QUANTIFYING THE ROLE OF GROUNDWATER IN SUSTAINING GROENVLEI, A SHALLOW LAKE IN THE SOUTHERN CAPE REGION OF SOUTH AFRICA

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Promoter:

Prof. PD Vermeulen

Oblique aerial photograph of Groenvlei, looking eastwards from Sedgefield
(Photo: D. Phelp).

In memory

Professor Gerrit van Tonder

a friend who taught many much about groundwater

DECLARATION

I declare this PhD dissertation submitted to the University of the Free State is my own work. It has not been previously submitted, or submitted to a university other than the University of the Free State.

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SUMMARY

Eight of the 21 Ramsar-designated wetlands in South Africa are located in similar geohydrological settings as Groenvlei, a 359 ha lacustrine wetland found east of Sedgefield in the southern Cape. Groenvlei is unique as it is isolated from the sea and neither fed nor drained by rivers. Consequently, the lake is fed only by rainfall and groundwater inflow. Losses comprise evaporation and groundwater outflow. These characteristics result in a relatively uncomplicated hydrological system that allows for the geohydrological component to be quantified and understood. Using climatic and lake data monitored by the Department of Water Affairs and geohydrological data collected over a period of a decade, research was conducted to quantify the groundwater contribution to the system and develop an improved understanding of the hydrology of Groenvlei.

A daily water balance based on rainfall, adjusted S pan evaporation data and lake levels was used to compute that the nett groundwater contribution to Groenvlei amounted to about 0.3 mm/d. It was shown that S pan evaporation data adjusted by coefficients prescribed by Midgley et al. (1994) should be used to quantify lake evaporation, and that the reed collar transpired 10% to 30% more during summer than evaporated from open water. No water is transpired by the reed collar in winter as the vegetation is dormant.

Integrating the water balance results with steady-state Darcian flow calculations and a chemical mass balance indicated direct rainfall (71.6%) and groundwater inflow along the western and northern boundaries of the lake (28.4%) constituted inflow into the system. This is balanced by evaporation from open water (61.7%), transpiration from the reed collar in summer (21.4%) and groundwater outflow along the southern boundary (16.9%). This latter component invalidates claims that Groenvlei is endorheic in character.

Recharge to the Eden Primary aquifer was estimated to be in the order of 20% MAP. It was calculated only 5.7% of rainfall in the lake catchment discharges into the lake. The balance of rain entering the subsurface is lost through terrestrial evaporation or discharges into the sea via the deeper part of the aquifer. It was interpreted that the deceptively thick vadose zone plays a buffering role in the hydrology of the area and that evapotranspiration losses are appreciable.

The importance of the reed collar was further exemplified by the retention of salts in the vegetative fringe. Salts are assimilated by the vegetation and retained in the hyporheic zone until re-entrained into the main water body through wind and wave action. This results in only part of the salt load leaving the lake along the southern boundary and affecting groundwater quality between the lake and the sea. Further research is required to confirm this.

The results of the research allowed for tools to be developed to assess the impact of groundwater abstraction from the lake's catchment on lake levels and water quality. These tools could also be used to demonstrate Groenvlei has long since lost its connection to the marine or estuarine environments, with a new equilibrium being reached within 120 years of disconnect. The young lake is dynamic in character and rapidly responds to hydrological change. In its short history, Groenvlei has adapted and responded to changes in both sea level and climate, collectively resulting in the present-day system.

In addition to highlighting the importance of sound conceptualisation, data quality and a convergence of evidence, the outcomes of this study challenged the findings of Roets' (2008) PhD research and found no scientific evidence to support his contention that Groenvlei is sustained by underlying Table Mountain Group aquifers. It was also found that the permeability south of Groenvlei is not low and the extent of the lake catchment is 25 km^2. Past research of Groenvlei has resulted in a number of misconceptions and it was argued a need exists to link hydrologists and ecologists to better understand wetlands, with each contributing specific skills and knowledge.

An important contribution of the research documented in this thesis is that the approach used can be applied to similar wetlands where the role of groundwater might be less obvious because of river flows and tidal exchange. The importance of sound conceptualization and direct rainfall onto wetlands, quantification of evaporative losses using S pan data and coefficients prescribed by Midgley et al. (1994), and the relationship between open water losses and transpiration losses are three aspects that could improve the understanding and quantification of lake – groundwater interaction elsewhere.

A limitation to understanding the geohydrology of Groenvlei is the lack of information pertaining to aquifer thickness. It is therefore recommended four boreholes be drilled to either bedrock or at least 100 m in depth (whichever is reached first) to quantify the thickness of the aquifer. Other limitations that require attention include:

- Uneven spatial distribution of the geohydrological data;
- Lack of information of losses from open water and the reed collar; and
- Absence of monitored groundwater data needed to address temporal relationships between groundwater and the lake.

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ABBREVIATIONS AND NOTATIONS

b	aquifer thickness (m)
CCPP	calcium carbonate precipitation potential
CMB	chloride mass balance
conc.	concentration
CRD	cumulative rainfall departure
δ	delta
D	deuterium – isotope of hydrogen
DWA	Department of Water Affairs
DEAT	Department of Environmental Affairs and Tourism
DEM	digital elevation model
DTM	digital terrain model
EC	electrical conductivity (mS/m)
Fe	Iron
Fm	Formation
g	gram
GIS	geographical information system
GPS	global positioning system
Grp	Group
i	hydraulic gradient
K	hydraulic conductivity (m/d)
ka	thousand years
kg	kilogram
km	kilometres
L/s	litres per second
m	metres
Ma	million years
MAE	mean annual evaporation
MAP	mean annual precipitation
m/d	metres per day
m^3/d	cubic metres per day
mg/L	milligrams per litre

mS/m	millisiemens per metre
NGA	National Groundwater Archive
NGDB	National Groundwater Database
^{18}O	isotope of oxygen
‰	permil (parts per thousand)
Q	yield (usually in L/s or m^3/d)
^{222}Rn	isotope of Radon
S	storativity or specific yield
^{86}Sr	isotope of Strontium
^{87}Sr	isotope of Strontium
T	transmissivity (m^2/d)
t	time (usually in days)
TMG	Table Mountain Group
w	width (m)

DEFINITIONS

Baseflow: sustained low flow in a river during dry or fair weather conditions, but not necessarily all contributed by groundwater; includes contributions from delayed interflow and groundwater discharge.

Biome: a major regional group of distinctive plant and animal communities best adapted to the region's natural physical environment, latitude, elevation, and terrain.

Colmation: the retention process where the clogging of the top layer of channel or lake sediments by fine material settling out of the water column or filtered out by passage of infiltrating or percolating water causes a reduction of porosity and permeability of the porous media.

Ecosystem: an organic community of plants and animals and the physical environment they inhabit.

Endorheic: a blind or closed drainage system without any drainage outlet i.e. to flow within.

Eutrophic: water rich in minerals and organic nutrients.

Evaporation: the process by which a liquid or solid is changed into a gas.

Evapotranspiration: a general term that includes the process by which water is transferred from land to the atmosphere by evaporation from soil and other surfaces and by transpiration from plants, often used synonymously with evaporation.

Geohydrology: the study of the properties, circulation and distribution of groundwater; used interchangeably with hydrogeology.

Groundwater: water found in the subsurface in the saturated zone below the water table or piezometric surface i.e. the water table marks the upper surface of a groundwater system.

Hyporheic zone: water-saturated transitional zone between surface water and groundwater.

Lacustrine wetland: means "of a lake" and includes wetlands situated in a topographic depression or a dammed river channel. Lacustrine wetlands have a total area greater than 8 ha and the surface area covered by vegetation (mosses, lichens, trees, shrubs or persistent emergents) is less than 30% of the total.

Macrophyte: an aquatic plant that grows in or near water and is either emergent, submergent, or floating.

Palustrine wetland: includes any inland wetland which lacks flowing water, contains ocean-derived salts in concentrations of less than 0.05%, is non-tidal and less than 8 ha in extent.

Recharge: the addition of water to the saturated zone by the downward percolation of precipitation.

Transpiration: the process by which water vapour escapes from living plants - principally the leaves – and enters the atmosphere.

Vegetation: the group of plants forming the plant cover of a geographic area.

Vlei: Afrikaans word meaning small or shallow body of water.

ACKNOWLEDGEMENTS

I have benefited greatly from interaction with people from a range of disciplines and walks of life. I will never be able to name them all, but that does not stop me recognising their role in shaping me. It never ceases to amaze me how much people know; and don't know.

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The many cups of coffee and moral support provided by Piet Erasmus when things were not going according to plan provided the early spark to learn more about the geohydrology of Sedgefield and the surrounding lakes. Also, much of my interest and understanding of surface – groundwater interaction developed while trying to get to grips with the groundwater component of the Reserve. Dr Heather MacKay (formerly of the Department of Water Affairs) and Dr Kevin Pietersen (formerly of the Water Research Commission) provided opportunities to meet a host of interesting people thinking about similar issues, but from other angles. Publication of the *GRDM Manual* and *Surface water–groundwater interaction in a Southern African context* **-** *a geohydrological perspective* are outcomes of this support. Dr Bill Harding fueled my interest in groundwater – lake interaction when assisting him with the restoration of Zeekoevlei on the Cape Flats.

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Meeting Prof. Danie Vermeulen through the Flemish-sponsored FETWater initiative led to a friendship and his prodding to get this thesis done. A chance encounter with the late Prof. Gerrit van Tonder on a farm near Graaff-Reinet in 1984 provided a platform for me to regularly tap into his immense groundwater knowledge over a period of 30 years. Their support of this research effort is gratefully acknowledged.

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On a personal note, parents John and Ina gave me a sound platform from which to grow. Long gone, but never forgotten. My wife Jenny has always been my support and encouragement; and somehow makes sure the ball keeps rolling forward – including this thesis. And of course, daughters Michelle and Samantha, who make it all worthwhile and provided much inspiration. Thank you for letting me put pen to paper.

Finally, I hope this study contributes to our understanding of groundwater's role in sustaining lakes and wetlands, and brings into perspective interaction between man and the environment. In an era obsessed with predicting climate change, we seem to have forgotten to support conceptualisation with measurement. We really do have to look back to go forward. To those that read this thesis, I hope you can take "a lot" from it.

1 INTRODUCTION

1.1 Problem Statement

Following promulgation of the National Water Act (Act 36 of 1998), water resources in South Africa now need to be managed holistically. This results in a need for an improved understanding of surface - groundwater interaction. Recent work showed this understanding is poor (Parsons, 2004a; Hughes et al., 2007; van Tonder and Hughes, 2008, Humphries et al., 2011; Tanner, 2013). This also extends to wetland – groundwater interaction, where failure to appreciate the linkage between the two systems can lead to poor decision-making and management. While most of South Africa is underlain by fractured rock aquifers, a number of internationally important coastal wetland systems are located on primary aquifer systems (Figure 1.1). These include Lake Sibaya, Lake St Lucia, Lake Mzingazi and Kosi Bay in KwaZulu Natal; Groenvlei, Swartvlei, Rondevlei, Langvlei and Voëlvlei in the Southern Cape; De Hoopvlei near the De Hoop Nature Reserve; Zeekoevlei, Rondevlei, Princessvlei and Rietvlei on the Cape Flats, and Langebaan and Verlorenvlei along the West Coast. Of the 21 sites in South Africa designated by the Ramsar Convention as wetlands of international importance, eight are located in geohydrological settings similar to those of Groenvlei. Groundwater plays a leading role in sustaining at least four of these. At those wetlands dominated by tidal exchange and / or river flow, groundwater may play a lesser role, but could be critical under certain conditions – as was the case at Lake St Lucia during the extended drought of the early 2000s (Taylor, 2006; Parsons, 2009a). To manage these systems, it is important to understand what drives them and to understand what impacts surrounding activities could have on them.

