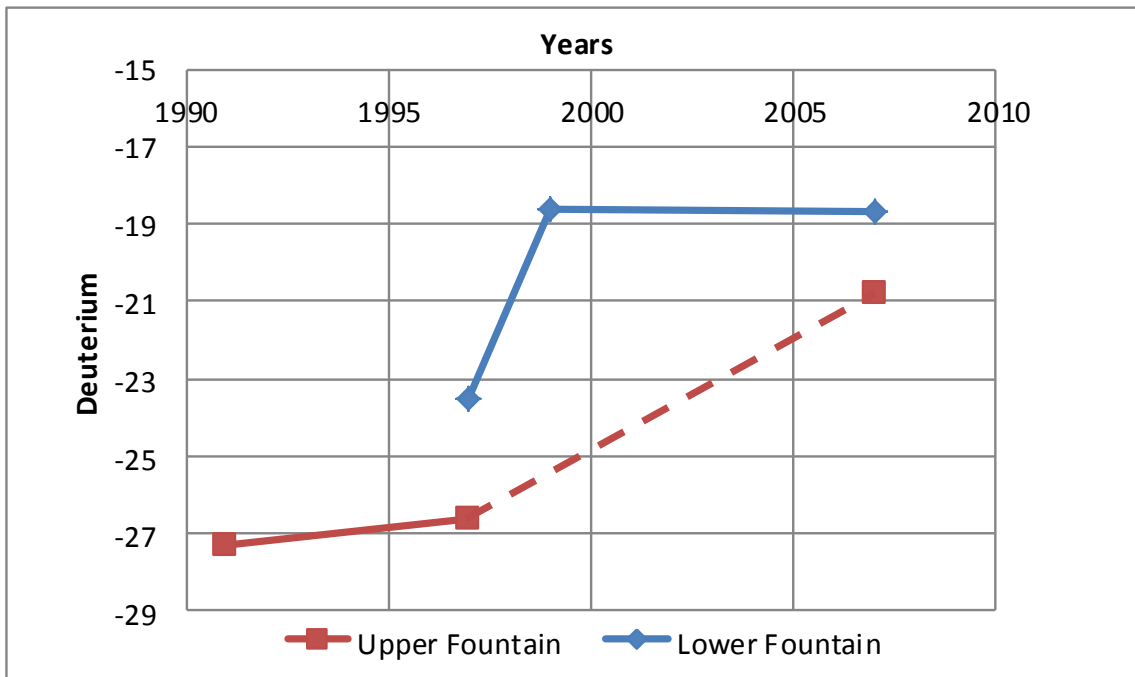


Figure 25. ^2H time series graph.

The evaporation data for the rainfall station Rietvlei at Rietvlei Dam can be seen in Figure 26. This graph was used in order to assess the relationship between evaporation and the change in concentrations of the stable heavy isotopes in the Upper and Lower Fountain springs between 1997 and 2007.

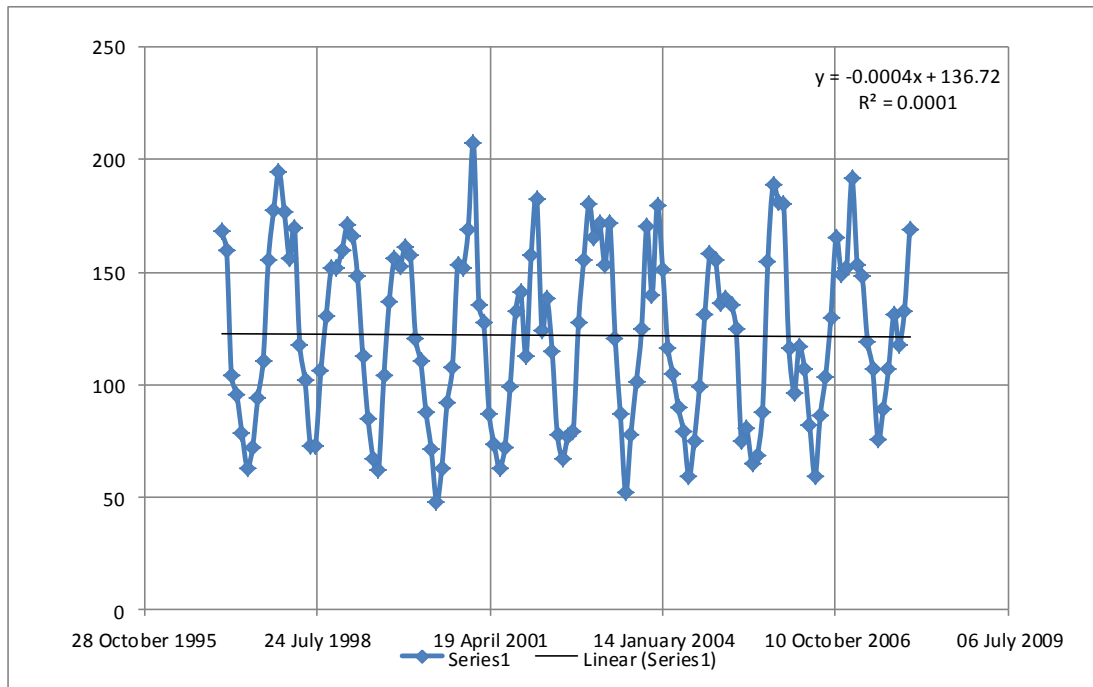


Figure 26. Evaporation data for the rainfall station Rietvlei at Rietvlei Dam (1997-2007).

The temporal data for oxygen and deuterium can be seen is illustrated in the graphs in Figure 27 and

Figure 28 respectively. The graphs show the heavy stable isotope data from 1996 until 2005 for the numerous rainfall stations in Pretoria.