Groenvlei, located directly east of Sedgefield along the southern Cape coast of South Africa (Figure 1.2), provided a unique opportunity to develop a better understanding of the role of groundwater in sustaining these water bodies. Absence of surface water inflows and the availability of monitored rainfall, evaporation and water level data suggested a water balance approach would allow for quantification of the groundwater discharge into and out of the system. By developing a sound understanding of the hydrological functioning of Groenvlei, a better understanding could be extrapolated to valued coastal wetland systems elsewhere in the country.

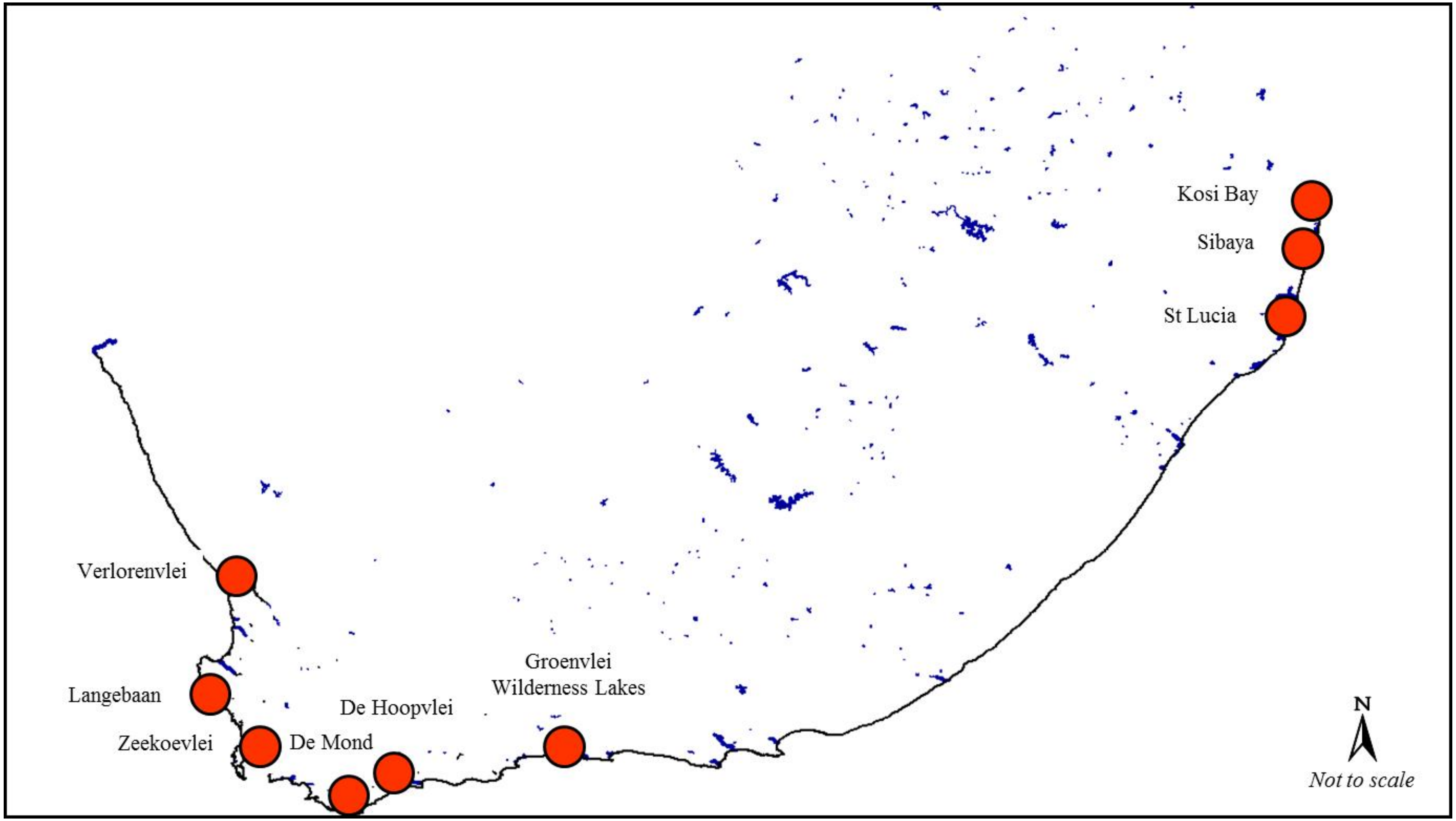

Figure 1.1: Ecologically important wetlands in South Africa located on coastal primary aquifer systems.

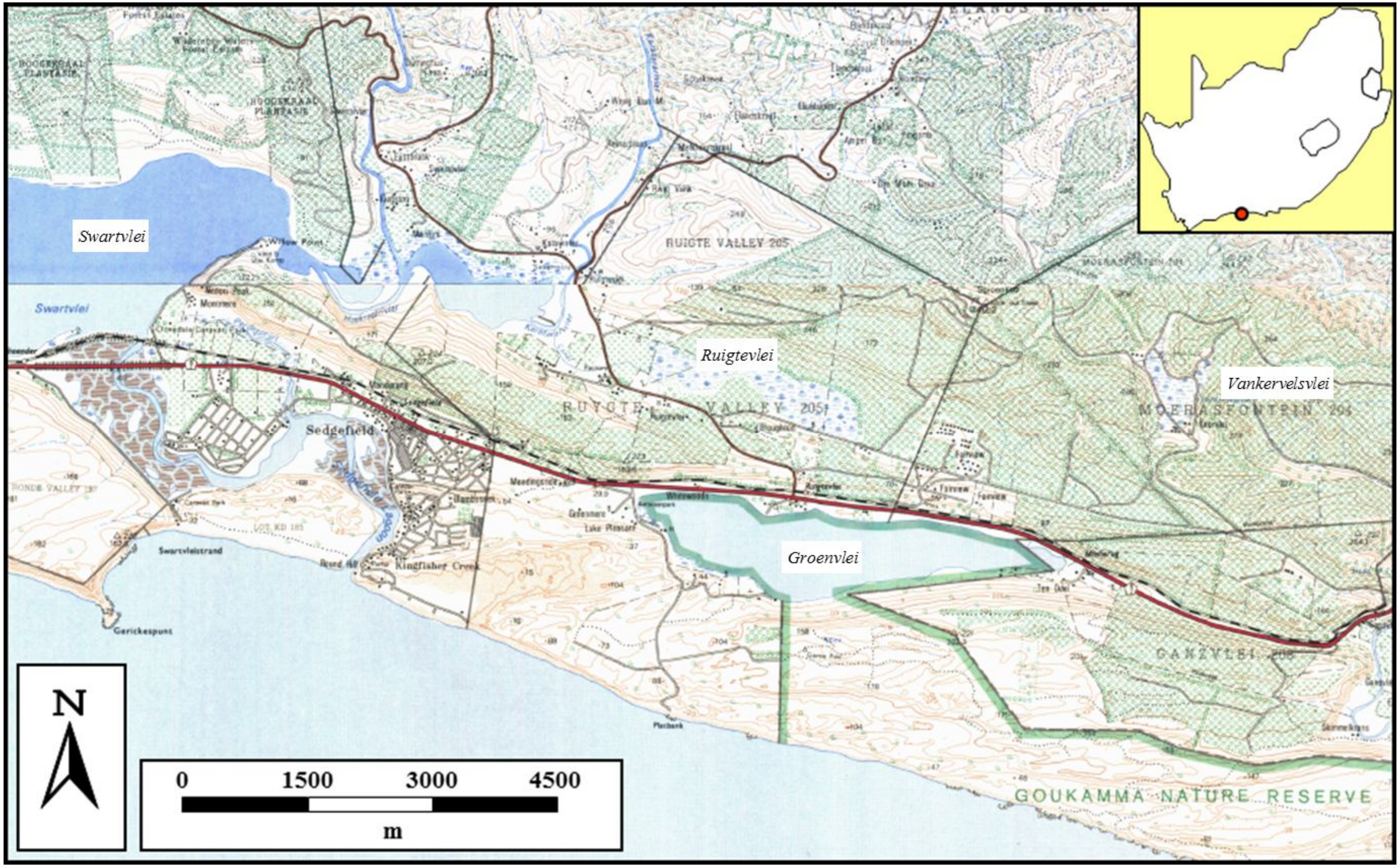

Figure 1.2: Locality map of Groenvlei.

Groenvlei is an isolated, brackish coastal lake known for its diverse bird life and one of the best venues for large mouth black bass angling. It is one of a series of coastal lakes in the area – the so-called Wilderness Lakes [1] – but the only one disconnected from the sea. It is located 5 km from the holiday town of Sedgefield. Urban development in and around Sedgefield potentially threatens Groenvlei. Abstraction of groundwater for water supply purposes could impact the volume of groundwater discharged into the lake, while sewage disposal could compromise water quality. Before this research, knowledge of the hydrological function was based on the inappropriate use of a rainfall – runoff model (Fijen, 1995) and incorrect conceptualisation (Roets, 2008). This made it difficult to identify and implement appropriate protection measures, especially when a balance was sought between protection, use of water resources and sustainable economic development. A need existed to develop an understanding of the hydrology of Groenvlei based on a sound conceptual model supported by available information.

1.2 Objectives of Research

The primary objective of the research was to develop an understanding of the hydrological functioning of Groenvlei. Specific goals of the research were to:

- Quantify the groundwater contribution to Groenvlei; and
- Understand the chemistry of Groenvlei.

A secondary objective of the study was to contribute to a better understanding of surface – groundwater interaction in general, and provide a basis for assessing the role groundwater plays in sustaining the important coastal wetland systems identified in Section 1.1.

[1] Not all authors include Groenvlei in the Wilderness Lakes, e.g. Groenvlei is not included in the Ramsar delineation by Randall and Russell (1995). Because it is isolated from the other lakes and hydrologically distinctly different from them, it is probably not correct to include Groenvlei in this group.

1.3 Setting of Hypotheses

To provide structure to the research and ensure the objectives presented in Section 1.2 could be met, a number of hypotheses were developed. These were based on the findings of previous investigations and a conceptual understanding of the hydrology of the area:

- Groenvlei is a groundwater-fed system;
- Groenvlei is not fed by surface run-off;
- There is no hydraulic link between Vankervelsvlei and Groenvlei;
- Groenvlei is a flow-through system, with groundwater discharging into the lake from the north and water discharging from the lake into the subsurface in the south;
- A pan evaporation data can be used to approximate lake evaporation; and
- The quality of water in Groenvlei is reflective of its marine origin.

1.4 Structure of Thesis

After this introductory chapter, the findings of the literature study are presented in Chapter 2. The literature study covers topics such as previous studies of Groenvlei, groundwater - lake interaction and methods for quantifying groundwater - lake interaction (water balance methods, chemical mass balance model, quantification of component parts). Chapter 3 describes data collection and methods used in the research, including the collection of existing data, patching of rainfall and evaporation data, assessment of runoff and determination of hydraulic properties and hydraulic gradients. Chapters 4 and 5 provide a description of the study area, including its climate, physiography, drainage and geohydrological properties. The groundwater contribution to Groenvlei is quantified in Chapter 6, as is the salt balance of the system. The results of the research are discussed in Chapter 7, with the discussion also addressing the hypotheses presented in Section 1.3. Key lessons learnt during this research formed the basis of the guidelines for studying similar lakes and wetlands presented in Chapter 8. The main findings of the research and recommendations are summarised in Chapter 9. A paper reviewing the role of Table Mountain Group (TMG) aquifers in sustaining Groenvlei is presented in Appendix A while data used in this research is included in Appendix B (electronic).