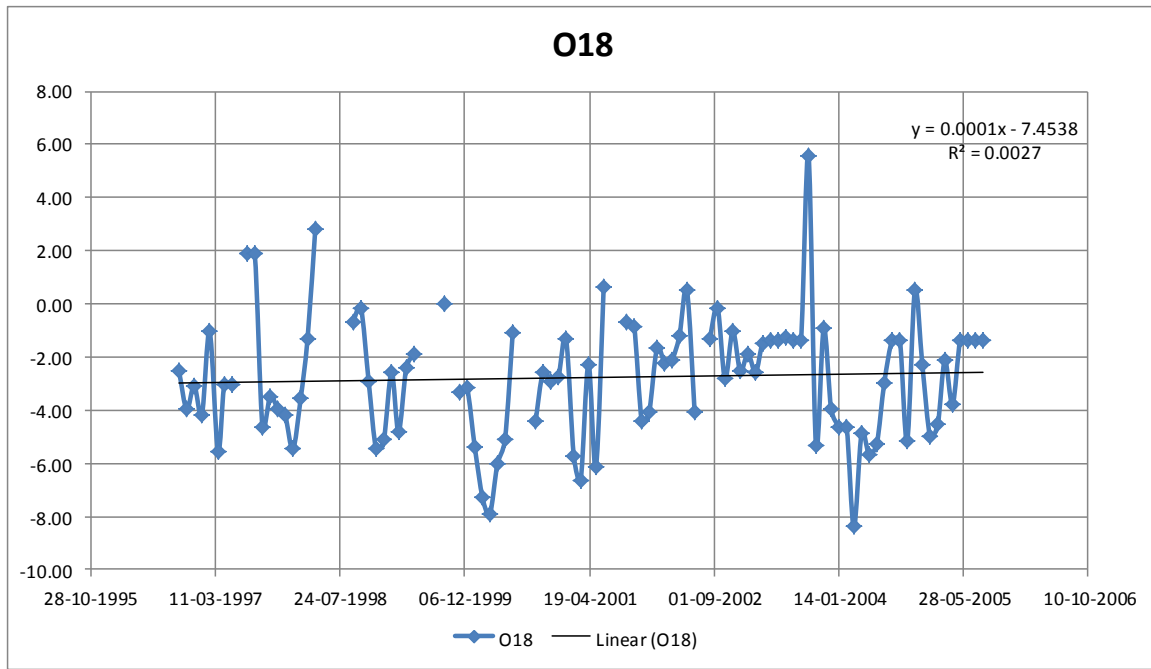


Figure 27. ^{18}O time series graphs for rainfall stations in Pretoria.

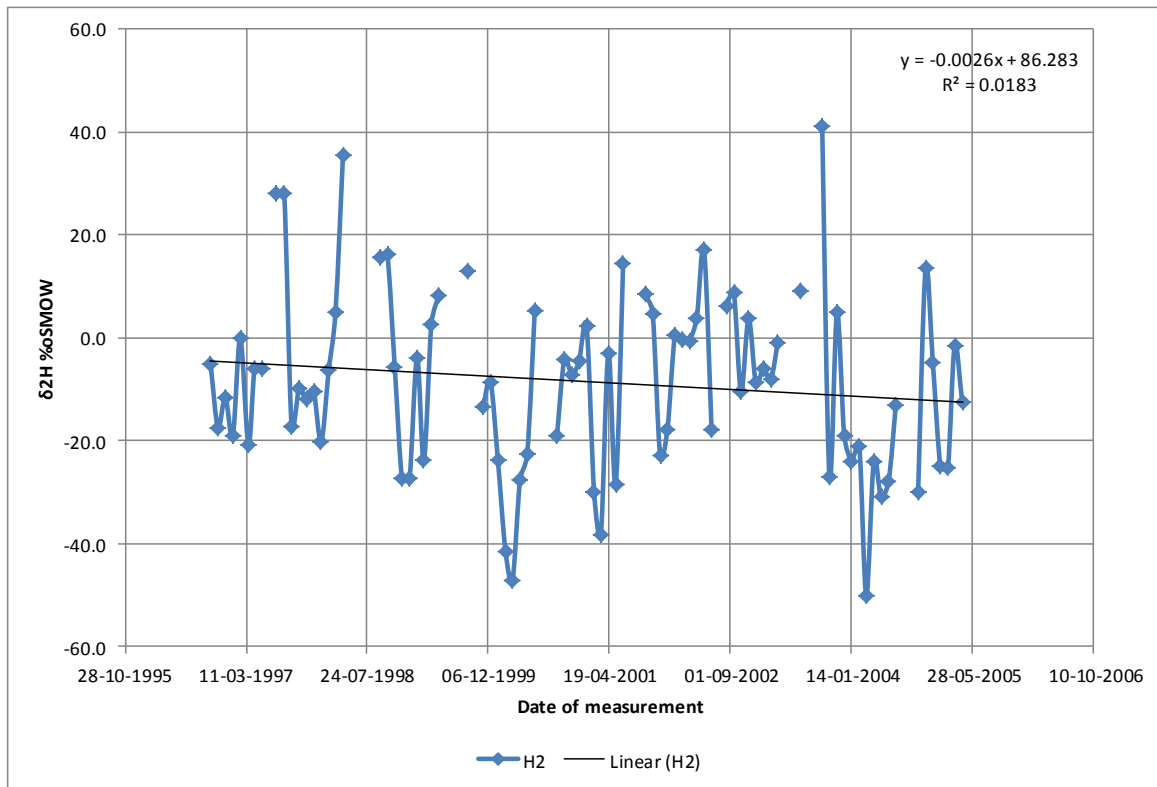


Figure 28. ^2H time series graphs for the rainfall stations in Pretoria.

From Table 7 it can be seen that there were a number of years where $\delta^2\text{H}$ was not analysed for. In order to draw the graph showing the relationship between $\delta^{18}\text{O}$ and $\delta^2\text{H}$ it

was therefore necessary to remove all data points except those with both $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values. The data that remained after removing redundant points can be seen in Table 8.

Table 8. Remaining ^{18}O and ^2H data.

Spring	Type of sample	Year	$\delta^2\text{H}$	$\delta^{18}\text{O}$
Lower Fountain	Spring	1997	-23.50	-3.92
Lower Fountain	Spring	1999	-18.60	-3.58
Lower Fountain	Spring	2007	-18.70	-3.20
Upper Fountain	Spring	1991	-27.30	-4.47
Upper Fountain	Spring	1997	-26.60	-4.48
Upper Fountain	Spring	2007	-20.80	-3.32

By using the data in Table 8 the relationship between the $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in the Upper and Lower Fountain springs were plotted in the graph in Figure 29. Three points for each spring can be seen on the graph as well as the $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values for various rainfall stations in and around Pretoria and the GMWL that was discussed in Section 2.3.1. The deuterium and $\delta^{18}\text{O}$ concentrations are represented as permille with respect to SMOW.

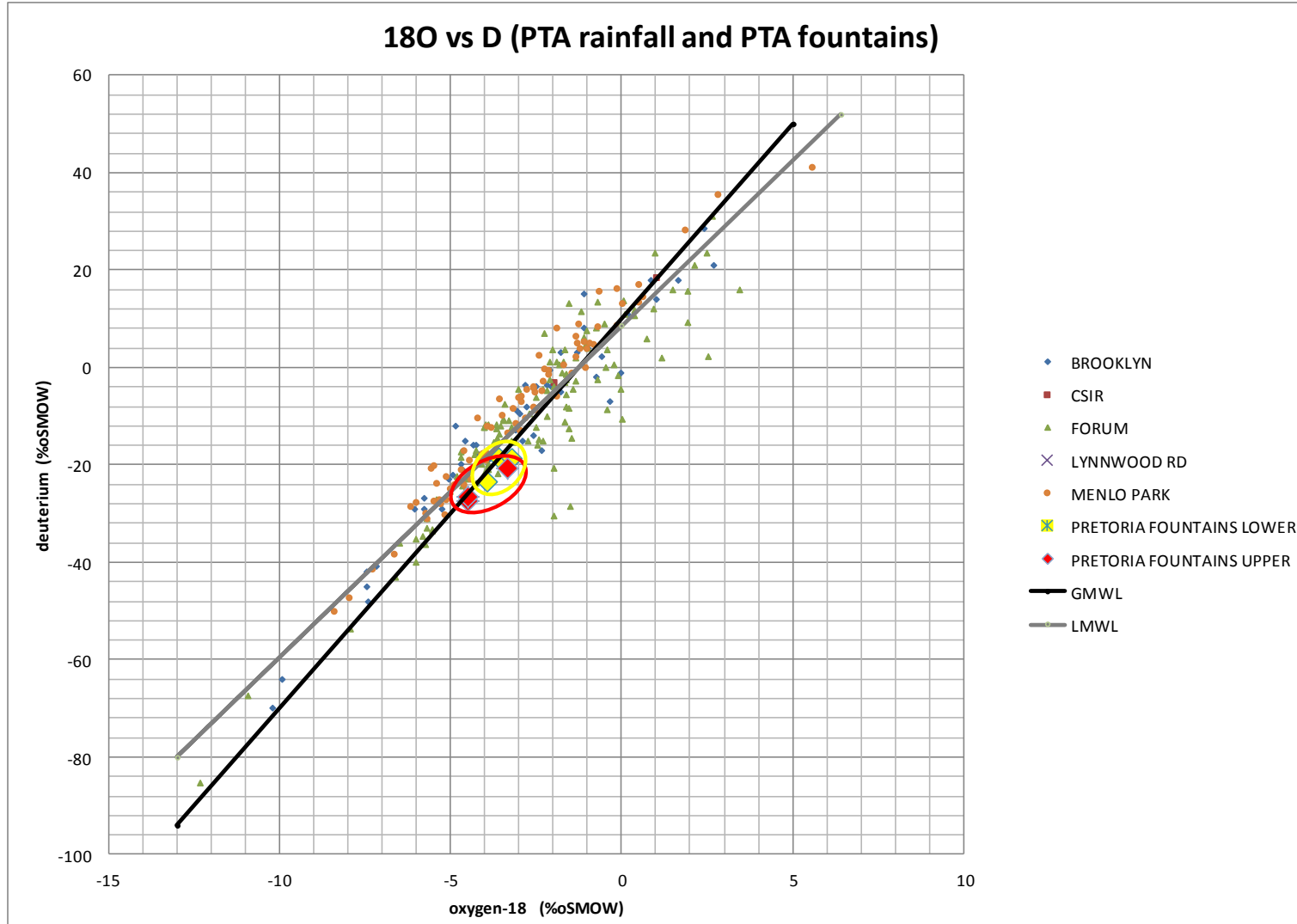


Figure 29. Relationship between 18O and 2H in the rainfall in the Pretoria area and the Upper and Lower Fountain springs.

5.5. Borehole logs and Conceptual Model

A number of simplified borehole logs in the FE and FW compartments are were found in the National Groundwater Archive (NGA). Selected boreholes situated within an approximately 3 km radius of the springs were chosen and a statistical analysis of the data can be seen in Table 9. A simplified local geological understanding from the area can be derived from these logs. From the table it can be seen that the main lithologies encountered in the boreholes were soil, chert and dolomite. The soil horizon has an average depth to bottom of 4.78 mbgl and an average thickness of 4.78 m. The soil horizon was assumed to be composed of weathered dolomite and chert (wad) if situated above the Malmani dolomites and clayey sand where the soil horizon was underlain by the Pretoria syenite dyke.

The depth to bottom of the chert layers ranged between 4 and 32.38 mbgl while the average thickness was 21.17m. The dolomite bedrock was encountered from surface in some areas, to 130 mbgl and had an average thickness 21.37 m.

Table 9. Statistical analysis of NGA borehole logs.