2 LITERATURE STUDY

2.1 Previous Studies of Groenvlei and Surrounds

Groenvlei has not been the subject of many detailed hydrological studies in the past. Martin (1956, 1959, 1960a, 1960b, 1962) provided an early account of the origin and ecology of the lake, while geological aspects were described by Toerien (1979), Illenberger (1996), Marker and Holmes (2005), and Illenberger and Burkinshaw (2008). In addressing the management requirements of Groenvlei, Fijen (1995) attempted to quantify the hydrology. Paridaens and Vandenbroucke (2007), Roets (2008), Vivier (2009) and Dennis (2010) considered geohydrological aspects.

Between 1992 and the present, the author undertook a range of geohydrological projects related to water supply, monitoring, groundwater contamination and environmental assessments (see reference list). Data and information gathered during these projects were used during this study, and are referenced where appropriate.

Information about Groenvlei and its surrounds was sourced from a wide variety of literature, including Coetzee (1980), Watson and Lang (2003), Duncan (2006) and Kirsten (2008). Of direct interest, van der Merwe (1976), Coetzee (1980) and Allanson (2006) presented descriptions of the morphology and freshwater ecology of Groenvlei. Russell et al. (2012) provided a list of documents that contribute to the state of knowledge of the Garden Route National Park and Wilderness Coastal Area.

2.1.1 Martin (1956, 1959, 1960a, 1960b, 1962)

Martin (1956, 1959, 1960a, 1960b, 1962) prepared a series of documents that described the geology and vegetation of Groenvlei. He referred to the lake as a fen – peat-forming groundwater-fed mineral-rich wetland neutral or alkaline in character. Early sediments indicate a freshwater system that was then covered by marine mud. Also, Martin (1959) demarcated the position of a buried channel connecting Groenvlei to Swartvlei.

Martin (1960a) speculated as to the origin of the salinity of Groenvlei, and provided the following possibilities:

- The salt is residual from the time when sea water entered Groenvlei;
- The salt is derived from spray blown over the dunes;
- The lake is fed by saline springs;
- The salt is the concentration residue left by evaporation.

He was not able to decide which was the most likely possibility, but favoured the last.

2.1.2 Fijen (1995)

Fijen (1995) provided a holistic description of Groenvlei, including its climate, drainage and hydrology. He highlighted the management needs of the water body. A weakness of this study was the use of the Pitman model to quantify the hydrology of the area. The model, described by Pitman (1973), Hughes (2013), Tanner (2013) and others, is a rainfall – runoff model and is inappropriate to use because of the absence of any surface water features that drain into Groenvlei. Fijen (1995) acknowledged this by exploring the role of groundwater in sustaining the wetland. Based on a 10 year water balance, and using annual data, he estimated inflows to the lake amounted to 0.91 Mm^3/a (run-off and groundwater) with minimal seepage losses. He postulated small seepage losses could be expected because the area between Groenvlei and the sea consists of cemented dune rock or aeolianite of low permeability.

Fijen (1995) calculated inflows and outflows using a salt water balance. He proposed the salinity was 35 000 mg/L when the lake formed 5 000 to 7 000 years ago, and current salinity levels reflected a balance between inflows and outflows. Based on the saltwater balance calculations, he set inflow at 1.8 Mm^3/a and seepage losses at 0.1 Mm^3/a.

2.1.3 Illenberger (1996)

Illenberger (1996) and Illenberger and Burkinshaw (2008) noted dunes form where there is an adequate sand supply and sufficient wind energy to move the sand, and offered that the Wilderness dune cordon is of worldwide importance. Good examples of fossil coastal dunes are found at Platbank, directly east of the Swartvlei mouth, where dune sand was cemented by calcium carbonate to form aeolianite. The Wilderness dune cordon comprises steep-sided ridges separated by coastal lakes. Three major dune cordons are developed on land (seaward cordon, middle cordon, landward cordon), while submerged cordons occur offshore down to depths of 50 m (Figure 2.1). The morphology of the landscape is the result of many changes of sea level since the onset of the Quaternary period, with numerous phases of dune building and erosion. The seaward dune cordon comprises stabilised dunes (6 ka) that overlie fossilized dunes (115 ka). The middle cordon was dated at 200 ka while the landward cordon comprises dune sediments in excess of 600 ka in age. Marker and Holmes (2005) observed these ages were based on rates of sand accretion and could not be regarded as absolute ages. Using luminescence dating techniques, they set the age of the basal aeolianite at 128 ka and measured 90 ka towards the top of the geological unit. Later work by Bateman et al. (2011) essentially confirmed and refined these ages.

2.1.4 Paridaens and Vandenbroucke (2007)

This joint MSc. thesis provided a description of the geohydrology of Groenvlei and included field measurements of hydraulic conductivity (K) using a mini disk infiltrometer, slug tests and a seepage meter. Using a digital terrain model, the catchment area of Groenvlei was set at 14.9 km^2. Groundwater inflow and outflow was determined using a long-term mass balance approach (i.e. included both depth and water quality terms) to be 1.21 mm/d and 0.23 mm/d. This equates to 1.60 Mm^3/a and 0.28 Mm^3/a when the area of the reed collar is included in the total area of the lake.

Paridaens and Vandenbroucke (2007) measured the hydraulic conductivity of soil and sand at Groenvlei using mini disk infiltrometers, slug tests and seepage meters (Table 2.1). The mini disk infiltrometers measured vertical K at surface, while the slug tests measured K at a depth of about 6 m. Their seepage meter measurements must be treated with some caution as cognisance was not taken of (a) whether the measurements were taken at points of inflow or outflow and (b) the changes of hydraulic gradient around the lake.

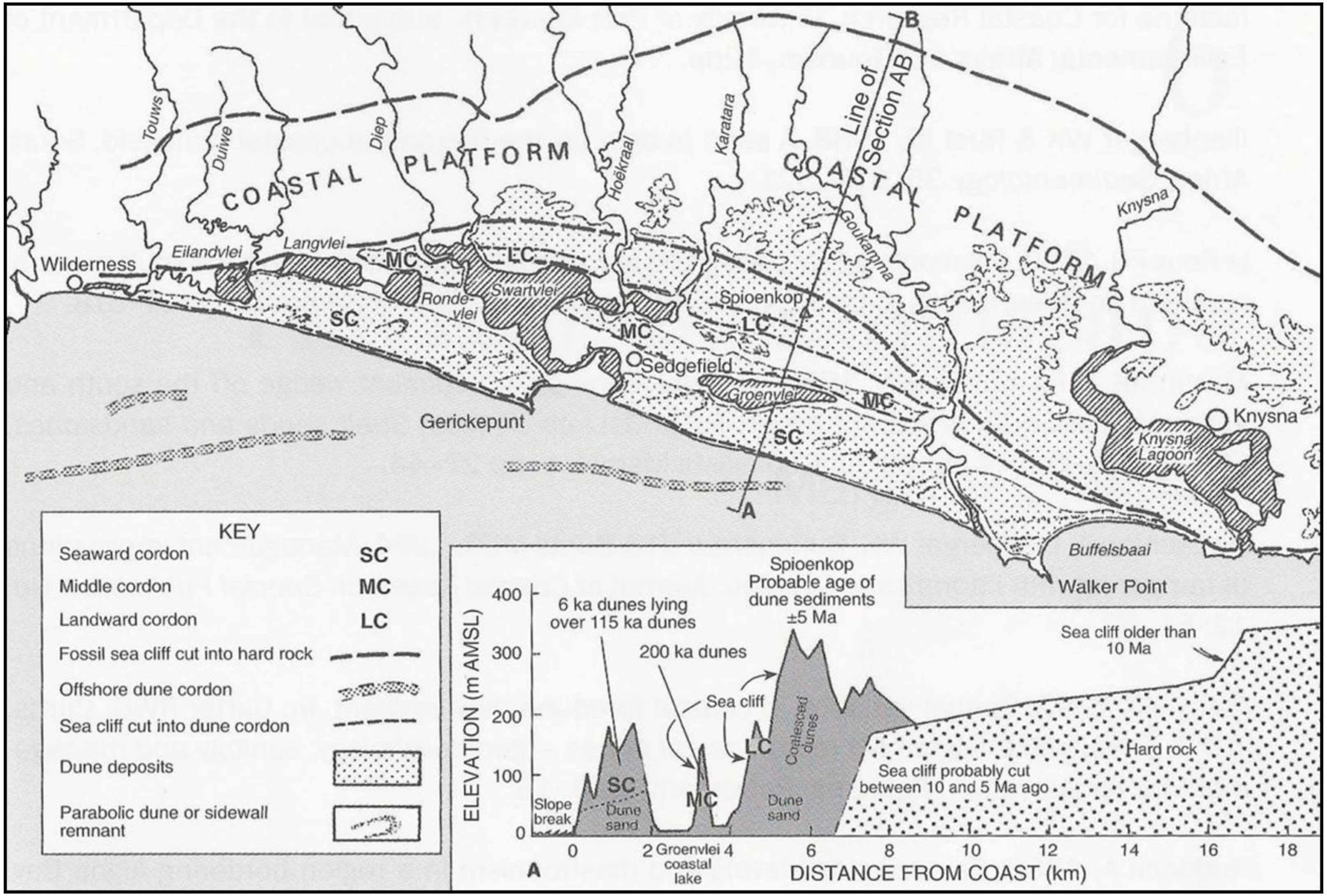

Figure 2.1: The extent and age of dune cordons in the vicinity of Groenvlei (Illenberger and Burkinshaw, 2008).

Table 2.1: Summary of hydraulic conductivity measurements by Paridaens and Vandenbroucke (2007)

Method	**K Ave (m/d)**	**K Min. (m/d)**	**K Max. (m/d)**	**n**
Mini infiltrometer	2.5	1.33	4.85	7
Slug & bail	9.9	2.14	21.6	5

Applying a method described by Arriaga and Leap (2006), Paridaens and Vandenbroucke (2007) used temperature – depth profiles along the edges of Groenvlei to determine whether groundwater was flowing into or out of the lake. Profiles were taken 500 m apart, with two profiles taken at each site. In essence, their results showed groundwater flowed into the lake along the northern shores and out of the lake along its southern shores. However 5 of their 22 accepted profiles points were anomalous to this pattern. It could be interpreted from the data that the inflow boundary was longer than the outflow boundary, with the inflow boundary being 8.24 km long and the outflow boundary 3.28 km in length.

2.1.5 Roets (2008)

Roets (2008) – together with Roets et al. (2008a, 2008b) and Roets (2009) – considered the role of discharge from the underlying TMG Aquifer in sustaining Vankervelsvlei [1] and Groenvlei (Figure 1.2). They contended groundwater from the TMG Aquifer discharged into the unconfined primary aquifer system and into Vankervelsvlei. Water then discharged back into the primary aquifer and discharged into Groenvlei. These studies were critically reviewed by Parsons (2009b) [2] and found to be without merit as the conceptual model was flawed. They did not take account of measured groundwater level data from seven boreholes drilled on dune ridges in the area. No scientifically credible data or information was presented to support their hypotheses and they failed to consider sources of water other than that of discharge from the TMG Aquifer. Specific problems included:

[1] Vankervelsvlei is a unique "floating bog" wetland located 2.5 km northeast of Groenvlei and described by Irving and Meadows (1997).

[2] This review is presented as Appendix A of this thesis.

- Their hypothetical cross section misrepresented the topography of the area and incorrectly portrayed the elevation of Groenvlei in relation to both the sea and Vankervelsvlei;
- The link between Vankervelsvlei and Groenvlei was dependent on a hydraulic gradient of 0.054. Groundwater level measurements in the area and a review of the international literature indicated this not to be possible;
- The conceptual geohydrological model presented by Roets (2008) did not conform to the geological map of the area (Coetzee, 1979), and the confining layer required to lift groundwater some 150m does not exist. The water levels measured at Vankervelsvlei are textbook examples of perched levels;
- The electrical conductivity (EC) range used by Roets (2008) to motivate a TMG Aquifer origin applied to 56% of groundwater sampled in the Western Cape province – irrespective of aquifer – and hence had no value as a "fingerprint";
- The iron (Fe) concentration of 382 mg/L recorded in a shallow wellpoint at Vankervelsvlei was either a laboratory error, a typing error or possibly related to the vegetative mat of the vlei, but cannot be attributed to discharge from the TMG Aquifer;

Consequently, the geohydrological findings of this research are rejected, and not considered to contribute to knowledge about Groenvlei.

2.1.6 Vivier (2009)

Vivier (2009) undertook a Reserve determination study of the Outeniqua catchment between Gouritzmond and Plettenberg Bay. The study included compiling a finite element model of the Groenvlei subcatchment using the FEFLOW modelling package. The modelling process had a major weakness with Vivier (2009) reporting 13 of the 68 groundwater levels available to him from the National Groundwater Archive (NGA) were below sea level. As most of these groundwater levels were measured by the author and supplied to the Department of Water Affairs (DWA), this is simply not true. The incorrect groundwater levels used in the model were probably the result of database or computation errors.

Vivier (2009, pg 42) reported “recharge in the Groenvlei catchment is almost 100% as the sands are highly permeable with a hydraulic conductivity in the order of 4 m/d which represents a transmissivity of 120-150 m^2/d”. He used a recharge estimate of 45% of mean annual precipitation (MAP) in his model. Using parameters documented in Table 2.2, he computed a groundwater flow balance for the Groenvlei catchment. Input included recharge from rainfall on the wetland (5 770 m^3/d), surface water run-off (1 293 m^3/d) and groundwater inflow from the upstream catchment (2 653 m^3/d). Losses included evaporation (9 321 m^3/d), outflow to the ocean (900 m^3/d) and borehole abstraction (500 m^3/d).

Table 2.2: Parameters used by Vivier (2009) in the modelling of the Groenvlei subcatchment.

Component	Unit	Quantity
Groenvlei catchment area	km^2	10.53
Upstream catchment	km^2	3.31
Groenvlei surface area	km^2	3.24
Depth	m	3
MAP	mm/a	650
MAE	mm/a	1400
Recharge	% MAP	45
Steady-state outflow to ocean	m^3/d	32 000
Aquifer hydraulic conductivity	m/d	3.5
Aquifer thickness	m	100
Initial TDS	mg/l	3500

Using a drawdown limitation of 0.5 m and restricting all groundwater abstraction within 200 m of Groenvlei , the model was used to compute that 400 m^3/d could be abstracted from a zone extending from 200 m to 1 km from the wetland without causing significant impact. Abstraction of 1 500 m^3/d from the entire subcatchment would not have a detrimental impact on the wetland if the buffer zones were maintained.

2.1.7 Dennis (2010)

A numerical model (MODFLOW) was compiled to predict aquifer response to artificially recharging treated wastewater into the subsurface at the Sedgefield wastewater treatment works (WWTW) positioned 800 m west of Groenvlei. The initial transmissivity and specific yield values applied in the model were 300 m^2/d and 0.2 respectively. It was determined groundwater flows southward toward the sea.

2.2 Studies of Other South African Coastal Lake Systems

Hill (1974) observed there are two groups of southern African lakes in close proximity to the sea where most are connected to the sea. Both groups – the five lakes in the Wilderness and the lakes of KwaZulu Natal and southern Mozambique – are located on primary aquifer systems and each contains a lake not connected to the sea i.e. Groenvlei and Lake Sibaya.

2.2.1 Lake St Lucia

Lake St Lucia is a 350 km^2 lake of national and international importance located on the extensive Zululand Coastal Aquifer in northern KwaZulu-Natal. Because of potential impacts of both afforestation and mining (Rawlins, 1991; Kelbe and Rawlins, 1992a, 1992b; Kelbe et al., 1995), the area has been the focus of a number of geohydrological investigations. The St Lucia System is more complex than Groenvlei because:

- The open body of water is surrounded by a number of smaller wetlands;
- The system is driven by direct rainfall, variable inflow from five rivers, groundwater, exchange through the estuary mouth and evapotranspiration losses; and
- The hydrological and ecological functioning is dependent on the state of the estuary mouth (Taylor, 2006).

Knowledge about the groundwater component of the system is largely driven by the outcomes of numerical modelling (Kelbe and Rawlins, 1992a, 1992b; Kelbe *et al.*, 1995; Wejden, 2003; Voeret *et al*. 2007 and others), dependent on assumptions relating to plant rooting depths and evapotranspiration losses by different land covers. These components are difficult to quantify, as it is not possible to assess whether vegetation is dependent on water from the vadose or the saturated zone. This is compounded by some vegetation being facultative in character i.e. water use is dependent on the water available to plants (Le Maitre et al., 1999) [1]. Unfortunately the outcomes of modelling were not calibrated against measured aquifer response to afforestation and deforestation. While it is generally accepted groundwater makes a small volumetric contribution (~7%) [2] , its vital role in sustaining aquatic ecosystems was highlighted during the severe drought experienced between 2002 and 2007 (Taylor, 2006; Voeret et al., 2007; Vrdoljak et al. 2007; Parsons 2009a). Groundwater provided the only freshwater input (~ 100 mS/m) to flora and fauna escaping ever-increasing hypersaline conditions (~ 23 000 mS/m) of the shrinking surface water.

Rawlins (1991, pg. 110), in his monitoring of groundwater levels in the Zululand Coastal Aquifer adjacent to Lake St Lucia, observed intermittent extreme events are capable of redressing any imbalance caused by other factors. He also noted no recharge takes place when rainfall events do not exceed 10 mm.

2.2.2 Lake Sibaya

Lake Sibaya, also located on the Zululand Coastal Aquifer, covers an area of 65 km^2. It is some 20 m above mean sea level and 41 m deep. The area experiences a MAP of 1 300 mm/a, almost double that of Groenvlei. Pitman and Hutchison (1975) provided a water balance of the lake, but were challenged by surface flow estimations and separating the nett groundwater contribution (nett lake recharge). They found the nett groundwater contribution to vary erratically from month to month, with several negative values in their 5 year record. Average daily nett groundwater contribution equated to 1 mm/d.

A groundwater level contour map of the Zululand Coastal Aquifer in northern KwaZulu-Natal prepared by Meyer and Godfrey (1995) showed the effect of evaporation from Lake Sibaya on groundwater; and indicated the lake to be a groundwater sink. Meyer *et al*. (2001) compiled a numeric model to simulate the impact of afforestation on the lake. They concluded groundwater flow directions around the lake would change if afforestation resulted

[1] A fundamental tenet of ecology is that ecosystems generally use a resource in proportion to the availability of the resource (whether it be water, light, nitrogen or some other resource), and the availability of the resource will be a significant determinant of the structure, composition and dynamics of an ecosystem (Tilman, 1988 as quoted by Brown *et al.*, 2003). Where groundwater is accessible, ecosystems will develop some degree of dependence on it.

[2] Direct rainfall provides the biggest input into Lake St Lucia, accounting for 42% of the total input and 48% of the freshwater input.

in additional evapotranspiration losses of 1 000 mm/a or more. In turn, this would reduce groundwater inflow, increase losses and result in a lowering of the water level in Lake Sibaya. This report did not provide detail on the water balance of the system or on fluxes in and out of the lake.

Between 1967 and 2002, the water level in the lake varied between 18.2 mamsl and 20.4 mamsl (Meyer and Godfrey, 1995). Since then, the level has steadily declined to 16.0 mamsl (Weitz and Demile, 2013), presumably in response to increased water abstraction for urban and rural water supplies. Using preliminary estimates provided by Weitz and Demile (2013), the relative importance of the component parts of the water balance could be determined (Table 2.3).

Table 2.3: Preliminary water balance of Lake Sibaya

Inflow		**Outflow**	
Component	**%**	**Component**	**%**
Rainfall	34.1	Evaporation	69.5
Run-off	26.3	Groundwater	17.7
Groundwater	39.6	Abstraction – lake	1.1
		Abstraction groundwater	11.7

(after Weitz and Demile, 2013)

Lake Sibaya and Groenvlei have a similar geological history. Both are flow-through systems principally fed by rainfall and groundwater inflow. However, Lake Sibaya differs from Groenvlei in that it is significantly larger (7 750 ha versus 248 ha), at a higher elevation (20 mamsl versus 3 mamsl), significantly deeper (43 m versus 6 m), and partially fed by a short river. Additionally, its water is less saline (60 mS/m versus 350 mS/m).

2.2.3 Zeekoevlei

Zeekoevlei is located on the western edge of the Cape Flats Aquifer and covers an area of 2.6 km^2. The lake is at an elevation of 5 mamsl and has an average depth of 1.9 m. Unlike

Groenvlei, Zeekoevlei is significantly impacted by surrounding urban development and the subject of artificial water level regulation since 1948. It is also subjected to marked seasonal variations, with winter inflows from the Lotus and Little Lotus rivers reducing EC from 200 mS/m in summer to 100 mS/m in winter. Efforts to compile a water balance of Zeekoevlei by Morrison (1989) and Harding (1995, 1996) did not take groundwater into account. Parsons (2000) estimated groundwater contributions using a water balance, Darcy's Law and fluxes presented in the literature. Good agreement of results was obtained, but the flux method underestimated the groundwater contribution from the eastern shore - characterised by a hydraulic gradient significantly steeper than elsewhere. Groundwater input was set at 4 mm/d or 15% of the average annual input. However, groundwater is the only source of water in the summer months.

2.3 Studies of Similar Geohydrological Environments

While South Africa is dominated by fractured, hard rock secondary aquifers, a few significant or important coastal primary aquifers exist. These are listed below together with key references that describe conditions of each aquifer system:

- Zululand Coastal Aquifer – Kelbe and Rawlins (1992a, 1992b), Kelbe and Germishuyse (1998a, 1998b), Germishuyse (1999), Kelbe et al. (2001), Meyer et al. (2001).
- Cape Flats Aquifer – Henzen (1973), Wessels and Greef (1980), Wright and Conrad (1995).
- Atlantis Aquifer – Bredenkamp and Vandoolaeghe (1982), Fleisher (1990), Wright (1991), Fleisher and Eskes (1992).
- Langebaan Road Aquifer – Timmerman (1985a, 1985b, 1985c), Du Toit and Weaver (1995), Weaver et al. (1997), Woodford (2005).
- Sandveld Primary Aquifer – Timmerman (1986), Conrad and Münch (2004).

Information about these and similar systems elsewhere in the world assisted in conceptualising and understanding the geohydrology in the vicinity of Groenvlei.

2.4 Surface – Groundwater Interaction

2.4.1 Preamble

Promulgation of the National Water Act (Act 36 of 1998) necessitated water resources be addressed in a holistic manner, requiring a better understanding of surface – groundwater interaction. This was in line with world trends, with important treatises on the topic being presented by Lerner (1996), Gardner (1999), Winter et al. (1999) and Sophocleous (2002). Locally, attention was also given to the surface – groundwater interaction by Vegter and Pitman (1996), Parsons (2004a), DWAF (2005), Witthüser (2006), van Tonder et al. (2007), Gomo, 2011 and Moseki (2012). Interest in the topic has evolved to such an extent that the well-known Pitman rainfall-runoff model has been modified to account for the groundwater component of the hydrological cycle (Hughes, 2013; Tanner, 2013). This thesis focused on groundwater – lake interactions, a subset of the broader topic.