Lithology	Depth to Top (mbgl)			Depth to Bottom (mbgl)			Thickness (m)		
	Min	Max	Ave	Min	Max	Ave	Min	Max	Ave
Soil	0.00	0.00	0.00	2.00	7.62	4.78	2.00	7.62	4.78
Chert	0.00	35.00	11.21	4.00	92.50	32.38	4.00	57.50	21.17
Dolomite	0.00	37.00	17.29	9.00	130.00	38.66	9.00	93.00	21.37

The borehole log statistics, the 1: 250 000 Geological Map Series 2528 Pretoria, the Aster Digital Elevation Model (DEM) for the area as well as the average water level data from Section 5.1 were used to construct the conceptual model of the springs and their respective compartments as shown in Figure 30. The free ASTER DEM version 2 data was acquired from the Japan Space Systems website. According to the website this DEM was created by the Japan-US ASTER Science Team using ASTER data which had been acquired from the beginning of the observation in 2000 until the end of August 2010.

Indicated on the conceptual model is satellite imagery, the surface topography, geology (Pretoria Group, Malmani Subgroup and the Pretoria Syenite dyke), the depth of the soil (indicated in brown), the water table (indicated in blue), the springs (blue stars) and the fault which runs in a NE- SW direction in the Fountains East compartment (indicated in red). In order to show the topography of the area the DEM was used. This DEM was overlain by the satellite imagery for the area. The geological data were digitised from 2528 Pretoria geological map and include the contacts between the Pretoria Group and the Malmani Subgroup which were assumed to have a dip of 15 degrees (deduced from the map) to the NE (Fountains East compartment) and to the NW (Fountains West compartment).

The bottom of the soil horizon was constructed by subtracting the average soil depth within a 3 km radius of the springs from the DEM and the water table was constructed by subtracting the average water level depth (from DWA boreholes) from the DEM.

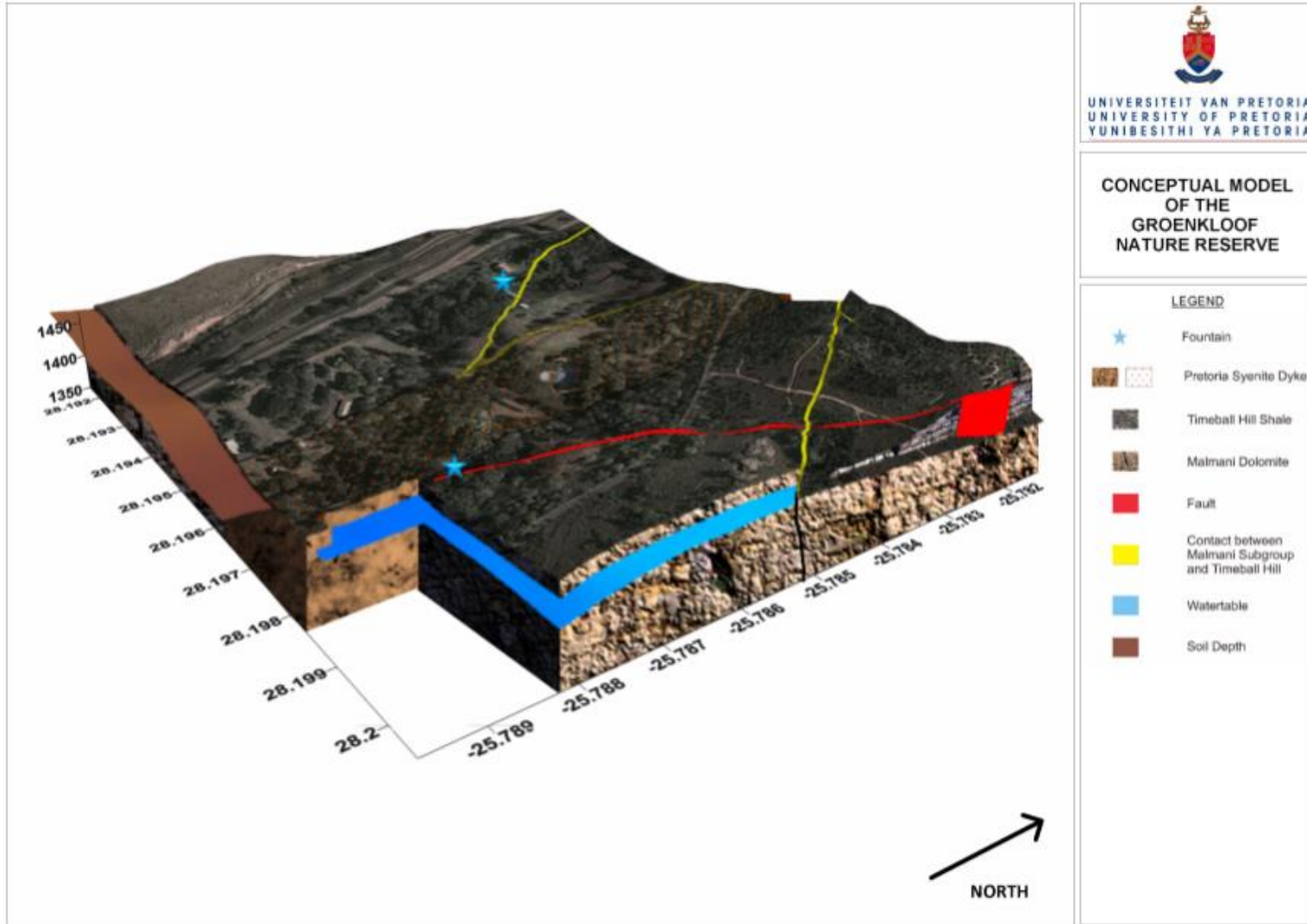


Figure 30. Conceptual model of the Upper and Lower Fountain springs.

6. Discussion

6.1. Groundwater Levels

From the statistical analysis of the water levels in Table 4 it can be seen that the majority of boreholes have a range of less than 6m between the maximum water level and the minimum water level with the exception of A2N0637 (15.18), A2N0642 (15.4) and A2N0682 (135.55). The average and median water levels vary greatly from borehole to borehole with values ranging from approximately 9 to over 100 mbgl. The average water level in the DWA boreholes situated in the Fountains East compartment is 46.47 mbgl and the average water level in the Fountains West compartment is 54.99 mbgl.

All the average consecutive changes in water level for the boreholes are less than 0.5m. This means that on average there are no large changes in the water level at each borehole. The average changes in water level as well as the range of less than 6m between the minimum and maximum water levels in most of the boreholes are expected as karst systems generally have high storage capacities. Excluding the effects of abstraction from the aquifer a large amount of water would have to be added or removed to achieve a moderate change in water level (see Section 2.1.6). The large maximum change in water level in borehole A2N0682 may be a result of the gaps in the data or as a result of abstraction occurring intermittently in close proximity to the DWA borehole. Intermittent abstraction nearby may be a probable reason for large water level changes as the borehole is situated in Irene. There is also a decrease of approximately 11m at borehole A2N0642 at the beginning of 2008. This decrease may be a result of the location of the borehole at the contact between the Pretoria dyke and the dolomite (see Figure 10).

The change in groundwater levels in the DWA boreholes in the Fountains East and Fountains West compartments from 1984 to 2013 can be seen in Figure 11 and Figure 12. These time series graphs shows that borehole A2N0682 which only has data until 2001 has a highly erratic water level. All the boreholes appear to have also had inconsistent water levels between 1984 and 2001 and thereafter the water levels appear to become more comparable between measurements. An exception to this is the water level in A2N0642 which decreases by approximately 11 m in 2008 and becomes more stable thereafter. The erratic appearance of the time series graphs prior to 2001 may be a result of erratic measurement times (differing intervals between measurements) and the large intervals between some measurements (up to 10 years in some cases). After 2001 the water levels in the boreholes increase in depth in the general order: A2N0642 < A2N0640 < A2N0641 < A2N0637 < A2N0638 < A2N0647.

The similar trends followed by the majority of boreholes in the Fountains East compartment and A2N0647 in the Fountains West compartment could possibly indicate that there is water movement between the compartments through the Pretoria dyke. This would imply that the Pretoria dyke is not completely impermeable as many dykes which compartmentalise South African dolomite systems are thought to be.

Another possible reason for the unstable water levels in the 1980s and 1990s is the drought that occurred at that time. During this time large amounts of abstraction would have occurred while very little precipitation would have infiltrated as recharge. After 2003 the water levels appear to stabilise and show a gently increasing trend until 2011 as groundwater levels recovered. From 2011 to 2013 the water levels in all boreholes appear to follow a decreasing trend or remain similar. The boreholes all display a seasonal variation in water level, especially after 2001.

The contour map in Figure 14 shows that the groundwater flow is generally from the south towards the north. The boreholes with the deepest water levels are A2N0647 (Fountains West compartment), A2N0637 and A2N0638 (Fountains East compartment). There appears to be a ridge on shallower water levels between these two deep points running approximately southwards from the Upper and Lower Fountain spring. The ridge of shallower water levels appears to coincide with the Pretoria Dyke. Increased data may, however, change the picture.

A good relationship is usually expected to hold between topography and the groundwater level. This relationship can be used to distinguish between boreholes with water levels at rest, and boreholes with anomalous groundwater levels due to disturbances such as pumping or local hydrogeological heterogeneities. With reference to Figure 13 the relationship between the elevation of the boreholes and the average water levels can be seen to have a coefficient of determination of 7.48. This low value indicates that the water levels in the Fountains East and Fountains West karst aquifer compartments do not correlate well with the elevations in the compartments.

The poor correlation between the elevation and water levels is expected in a karst aquifer as flow is mainly through solutional conduits and there is a large number of heterogeneities present in the aquifer. The highly permeable nature of the dolomite aquifers also causes the water levels to not follow topography in these aquifers and the results are generally an essentially flat water table (DWAF, 2008).

6.2. Spring Discharge

The monthly spring discharge volume data for both springs for the period of a year was totalled and averaged in Table 5. The average discharge volume from the Upper Fountain spring is 22.28 MI/day while the average discharge volumes from the Lower Fountain spring is 25.81 MI/day. The difference in average discharge volumes between the springs is therefore approximately 4 MI/day with the Lower Fountain spring having the higher average discharge. The difference in spring discharge between the two springs may indicate a lack of connectivity between the compartments, or may alternatively be indicative of differences in storage, recharge or other hydrological parameters.

The spring discharge data was displayed as spring hydrographs for the Upper and Lower Fountain spring in Figure 15. Also presented on the graphs are the rainfall and

evaporation data for the Pretoria area. Both springs have similar trends in discharge volume until June with the Upper Fountain spring having a higher monthly discharge than the Lower Fountain. However, after June the Lower Fountain spring has a more erratic monthly discharge than the Upper Fountain spring which has a monthly discharge that has fewer fluctuations throughout the year and is generally lower than that of the Lower Fountain spring.

With relation to the climatic data the relatively stable discharge volume through the year from the Upper Fountain spring and the more erratic discharge from the Lower Fountain spring both show no obvious connection to the higher rainfall months (October to March) and the lower rainfall periods (April to September). This may indicate that the recharge area of the springs is not situated within the Pretoria area. In Section 6.4 the stable isotopes of hydrogen and oxygen were used to further assess whether the recharge area of the springs is situated within Pretoria or not and will be discussed later.

No currently operating abstraction boreholes could be found in the close vicinity to the springs (from the NGA or the City of Tshwane Municipality data) in either the Fountains East or Fountains West compartments. It could therefore not be assessed whether the erratic discharge volumes from the Lower Fountain spring were a result of abstraction taking place in the karst compartments.

6.3. Inorganic Chemistry

The bar graphs in Figure 16 illustrate the average concentrations of the major ions for the Upper and Lower Fountain springs. From the bar graphs it can be seen that potassium is the constituent with the lowest concentration of those analysed while bicarbonate has the highest concentration. The three constituents with the highest concentrations are calcium, magnesium and bicarbonate which is expected as a result of the dissolution of dolomite. This figure also indicates that calcium, magnesium, sodium, chloride, sulphate and nitrate all have a higher concentration in the Lower Fountain spring. From Table 6 it can be deduced that the above ions are respectively 4.23 mg/l, 2.09 mg/l, 5.44mg/l, 16.23 mg/l, 9 mg/l and 2.85 mg/l higher in the Lower Fountain spring. The only ion that is found to have a higher average concentration in the Upper Fountain spring is HCO_3 (14.4 mg/l). The differences in averages are however too minor to indicate any varying processes or factors present in the environment of each compartment.