2.4.2 Surface – groundwater interaction

Focus on surface – groundwater interaction resulted in many conceptual understandings being challenged, and in some cases shown to be wrong. Van Tonder and Hughes (2008), for example, reported much baseflow in rivers is from interflow, and little groundwater from adjacent aquifers reaches rivers. Parsons (2004a) argued incorrect use of terminology was one of the greatest challenges to developing a better understanding of surface – groundwater interaction. For example, hydrologists understand baseflow to be the sustained low flow in a river during dry or fair weather conditions, while geohydrologists consider baseflow to be that component of flow contributed by groundwater. Like perennial rivers through to seasonal and then intermittent rivers, storm flow through to interflow and then baseflow are points on a continuum where distinction is usually difficult. It is recognised groundwater contributes to baseflow and Parsons (2004a) proposed the term "groundwater contribution to baseflow" as a means of clarifying the process and origin of the water. Recognising the complexities of surface – groundwater interaction, Moseki (2012) observed interactions are difficult to quantify without the requisite data, and the use of multiple techniques is essential for reducing uncertainty.

2.4.3 Wetlands

Notwithstanding the recognition of a unitary hydrological cycle in the National Water Act, the role of groundwater as a driver of wetlands is yet to be fully appreciated in South Africa. Groundwater-driven systems or systems partially dependent on groundwater have a higher degree of permanency and very different characteristics to wetlands driven by rainfall or surface water inflows, thus requiring different management and protection strategies. For example, the control of groundwater abstraction could be important in the protection of the former, but is not required in the case of the latter. Wetland classifications and management guidelines by McCarthy and Hancox (2000), Jones and Day (2003), Dickens et al. (2004) and DWAF (2007) all fail to explicitly address the geohydrological component of wetland systems. It is noted a shift to a geomorphic approach [1] proposed by Ollis et al. (2009) allows for greater recognition of groundwater input into wetland systems.

Wetlands are defined by the Ramsar Convention as “areas of marsh, fen, peatland or water; whether natural or artificial, permanent or temporary, with water that is static or flowing, fresh, brackish or salt, including areas of marine water the depth of which at low tide does not exceed six metres.” This broad definition includes lakes and rivers, swamps and marshes, wet grasslands and peatlands, oases, estuaries, deltas and tidal flats, near-shore marine areas, mangroves and coral reefs, and human-made sites such as fish ponds, rice paddies, reservoirs, and salt pans. In a South African context, the National Water Act (Act 36 of 1998) defines wetlands as “land which is transitional between terrestrial and aquatic systems where the water table is usually at or near the surface, or the land is periodically covered with shallow water, and which land in normal circumstances supports or would support vegetation typically adapted to life in saturated soil.”

Wetlands are features of most landscape types in South Africa, and provide important goods and services to the environment. Inter alia, wetlands have been found to cleanse polluted water, reduce flooding, prevent soil erosion, store water, provide unique habitat for a variety of flora and fauna, They can also be used for recreation and tourism. However, wetlands face

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1 Opposed to the Cowardin approach where wetland units are distinguished at the broadest level on the basis of structural features (such as size, depth, vegetation cover and presence of surface water) that are relatively easy to identify from aerial photography and other remote sensing information sources.

ever-increasing pressure from urban and agricultural development. The National Biodiversity Assessment (2011) found wetlands make up only 2.4% of the area of the country, and 48% of wetland ecosystems in South Africa are critically endangered. An international realisation of both the importance of wetlands and the threats they face led to the Ramsar Convention of 1971, an intergovernmental treaty signed by 136 countries that provides the framework for national action and international cooperation for the conservation and wise use of wetlands and their resources.

Wetlands are dynamic features within the landscape, as evidenced at both Groenvlei and the St Lucia – Mkuze Wetland System (Ellery et al., 2003; Taylor, 2006). Both systems are relatively young (~ 6 ka) and formed as a result of the last significant change in sea level. However, the St Lucia – Mkuze Wetland System appears more transient than Groenvlei as sediment brought into the system by five rivers has reduced the area of open water by about 60% over the past 6 000 years.

The management of wetlands poses unique difficulties in that the benefits of wetlands are seen by various agencies in different ways (Dickens et al., 2004). Also, wetlands need to be understood in terms of their drivers (hydrology, geomorphology, water quality) and modifiers (land use, including vegetation alteration) (DWAF, 2007). Consequently input is required from a range of disciplines, including hydrology, geohydrology, geomorphology, soil science, botany and ecology.

Hydrological drivers of wetlands include direct rainfall and evapotranspiration [1], river inflow and outflow (including flooding), estuarine exchange and groundwater inflow and outflow. The lack of importance given to the role of groundwater in driving wetlands may result from a failure to appreciate the distinction between hydrology and geohydrology, or a lack of understanding of the subsurface, and may result in either over-protecting a water body by being too restrictive or not protecting it enough. In both instances, the Ramsar vision to promote the wise use of wetlands would have failed.

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1 In this study, the general term of evapotranspiration is used to include both evaporation from open bodies of water and soil and transpiration by plants. When either evaporation or transpiration is used, it refers specifically to that form of loss.

2.4.4 Groundwater – lake interaction

Lakes are a type of wetland, with Groenvlei being classified as a lacustrine wetland (see Section 4.8). Interaction between groundwater and lakes has been widely researched, both internationally and locally. Early work was done by McBride and Pfannkuch (1975) and Winter (1976). More recently, the coastal lakes of Western Australia have been investigated by Townley et al. (1993), Turner et al. (2000), Turner and Townley (2006) and others. Winter (1976, 1981, 1999, 2000 and 2001) and Winter et al. (1999, 2003) documented their findings and understanding of groundwater – lake interaction from their research of American lakes, both big and small. The groundwater contribution to Lake Sibaya on the Zululand Coastal Plain along the north-eastern coast of South Africa has been studied by Pitman (1980), Meyer and Godfrey (1995), Godfrey and Todd (2002), Meyer and Godfrey (2003) and Weitz and Demlie (2013). Other particularly insightful case studies are those of the Sparkling Lake (Krabbenhoft et al., 1990) and Wilton wetland complex (Hunt et al., 1996), both located in Wisconsin; the Kenyan Rift Valley lakes (Becht et al., 2006) and Lake Starr in Florida (Virdi et al., 2013). Knowledge from these and similar studies facilitated development of a sound conceptual understanding of groundwater – lake interactions.

As with rivers and other wetlands, the interaction between groundwater and lakes is affected by the position of water bodies with respect to groundwater flow systems, the geological characteristic of their beds and their climatic settings. Born et al. (1979) and Winter et al. (1999) observed that lakes interact with groundwater in three basic ways:

- Some lakes receive groundwater inflow throughout their entire bed (discharge system);
- Some lakes have seepage loss to groundwater throughout their entire bed (recharge system); and
- Most lakes receive groundwater inflow through part of their bed and have seepage loss to groundwater through other parts (flow-through system).

Interactions can be transient, shifting from flow-through to groundwater recharge to groundwater discharge in a season. Long-term study of lakes in Nebraska showed mounds formed around lakes directly after precipitation, causing groundwater discharge into the lake.

Evaporation and transpiration from the fringe of the lake, where the water table is shallow, caused the water table to drop and reverse the direction of groundwater - lake interaction.

While mechanisms for river – groundwater interaction and lake – groundwater interaction may be similar, there are some important differences. The water level of lakes generally does not change as rapidly as the water level of rivers. As a result, bank storage is of lesser importance in lakes than rivers. Evapotranspiration generally has a greater effect on lake levels than river levels because the surface area of lakes is generally larger and less shaded than many reaches of rivers, and because lake water is not replenished as readily as a reach of a river. Furthermore, lake sediments commonly have greater volumes of organic deposits than rivers. These poorly permeable organic deposits can affect the distribution of seepage and biogeochemical exchanges of water and solutes more in lakes than in rivers.

Boyle (1994) classified six lake types based on contribution of surface and groundwater (Figure 2.2). The complexity of relationships between lakes and surrounding groundwater is heightened by seasonal or longer-term temporal changes. Born et al. (1979) reported groundwater dominated the hydrology of Lake Pickerel, with groundwater accounting for 72% of inflow and 77% of losses. By contrast, groundwater accounted for only 12% of inflow into East Twin Lake in Ohio, and groundwater outflow was negligible. Turner et al. (2000) found rainfall (55% - 60%) and groundwater (40% - 45%) accounted for inflow into Lake Jasper – a surface water body of similar size as Groenvlei – while evapotranspiration (50% - 55%) and groundwater discharge (45% - 50%) accounted for outflow.

Following the early work of McBride and Pfannkuch (1975), hydrologists accept most groundwater that enters a lake system does so through the littoral zone. They found the decrease in seepage rate was approximately exponential, with the rate of decrease being an order of magnitude for every 60 m of distance (Figure 2.3). Using measurements from seepage meters, Boyle (1994) showed a threefold drop in seepage influx from a point 8 m from shore (806 $cm^3/m^2/s$) and a point 20 m from shore (265 $cm^3/m^2/s$). A further twofold drop was measured at a point 50 m from shore (19 $cm^3/m^2/s$). Measurements in the non-littoral zone ranged between 0.0 and 2.6 $cm^3/m^2/s$, with an average of 0.6 $cm^3/m^2/s$. Little groundwater was contributed to lakes in the deeper portions. In his evaluation of Lake Moses (Washington State, USA), Pitz (2003) also observed most discharge into the lake took place through the littoral zone.