Figure 17 is a bar graph of the average major cation (blue) and average major anions (red) for the Upper and Lower Fountain springs. The bar graphs indicate that the concentrations of the average major anions in both springs are greater than concentration of the average major cations. This could indicate that the minor cations (cations with a concentration of less than 10mg/l such as iron, manganese or fluoride) make up a significant proportion of the cation concentrations as the concentrations of cations should be equal or similar to that of the anions however the average cation anion balance of the UF is 7.33 % and the LF is 8.99 which is slightly higher than 5 % and less than 10 %. 5 % is generally

seen as the ideal error balance to be below and 10 % is the maximum (Appelo and Postma, 2010 and HP, 1999). The reason for the slight imbalance in cation and anion concentrations may therefore be due to a lab error or the charges in the soil environment. According to Deutsch (1997) anion concentrations that are higher than cation concentrations in a carbonate aquifer are a result of a failure to filter alkalinity samples.

The bar graphs also indicate that the concentration of the major cations in the Lower Fountain is higher than the concentration in the Upper Fountain. The average anion concentration is also higher in the Lower Fountain spring. This means that all together the concentration of the major ions in the Lower Fountain spring is higher in the Upper Fountain spring which was also displayed in Figure 16.

A tri-linear Piper diagram for the Upper and Lower Fountain springs can be seen in Figure 18. This diagram shows that both springs have a Ca(Mg)-HCO₃ signature which is typical of karst groundwater. This groundwater signature indicates a water type of fresh recently recharged groundwater.

Figure 19 illustrates the changes in the concentrations of the major cations in the springs from 2007 to 2012. The concentrations of calcium, magnesium and sodium in the Lower Fountain spring vary seasonally through the 5 year period until the middle of 2010 when the measurements become more infrequent and the concentrations appear to level out from measurement to measurement. The K concentrations at the Lower Fountain spring remain similar at around 2 meq/l from 2007 until 2012. The Ca and Mg concentrations have increased slightly from 2007 until 2012.

In general the concentrations of all the major ions excluding K follow the same trends throughout the 5-year period for the Lower Fountain. This means that when one of the ions shows a decreasing or increasing trend the other major cations follow the same general trend. For example in the beginning of 2010 all the major cations except K decreased dramatically and the concentrations of all these cations increased again in the middle of 2010.

The changes in major cation concentrations for the Upper Fountain are also illustrated in Figure 19. The concentrations of Ca, Mg and Na in the Upper Fountain spring vary seasonally while the concentration of K remain relatively stable from 2007 until 2012 at approximately 1 meq/l. Ca and Mg followed an increasing trend in the Upper Fountain between 2007 and 2012 while the concentrations of Na and K have remained similar overall throughout the 5 year period. There appears to be a correlation between the concentrations of Mg and Ca in the Upper Fountain spring, as one decreases or increases so does the other.

Similar concentrations of all major cations can be seen for both springs overall but the changes in cation concentrations between the two springs appears to be different and the trends followed by the cations seems to be dissimilar also. An example of this can be seen at the start of 2010 when there is a large decrease in the concentrations of Ca, Mg and

Na in the Lower Fountain (between 5 and 25 meq/l) and the concentrations of Ca, Mg and Na in the Upper Fountain do not display any significant variations.

Figure 20 illustrates the changes in major anion concentrations for the Upper and Lower Fountains. The concentrations of HCO_3 vary seasonally and are approximately four times higher than all the other anions in both springs throughout the 5 year period. In the Lower Fountain the concentrations of bicarbonate increase from 2007 to approximately mid-2008 after which the concentrations follow a decreasing trend. There is a large decrease in HCO_3 concentrations in mid-2011 (approximately 100 meq/l) and a subsequent increase at the next measurement in early 2012. The other major anions in the Lower Fountain spring do not display any large variations in concentration and remain similar from 2007 until 2012. An exception to this is NO_3 which increases and then decreases by approximately 70 meq/l at the start of 2009. The changes in concentration of NO_3 in the Lower Fountain spring during that start of 2009 may be a result of human error during the input of data or it may be a result of an anthropogenic source of NO_3 leaching into the aquifer from surface and then that source being removed. Examples of sources of NO_3 include human and animal wastes (faeces). The proximity of these springs to urban areas means that they are highly vulnerable to impacts from human activities.

The concentrations of HCO_3 in the Upper Fountain show a gently increasing trend until the start of 2008 after which they follow a slightly decreasing trend until 2010 where the concentration increases by approximately 25 meq/l and decreases marginally until the last measurement in 2012. In mid-2009 there is a substantial increase in the HCO_3 concentration to roughly 2200 meq/l (not shown in Figure 20). This appears to be a result of a poor analyses or cross-contamination during the sampling processes as only one measurement displays this exceptionally high concentration of HCO_3 and concentrations return to the similar values as those previously recorded at the next measurements. The concentrations of the other major anions do not exhibit any significant variations and all three remain at relatively low concentrations throughout the 5 year period when compared to the HCO_3 concentrations.

From Figure 20 it can be observed that the HCO_3 concentrations in the Upper and Lower Fountain springs follow different trends between 2007 and 2012. A circumstance that exemplifies this difference can be seen in mid-2011 where there is a large and sudden decrease and increase in HCO_3 concentrations can be seen in the Lower Fountain spring while the concentrations in the Upper Fountain remain similar during this time. With regards to the other major anions a similar trend can be seen given that the concentrations at both springs do not vary greatly throughout the 5 year period. The change observed in the HCO_3 concentrations in the Lower Fountain spring may be due to human error during the input of data or due to buffering reactions caused by a change in the pH in the water.

The trends followed by the major cations and anions may indicate very little inorganic contamination in the aquifer and the hydrochemistry is directly linked to dissolution/precipitation in the aquifer. This is supported by the fact that when the chemistry

results for each spring are compared with the SANS 241: 2011 guideline for drinking water both springs comply.

The Schoeller diagrams in Figure 21 (Lower Fountain) and Figure 22 (Upper Fountain) were utilised to show the changes in the yearly concentrations of Ca, Cl, HCO₃, K, Mg, Na, NO₃ and SO₄. The calcium concentrations in both springs are similar but the calcium concentration in the lower fountain spring varies more from year to year. The nitrate concentrations in the Lower Fountain spring are higher than those found in the Upper Fountain spring while the bicarbonate concentrations appear to be similar in both springs with more variation in the Lower Fountain spring. Potassium and magnesium concentrations of both springs appear similar while there is a slightly higher concentration of sodium in the Lower Fountain spring. The nitrate concentrations in the Lower Fountain spring start at approximately 0.08 meq/l (2007) and remain similar throughout the 6 year period whereas the nitrate concentrations in the Upper Fountain spring begin at approximately 0.5 meq/l (2007) and vary approximately 0.25 meq/l.

In order to evaluate the spatial variation between the two springs pie diagrams were plotted onto Google Earth imagery as can be seen in Figure 23. From the diagrams it can be observed that HCO₃ is the major constituent in both springs but the other ions in the pie diagrams vary at each spring. Furthermore, these diagrams show that Cl is more abundant in the Lower Fountain spring than in the Upper Fountain spring. The SO₄ in the Lower Fountain spring also appears to be greater than in the Upper Fountain spring. The differing radii of the pie diagrams indicate that the average TDS at the Lower Fountain is larger than at the Upper Fountain. The higher constituent concentrations and TDS in the Lower Fountain spring indicates that the water emerging from this spring is more mineralised than at the Upper Fountain spring. This may be due to more dissolution occurring within the Fountains West compartment, differing chemical reactions occurring within the two compartments or it may be due to higher evaporation occurring within the Fountains West karst compartment.

6.4. Stable Isotopes

Stable isotopes were used to help with conceptualising the differences and similarities between the two springs and the compartments to which they belong, in conjunction with the hydrochemical and spring discharge data.

The scatter plot in Figure 24 is a time series graph of the $\delta^{18}\text{O}$ data for more than 20 years for both springs. The graphs show that the oxygen 18 concentrations in the Lower Fountain followed a generally increasing trend between 1979 and 2007. This means that the Lower Fountain spring has become slightly more enriched in oxygen 18. The $\delta^{18}\text{O}$ in the Upper Fountain appear to follow a more irregular trend than the Lower Fountain but overall seems have been less depleted in 2007 than it was in 1991.

Temporal plots for the deuterium data show that the Lower Fountain spring became slightly more enriched by approximately 4.8 during the 10 year period that data is available

for the spring (‰SMOW) and the Upper Fountain spring experienced an increase in deuterium from -27.3 (1991) to -20.81(2007).

Increasing evaporation (see Section 2.3.1) may be the cause of the decrease in the depletion of the oxygen 18 and deuterium concentrations in the Upper and Lower Fountain springs between 1997 and 2007. In order to assess the relationship between the decrease in depletion and evaporation, the evaporation data for the Pretoria area was illustrated in Figure 26. This graph shows that the evaporation in the Pretoria area fluctuates seasonally but remains the same on average from the beginning of 1997 until the end of 2007. This gives an indication that the groundwater emerging from the springs is recharge in an area outside the City of Tshwane.

The time series graph for the oxygen 18 (Figure 27) shows that there is a minor decrease in depletion (approximately 0.4). This increase in ^{18}O concentrations is much smaller than that of the Upper and lower Fountain which become less depleted by approximately 1.15 and 1 respectively. Figure 22 is a temporal illustration of the deuterium data for the rainfall stations in Pretoria. This graph shows that, overall, the rainfall has become more depleted with respect to deuterium from the beginning to the end of the observation period. The most likely reason for the differences between the oxygen 18 and the deuterium in the springs and the rainfall is the recharge area of the groundwater emerging from the springs potentially being outside of Pretoria or at least outside of the relevant compartments.

Figure 29 is a scatter plot of the relationship between $\delta^{18}\text{O}$ and $\delta^2\text{H}$ at the Upper and Lower Fountain springs. This graph is utilised in order to fingerprint the water emerging at the springs by comparing the isotopic ratios at each spring. All the data points for each spring plot on or very close to the GMWL and plot slightly below the LMWL. Since both springs have an isotopic signature which is very close to that of rainfall the karst aquifer has probably recently undergone recharge from precipitation. The similar isotopic signature of both springs may indicate a shared origin of the recharge in both the Fountains East and Fountains West compartments. Additionally, the position at which the springs plot (just below the LMWL for Pretoria) may suggest that the recharge area of the compartments is not within the Pretoria area.

Both the springs are depleted in deuterium and oxygen 18 with values varying between -16 and -27, and -5 and -3 respectively. This indicates that a large amount of fractionation has occurred prior to recharge. One process which could lead to the isotopically depleted nature of the water observed at both springs is the rainout effect which was discussed in Section 2.3.1. The depleted nature of the isotopes observed at the springs may also be a result of recharge characteristics of the soil in the area of recharge. For example, the Pretoria Group sediments located to the north of the springs weather to coarse grained soils through which rain would infiltrate easily while the Hekpoort andesites which are situated further north of the springs weathers to a more clayey soil which may lead to ponding of the rainfall. The rain that infiltrates into the more coarse soil would therefore be depleted when compared to the rain which has ponded on the clayey soil and been exposed

to evaporation and therefore enrichment. Groundwater recharge into the dolomite aquifer from the area to the north of the springs (Pretoria Group) may be possible as these areas have a higher elevation than the dolomites. Precipitation may therefore move from the north into the dolomite areas through surface flow or near surface flow.