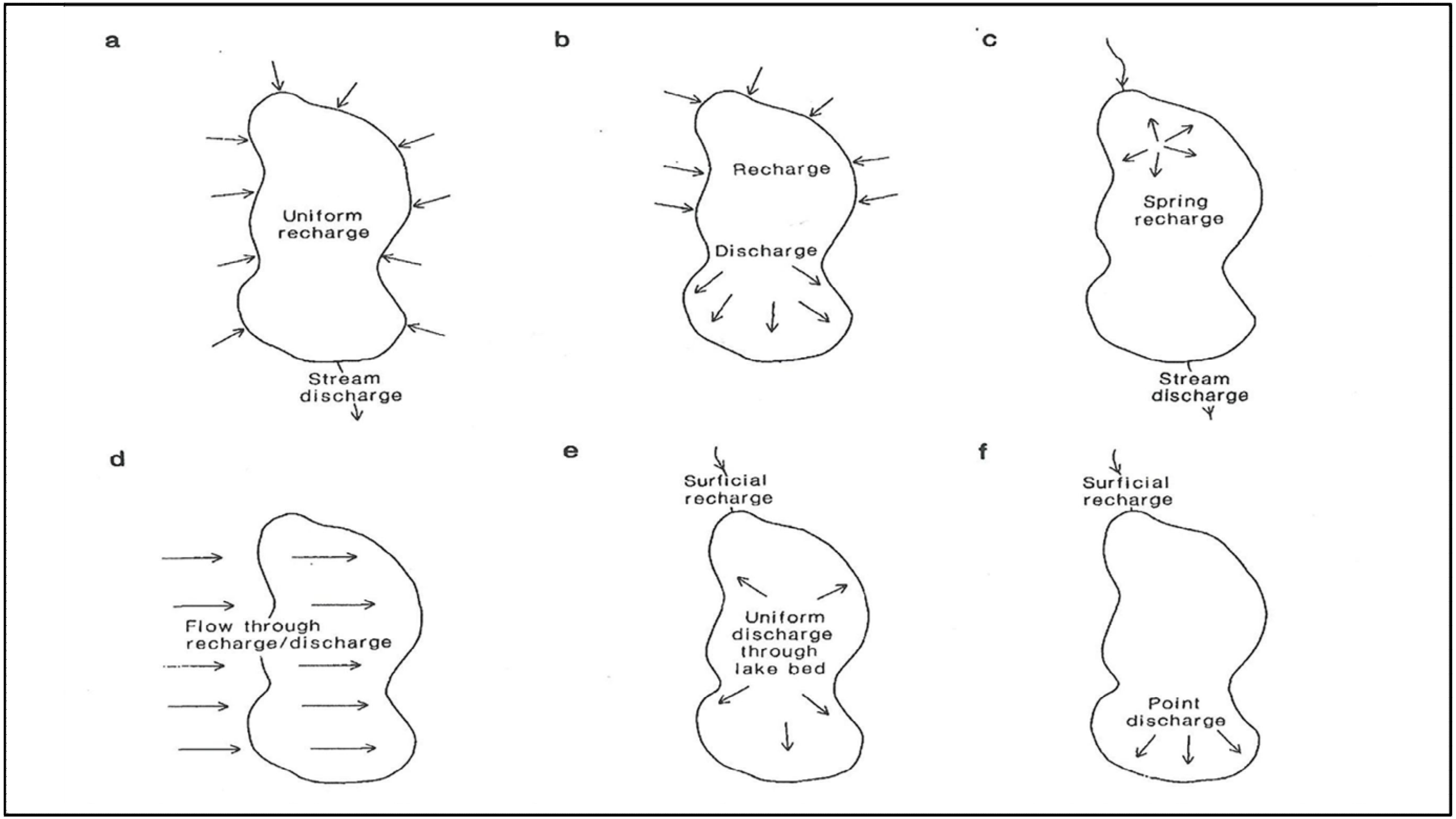

Figure 2.2: Classification of lake environments into six main classes based on relative contribution of surface and groundwater (Boyle, 1994).

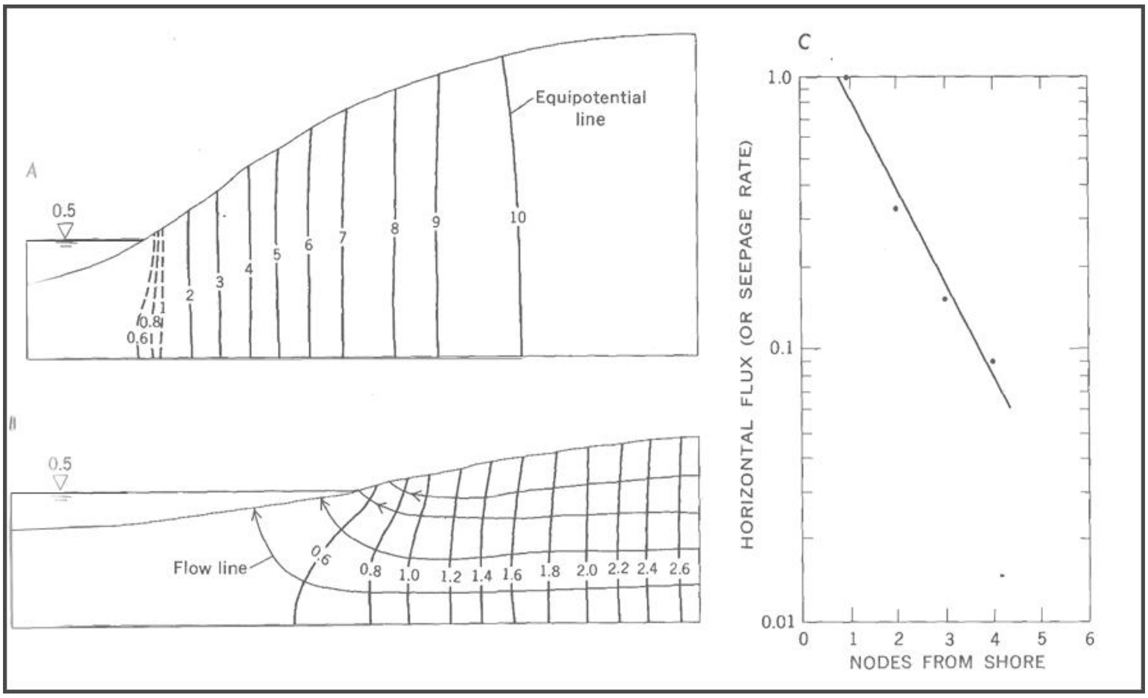

Figure 2.3: A typical vertical section model (a) Potential distribution in the entire section (b) Left third of entire section showing near-shore concentration of flow lines (c) Relation between horizontal flux (or seepage rate) and distance from shore (from McBride and Pfannkuch, 1975).

An important difference between lakes (lacustrine wetlands) and palustrine wetlands relates to wave action (Winter et al., 1999). Wind action across lakes creates wave action that disturbs the sediments in the shallow edges of a lake. This prevents a build-up of fine sediments in this zone, and in turn allows water to easily move through the lake beds. In the deeper parts of the lake, where the effect of wave action is less, fine sediments settle out of suspension to form a low-permeability bed that retards the movement of water through it. As a result, the transfer of water and solutes between groundwater and surface water is likely to be much slower.

Roningen and Burbey (2012) observed lakes and wetlands with no stream flow in or out of them are especially vulnerable to drought. This is not the case with Groenvlei where the drought of 2009 – 2010 had little effect on groundwater levels (see Section 5.4). Winter (2000) addressed vulnerability of wetlands to changes in climate. He found vulnerability depends on position within the hydrologic landscape and dependence on groundwater. Wetlands dependent primarily on discharge from regional groundwater flow systems are least vulnerable, while those dependent primarily on precipitation for their water supply are most vulnerable to climate change.

2.4.5 Water quality aspects of lake – groundwater interaction

It is as important to understand the hydrochemistry of groundwater and lakes as it is to understand the hydraulics. Two fundamental controls on water chemistry in drainage basins are the type of geologic materials present and the length of time water is in contact with those materials (Winter et al., 1999). Chemical reactions that affect the biological and geochemical characteristics of a basin include:

- Acid-base reactions;
- Precipitation and dissolution of minerals;
- Sorption and ion exchange;
- Oxidation-reduction reactions;
- Biodegradation; and
- Dissolution and exsolution of gases

Variance in geology, time and chemical reactions result in water quality being characteristics of particular facies and evolving in time (Figure 2.4). Appreciation of the hydrochemistry of groundwater and lake water – including isotopic composition – could demonstrate the exchange of water between surface and subsurface water bodies.

Winter (1999) observed the chemistry of lake waters reflects the magnitude of the groundwater flow system that discharges from them. Also, the interaction between water and vegetation can be important, especially providing a basis for using constructed or artificial wetlands to treat or manage impaired waters.

Chloride is recognised as a conservative tracer in that chloride ions do not enter into oxidation or reduction reactions, form no important solute complexes with other ions unless the chloride concentration is extremely high, do not form salts of low solubility, are not significantly adsorbed on mineral surfaces, and play few vital biochemical roles (Hem, 1985). Research by Svensson et al. (2010) addressed the cycling of chloride and its interaction with soil and vegetation. They reported that chloride participates in a complex biogeochemical cycle and uptake by biota and vegetation. In a study of 32 forested catchments in eastern North America and Europe spanning a wide range of climatic, hydrologic, geologic and vegetative conditions, they found less than half of their case studies had a nett Cl balance within 10%. Further, there were more sites (70%) showing a nett release than those showing a nett retention of Cl.

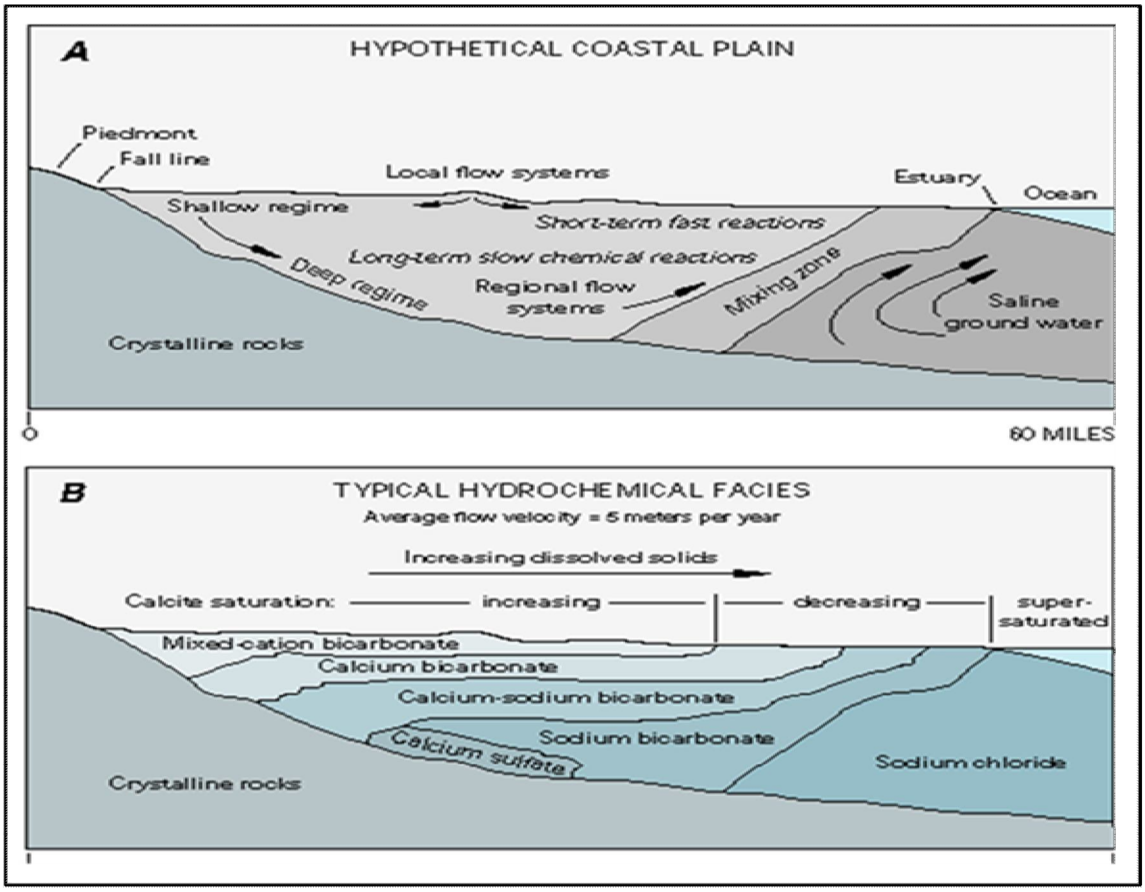

Figure 2.4: In a coastal plain, the interrelations of different rock types, shallow and deep ground-water flow systems and mixing with saline water (A) result in the evolution of a number of different groundwater chemical types (B) (Winter et al. 1999).

2.4.6 Hyporheic zone

The zone transitional between surface water and groundwater has emerged as being of importance in understanding surface – groundwater interaction. Smith (2005) noted the hyporheic zone as a critical interface between groundwater and surface water environments and a dynamic ecotone characterised by steep chemical and biological gradients. From his research, he further postulated that the potential for attenuation of pollutants is significantly greater in the hyporheic zone than in a surrounding aquifer. Consequently, interest has recently started to concentrate on the chemical pollutant attenuation capacity of the hyporheic zone and its role as a buffer in decreasing the impact of polluted groundwater on receiving surface waters.

Hydraulic properties of the hyporheic zone can be altered by biological activity, colmation and / or precipitation or dissolution of chemicals as water moves from surface to groundwater conditions. The degree of alteration would depend on site conditions. Environmental Agency (2009) presented two profiles of Cl concentration below the River Tame in Birmingham. In the first profile, Cl concentrations reduce from 150 mg /L to 60 mg/L over a distance of 0.6 m. The second profile was less distinct, but with a similar degree of reduction in concentration. Harvey and Jackson (1998) obtained a distinct pattern of Cl concentration reduction with depth, from 3.2 mg/L to 0.8 mg/L over a distance of 10 cm.

2.4.7 Anthropogenic impact on lake – groundwater interaction

Parsons (2000) found artificial water level regulation of Zeekoevlei and construction of the Cape Flats wastewater treatment works directly south of the wetland significantly impacted groundwater – lake interaction, reversing the direction of groundwater flow along the southern boundary.

Aguilera et al. (2013) reported groundwater abstraction from an area in the vicinity of an important semiarid wetland in Spain caused an inversion of groundwater flow, resulting in the system changing from a groundwater discharge zone into a recharge zone. They argued anthropisation should be seen as an added intrinsic property rather than an external disturbance. Management strategies should be based on realistic conceptual models which represent actual conditions rather than past natural conditions or future desired states.

2.5 Methods for Quantifying Groundwater - Lake Interaction

2.5.1 Preamble

A range of methods are available for quantifying groundwater - lake interaction, including water balance methods, chemical mass balance approaches and numerical modelling (Sacks et al., 1998; Winter, 1999; Virdi et al., 2013). The successful application of any of these methods has as its foundation a sound conceptual model. Also, use of a particular approach is dependent on available data and assumptions associated with the method.

Numerical models have been used by Vivier (2009), Virdi et al. (2013) and others to address groundwater - lake interaction. Computer-based modelling is a powerful tool for incorporating mathematical complexities and large data sets. However, a weakness of numerical modelling is:

- The type and volume of data and information required; and
- Assumptions that have to be made to be able to apply the model.

In the absence of a well-distributed data set and a poor understanding of the exchange processes through the lake bed (leakance), it was decided at the outset of this research to apply a more empirical and analytical approach better suited to the available data. An advantage of the analytical approach is that it allows for the continual checking of the conceptualisation and assumptions during the assessment.

2.5.2 Water balance method

2.5.2.1 Theoretical considerations

The water balance [1] is based on the conservation of mass and is widely used in hydrological studies (Healy et al., 2007). For example, Carter (1996) referred to a wetland water balance where individual components interrelate to create the hydrology of the wetland. Sacks et al. (1998) used a water balance to study 10 lakes in Florida. Bredehoeft (2002) used a water

[1] In American literature generally referred to as a water budget.

balance to explain the principles of sustainability and why recharge does not define the sustainable yield of an aquifer. The drying and salinisation of the Aral Sea was described by Benduhn and Renard (2004) using a water balance model. Birkhead et al. (2007) used a water balance approach to model Nysvlei, a Ramsar-designated wetland in the north-eastern parts of South Africa. Roningen and Burbey (2012) developed a daily water balance for Mountain Lake (USA) to understand hydrological controls of lake levels. McMahon et al. (2013) described the use of a simple water balance to estimate annual actual evaporation from an unimpaired catchment. A water balance was used as a preliminary investigative tool by Gomo and van Tonder (2013) to assess surface water – groundwater interaction along a section of the Modder River near Bloemfontein.

At first introduction, the water balance equation looks deceptively simple (Winter, 1981). Inputs and outputs are balanced against a change in storage. In its simplest form, the water balance is written as:

$$\mathbf{I + O = \Delta s}$$

where	I	=	input
	O	=	output
	Δs	=	change in storage

Expanding the equation to consider individual components of the input and output terms results in:

$$\mathbf{(P + R_i + G_i) - (E + R_o + G_o + A) = \Delta s}$$

where	P	=	precipitation
	R_i	=	run-off (in)
	G_i	=	groundwater (in)
	E	=	evapotranspiration
	R_o	=	run-off (out)
	G_o	=	groundwater (out)
	A	=	abstraction

Ward (1975) noted that if the equation can be solved, a quantitative assessment of the movement of water over, through and across the land is possible. In addition to being used to quantify water quantities, the water balance method can also be used to consider water quality by balancing loads (i.e. quantity multiplied by quality). The water balance is particularly useful if one component of the system is unknown. By rearranging the water balance, the unknown can be determined as the residual of the other terms.

The relative importance of each component of the water balance varies both spatially and temporally. Determining the balance is imprecise because as the climate varies from year to year so does the water balance. The accuracy of individual components depends on how well they can be measured and on the magnitude of the associated errors (Winter, 1981; Carter, 1996). However, water balances - in conjunction with information on the local geology - provide a basis for understanding hydrologic processes, water chemistry, wetland functions and predicting the effects of natural or human-induced hydrologic alterations on wetlands.

The water balance can be applied at a range of time scales (Healy et al., 2007). The temporal scale of application is generally governed by the data available. Annual balances can help develop a broad understanding of the hydrological functioning of a system, while daily or hourly balances may help develop a deeper appreciation of seasonal fluctuations.

A drawback of the water balance is that hydrological components not measured are lumped together with errors of the components that are measured. Errors and uncertainty relate to those components that are measured (e.g. rainfall) and those interpreted (e.g. evapotranspiration). Rainfall is widely measured in South Africa and measurement is relatively simple. However, errors result from instrumentation used, placement of rain gauges, measurement and data recording, and spatial and temporal variability of each individual rainfall event. Evaporation is also widely measured, but determining evaporation from an open body of water and transpiration by plants remains a challenge (see Sections 2.5.2.5 and 2.5.2.6). Failure to appreciate errors in the component data can undermine the outcome of water balance studies and the value of the residual determined.

Pitman (1980) and Fijen (1995) – engineers and hydrologists by training – applied a water balance approach to quantifying the hydrological components of Lake Sibaya and Groenvlei. In both instances, they used groundwater inflow to balance their water budgets and assumed little groundwater outflow. This assumption was flawed given the transmissive nature of the subsurface and the prevailing hydraulic gradients at both sites. Also, no attempt was made to verify or test the reasonableness of the outcome of the water balance.

All water balance calculations are subject to a degree of uncertainty, the major source of which is natural variability of component parts and errors in measurement (Healy et al., 2007). Although the method is widely used, it is adequate only if all the components of the balance are accurately measured.

2.5.2.2 Quantifying lake storage

Change in volume of water in the lake is the focal point of the water balance and associated directly with changes in water level [1] (Roningen and Burbey, 2012). The change in volume is readily calculated using monitored changes in water level and bathymetric information. A lake-specific area – capacity curve can be developed to facilitate the accurate determination of a change in storage.

2.5.2.3 Quantifying rainfall

Mean annual precipitation is a widely used parameter in hydrological studies, with direct measurement achieved using a rain gauge. It can also be achieved through indirect techniques such as those using radar. Though relatively simple, the measurement of rainfall and the associated errors are a research topic on its own (Rodda, 1967; Ward, 1975; Lynch, 2004; Sevruk, 2006; Sieck et al., 2007; Rodda and Dixon, 2012 and others). Rodda (1967) measured a 6.6% difference over a period of 5 years between a rain gauge installed at ground level and one 30 cm above ground. Monthly differences ranged between 5% and 10%. In a catchment covering 135 km^2 and using 49 rain gauges, Wood et al. (2000) recorded differences ranging between 33% and 65%. Winter (1981) noted error generally decreased with longer time periods of monitoring. He summarised instrument error could be in the order of 5%, and instrument placement error could be in the order of 5% to 15% for long-term data and as high as 75% for individual storms. Errors associated with aerial averaging of

[1] Sometimes referred to as the stage.

point data could be as high as 65%. Accepting rainfall data at face value will result in a failure to appreciate the error and uncertainty contained therein.

The use of multiple rain gauges at a lake can be used to overcome or reduce some uncertainty. However, this is not always practical. Multiple records from a single catchment could also be used, but the spatial distribution of rainfall events complicates interpretation.

2.5.2.4 Quantifying run-off

There is sufficient evidence to show that there is no surface water run-off into or out of Groenvlei (see Section 6.1.1). As a result, there is no need to address measurement or modelling of surface flows in this study.

2.5.2.5 Quantifying lake evaporation

Evaporation is a collective term covering all processes in which liquid water is transferred as a vapour to the atmosphere (McMahon et al., 2013). Direct measurement of evaporation losses from an open body of water is rarely undertaken. Clulow et al. (2012) noted that despite improvements to measurement techniques and the dominant role of evaporation in wetland water balances, there are few studies in southern Africa with actual measurements of evaporation from wetlands. A range of indirect methods are available to estimate such losses, including:

- Estimation based on measured pan evaporation data
- Water balance calculations
- Energy balance calculations
- Mass transfer procedures
- Combination of these

ASCE (1996) observed the method used to estimate actual evaporation is generally a function of available data, available resources and the level of accuracy needed. McMahon et al. (2013) provided a comprehensive summary of evaporation estimation techniques and pragmatic guidance on their application. Local perspectives on the topic were provided by

McKenzie and Craig (1999), Blight (2002), Kleynhans (2005) and Clulow et al. (2012). The lack of site-specific climate data required by many of the computational methods (e.g. Penman, Penman-Monteith and Priestly-Taylor) precluded their use and application in this study.

Watkins (1993) observed evaporation pans are the most commonly used low-technology measure of evaporation from an open body of water. In South Africa, evaporation is measured using either a Class A pan or Symons pan (or S pan). A S pan was employed at station K3E003 at Rondevlei. This is a square pan and is sunk into the ground, as opposed to an A pan which is circular and installed above ground. The S pan is made of galvanised steel, has sides 1 830 mm long and is 610 mm deep. The rim of the pan is installed 100 mm above ground level. Kleynhans (2005) warned evaporation pans can be sensitive to inaccuracies if not used correctly. There are more than 27 different types of evaporation pans (Winter, 1981), with all variation of design intended to overcome two basic problems:

- Effects of wind patterns; and
- Effects of advection of heat energy not representative of the natural environment.

Evaporation measurements from A pans and S pans differ significantly, and measured A pan evaporation is on average 1.1 to 1.3 times higher than the S pan values (Mare, 2007). Mare (2007) reported he used a factor of between 0.85 and 0.91 to convert A pan measurements to S pan measurements. Bosman (1990) (as quoted by Midgley et al., 1994) provided the following relationships for converting monthly A pan and S pan measurements and vice versa:

$$\textbf{A pan} = \mathbf{26.3622 + 1.0786 \times} \textbf{ S pan}$$

$$\textbf{S pan} = \mathbf{130 + 0.726}\textbf{ A pan}$$

Difficulties relating pan evaporation to evaporation from a dam or lake arise as a result of differences in wind, temperature, water depth and water quality (Ward, 1975; Linsley et al., 1982, Oroud, 1995; McMahon et al., 2013), with pan evaporation generally over-estimating actual evaporation. Many scientific studies have investigated this issue, a detailed discussion of which is beyond the scope of this thesis. Jarmain et al. (2008) compared 7 different methods of determining evaporation from an open water surface and concluded S pan measurements compared well with other techniques when the pan was well maintained.

McMahon et al. (2013) presented the following general equation for estimating actual evaporation from A pan data:

$$\mathbf{E_{fw.j} = K_j \, E_{pan.j}}$$

where:	$E_{fw.j}$	=	estimate of monthly (or daily) open-surface water evaporation in (mm / unit time)
	$_j$	=	specific month (or day)
	K_j	=	average monthly (or daily) A pan coefficient
	$E_{pan.j}$	=	monthly (or daily) observed A pan value in (mm / unit time)

They provided coefficients for 68 locations across Australia by correlating monthly evaporation values from A pans with corresponding Penman evaporation estimates. Monthly coefficients ranged between 0.558 and 1.546. McMahon et al. (2013) found a close correlation between their mean monthly pan coefficients (weighted for length of record) of 0.80 to that obtained using the Penman method (0.76).