6.5. Local Geology and Conceptual Model

The simplified local geology of the site was inferred from the simplified geological logs which were statistically analysed in Table 9 and the 1:250 000 geological map series (2528 Pretoria). A soil layer with an average thickness of approximately 5m was assumed at the springs. The soil layer was anticipated to be composed of a combination of wad and transported material. This layer is followed by, dolomite, chert or interbedded dolomite and chert. No lithologies below the depth of this final layer (approximately 100m) were investigated as water movement was assumed to be minimal below this depth. Running in a north south direction between the springs is the Pretoria syenite dyke and lying to the north of the springs are the rocks of the Timeball Hill formation. Borehole logs were not available during the time of this study, to confirm the thickness of the dyke in the vicinity of the springs and the dyke was therefore digitised from the geological map (Figure 10). The inaccuracy of using such a large scale map does not impact the findings in this study fundamentally as the conceptual model that has been compiled is largely for illustrative purposes.

It was also anticipated that cavities are present within the dolomite bedrock as karstification is a necessary process in order for the movement of groundwater to occur within the otherwise impermeable dolomite rock. The major water-bearing cavities were assumed to be concentrated in the vicinity of the springs as none of the boreholes within a 3.5 km radius showed any signs of air loss (no sample return) indicating no interception of underground cavities or conduits or cavities, based on limited drilling.

The 3D block conceptual model in Figure 30 is a simplified version of the reality that exists with the karst compartments and is based on simplified and sometimes incomplete data. It is a visual representation of a large portion of the data (including the geology data, DEM and water level data) that was acquired to assist in the characterisation of the springs and the compartments to which they belong.

7. Conclusions

Characterisation of the Upper and Lower Fountain springs and the compartments to which they belong was done using numerous sources and types of data. Examples of the data used include inorganic chemistry and spring discharge volumes. The deductions below are based on the analysis and interpretation of these data sets.

- A difference of 8.5 m in water levels can be seen between the adjacent compartments. This is notable and could indicate that movement of water, through the Pretoria syenite dyke, between the compartments is limited as a connection between the compartments would lead to a natural equilibration of water levels. However, the significant difference between the average water levels cannot be used alone to prove or disprove a connection between the two compartments as water levels are only available from one borehole in the Fountains West compartment. The lack of water level data could therefore be the reason for the large difference between the water levels and not the permeability of the dyke (as is assumed in most cases where dolerite or syenite dykes cross cut South African dolomite systems).
- The similar water level trends in boreholes from both compartments may indicate that the relationships between the water levels in the compartments are linked and that movement of water does occur between the two compartments. Another conclusion that can be made from the water level data is that irrespective of the compartment, groundwater flow is generally from the south to the north in the karst aquifer.
- With regards to the discharge volumes it was determined that the Lower Fountain spring has an average discharge volume that is approximately 3.53 Ml/day higher and more constant than the Upper Fountain spring which is found to have more erratic discharge volumes after June. The erratic nature of the discharge volumes for the Lower Fountain spring is thought to be a result of abstraction occurring in the vicinity of the spring within the Fountains West compartment. However, this could not be established as no abstraction boreholes in the area could be found from either the NGA or the DWA databases.
- The Upper and Lower Fountain springs can be characterised as fresh recently recharged groundwater (with a groundwater signature of $\text{Ca}(\text{Mg})\text{-HCO}_3$). The major cations of both springs are found to be Ca and Mg and the major anion is found to be HCO_3 . The similar Ca and Mg concentrations observed on either side of the dyke are to be expected as these ions are present as a result of dolomite dissolution. The Lower Fountain spring is also characterised by a higher average TDS than the Upper Fountain spring. The trends followed by some major ions vary significantly while others follow similar trends between the springs.
- From the isotope data it can be established that both the Upper and Lower Fountain spring are characterised by highly depleted isotope concentrations that indicate a recently recharged nature of the groundwater. The most likely source of the water emanating from both springs is rainfall which has undergone abundant fractionation and recharged the karst aquifer in a zone that lies outside of the city of Tshwane. The similar isotope values observed at both springs indicate that the recharge zone from the groundwater arising from the springs is the same.
- Stable isotope data and inorganic chemistry data were used to aid in the assessment of the connection between the compartments. The water level data was not used for

this purpose as only one borehole was found within the Fountains West compartment and the spring discharge volumes were also omitted from the assessment as no abstraction boreholes could be found in the Fountains West compartment and therefore the reason for the erratic discharge volumes at the Lower Fountain spring could not be established. Both the stable isotope and inorganic chemistry data indicate that movement of water between the two compartments does occur as they share similar groundwater chemistry's and a highly depleted isotope signature.

- The implication of the connection is that dissolved constituents may move from one compartment to another and if a contamination does occur in either of the compartments that contamination will affect the groundwater in the other compartment. Extra precautions should therefore be taken in order to protect this sensitive karst system and the groundwater emanating from the springs. It is however necessary to confirm this finding by attaining complete water level datasets for both compartments and assessing the reason for the erratic discharge volumes for the Lower Fountain spring.

8. Way Forward

In order to fully and accurately characterise the springs and the Fountains East and Fountains West compartments, the following is recommended:

- The reason for the erratic discharge volumes at the Lower Fountain spring should be investigated further as significant fluctuations in the discharge volumes may prevent the spring being available as a reliable source of water in the future.
- Boreholes should be drilled into the Pretoria dyke in order to assess the degree of fracturing. This will aid in assessing the movement of the water between the Fountains East and Fountains West compartments.
- Isotope data should be utilised to establish the recharge area for the water in the karst aquifer. Once the source area is established it will be possible to ascertain whether the decreasing depletion of the heavy stable isotopes in the springs are a result of increasing evaporation.
- All data used during this project (including the isotope and chemistry data) is available to the scientific fraternity and can be used to address temporal trends in the future regarding the water supply to the City of Tshwane.

9. References

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APPENDIX A: DWA water levels

APPENDIX B: Spring discharge volumes

APPENDIX C: Inorganic chemistry

APPENDIX D: Stable isotopes (springs)

APPENDIX E: Stable isotopes (rainfall)

APPENDIX F: NGA boreholes