Watkins (1993) observed A pan measurements are adjusted using a coefficient of 0.7, the approach also used by Roningen and Burbey (2012). Hounam (1973, as quoted by Winter, 1981) found the coefficient to range between 0.52 and 0.86. Hounam reported the monthly A pan coefficient at Lake Hefner (USA) ranged between 0.35 (May) and 1.32 (November). The range at Lake Eucumbene (Australia) was 2.04 (May) and 0.48 (November). As a result, Winter (1981) cautioned against applying the 0.7 coefficient to records of less than a year in duration.

In their water budget of Lake Sibaya, Pitman and Hutchison (1975) developed pan coefficients to apply to S pan data to determine lake evaporation (Table 2.4), with relatively large differences between summer and winter. Van Vuuren (2012) used pan coefficients provided by Midgley et al. (1994) to convert S pan data to lake evaporation (Table 2.5). This approach is widely used in rainfall – runoff modelling in South Africa, where the measured S pan evaporation is multiplied by a monthly pan factor to derive evaporation from open water.

Table 2.4: Pan factors used to determine lake evaporation of Lake Sibaya from S pan data.

Measure	O	N	D	J	F	M	A	M	J	J	A	S
Lake evaporation pan factor	0.75	0.85	0.90	0.94	0.97	1.00	1.00	1.00	1.00	0.75	0.70	0.65

(Pitman and Hutchison, 1975)

Table 2.5: Pan factors applied to S pan data to determine lake evaporation.

Measure	O	N	D	J	F	M	A	M	J	J	A	S
Lake evaporation pan factor	0.83	0.81	0.81	0.81	0.82	0.83	0.84	0.88	0.88	0.88	0.87	0.85

(from Midgley et al. 1994)

From their work in the Orange River Basin, McKenzie and Craig (1999) proposed A pan evaporation be used without correction to simulate evaporation losses from South African rivers. They found river evaporation to be higher than S pan values.

Blight (2002), Kleynhans (2005) and Birkhead et al. (2007) addressed evaporation losses from Nylsvlei. By comparing estimates of evaporation using energy balance measurements and empirical methods, it was recommended the evaporation rates presented in Table 2.6 be used to quantify evaporation losses from the wetland.

Table 2.6: Average daily evaporation (in mm/d) from Nylsvlei recommended by Kleynhans (2005).

Measure	O	N	D	J	F	M	A	M	J	J	A	S
Lake Evaporation	2.3	3.0	3.7	4.4	4.7	2.6	2.1	1.9	1.6	1.6	0.4	2.3

Fijen (1995) used a factor of 0.8 applied to S pan data for the months of July through to October to determine evaporation from Groenvlei. For the remaining months he applied a factor of 1.0. Vivier (2009) assumed evaporation from Groenvlei to be 75% of A pan evaporation measured at the George weather station 40 km west of the lake. Paridaens and Vandenbroucke (2007) argued it was better to use the S pan evaporation measurement directly and without correction as:

- The measurement provides an integrated effect of radiation, wind, temperature and humidity; and
- Pan measurements generally overestimate actual evaporation.

They justify their position by noting DWA adjusts the S pan measurement using a coefficient ranging between 0.8 and 1.0, and the good match between the S pan evaporation and their calculation of the reference crop evapotranspiration (Table 2.7).

Table 2.7: Comparison of S pan measurements and calculated reference crop evapotranspiration for Groenvlei. All units are mm/d.

Measure	O	N	D	J	F	M	A	M	J	J	A	S
S Pan	3.0	3.8	4.3	4.3	3.9	2.9	2.1	1.6	1.3	1.4	1.7	2.2
ET_o	3.0	3.7	3.8	4.1	3.7	3.0	2.1	1.6	1.2	1.3	1.8	2.3

(from Paridaens and Vandenbroucke, 2007)

However, their decision to use S pan evaporation without any correction is problematic as their water balance has the amount of groundwater flowing out of the system (0.22 mm/d) being significantly less than that flowing into the lake (1.21 mm/d). Accepting T and i are similar along the inflow and outflow boundaries, then the inflow boundary has to be 5.5 times longer than outflow boundary. Neither their temperature profile data nor a conceptual understanding of groundwater flow at Groenvlei supports such a condition.

Clulow et al. (2012) measured and modelled evaporation from the Mfabeni swamp directly east of Lake St Lucia in Maputuland using a surface renewal method. The situation of the mire has similarities to Groenvlei. Clulow et al. (2012) measured evaporation to be 900 mm/a with summer maximums of 6.0 mm/d and winter rates of 1.7 mm/d. Interestingly, their measured evaporation was much lower than the S pan (1 446 mm/a) and A pan (1 706 mm/a) values presented by Midgley et al. (1994) for nearby station W3E001.

While the situation of Nylsvlei is dissimilar to that of Groenvlei, the study by Birkhead et al. (2007) highlighted the difficulty of quantifying evaporation losses from wetlands (Figure 2.5). The seasonal variation – as predicted using a range of methods – is constant, but the quantum of loss ranges from the high reflected by A pan measurements (2 400 mm/a, with S pan equivalent of 1 870 mm/a) to that determined using an energy balance approach (929 mm/a). The researchers found the Pure Grassland factor applied to A pan data (Midgley et al., 1994) best reflected what they interpreted to be actual evaporation.

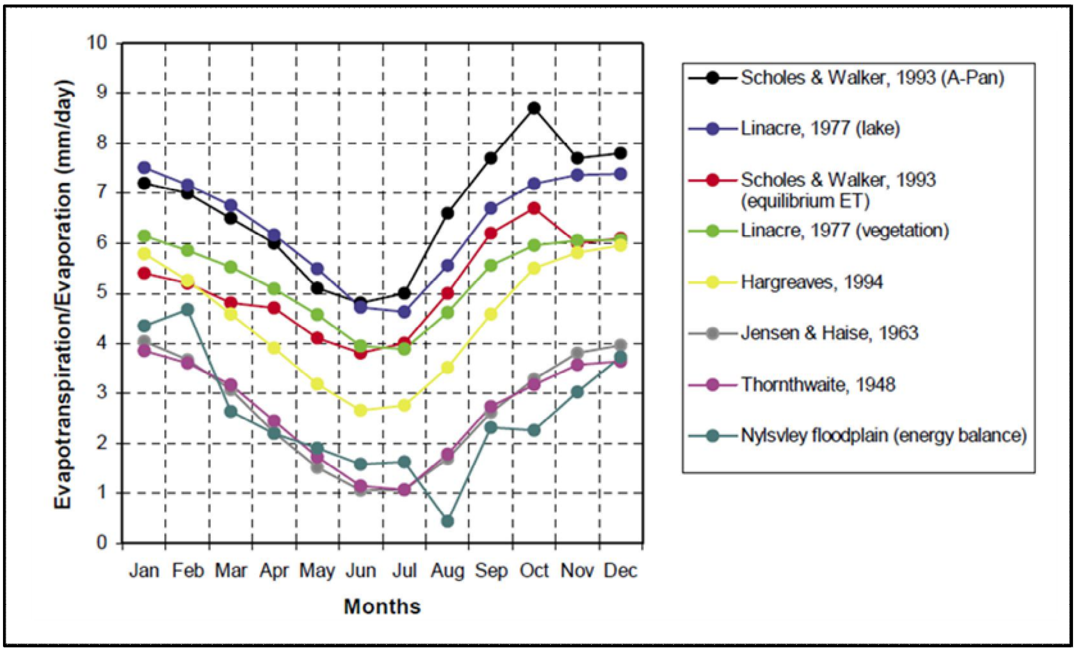

Figure 2.5: Estimates of evapotranspiration from the Nylsvlei floodplain presented by Birkhead et al. (2007).

The physical and chemical properties of water are changed by the inclusion of solutes into the solvent. In sea water, Na and Cl are typically the main solutes. Those substances with a high vapour pressure evaporate more easily than those with a low vapour pressure. At 20^{o}C, the vapour pressure of water is 2.3 kPa. By comparison, more volatile substances such as ethanol, butane and propane have vapour pressures of 5.8 kPa, 220 kPa and 1.0 MPa. According to Raoult's Law, the total vapour pressure of a solution is directly dependent on the vapour pressure of each chemical component and the mole fraction of the component present in the solution. Salt water, for example, contains non-volatile substances such as Na and Cl, and hence evaporates less readily than pure water. However, ITTC (2011) presented data showing the vapour pressure of seawater to be very similar to that of freshwater (Figure 2.6). Ward (1975) observed evaporation decreased by about 1% for every 1 % increase in salinity, so that evaporation of seawater (3.5% salinity) is some 2% to 3% less than evaporation from freshwater. Finley and Jones (undated) showed seawater could evaporate at a rate 4% less than freshwater. Benduhn and Renard (2004) determined the same difference in rates of loss when modelling the Aral Sea water and salt balance. The brackish character of Groenvlei's water is unlikely to significantly affect evaporation, and any effects are likely to be less than levels of uncertainty introduced by measurement errors and converting pan losses to those from open bodies of water.

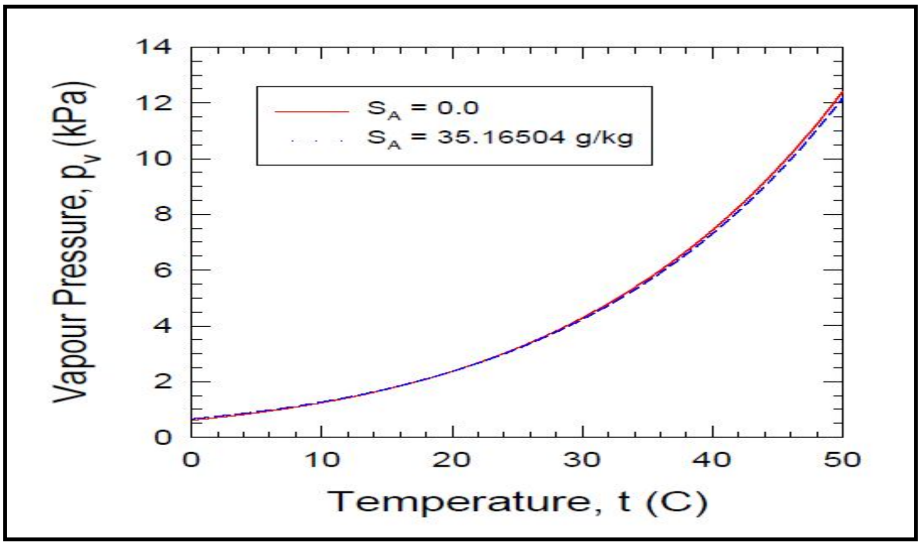

Figure 2.6: A comparison of the vapour pressure of freshwater to that of seawater (ITTC, 2011).

The turbidity of the water – also a component of its water quality – could affect the albedo of the water, and consequently its heat budget and temperature. However, this probably has little direct effect on evaporation and can be discounted.

Evaporation removes only water molecules, with salts such as NaCl dissociating from the water. If enough water evaporates, the remaining water becomes increasingly saline and eventually produces salty brine. The brine becomes supersaturated with continued evaporation and salt deposits begin to precipitate, forming evaporate minerals such as halite. However, a small amount of salts can be removed from saline water bodies when entrained in wind-blown spray, contributing to the salinity of rainwater.

2.5.2.6 Consideration of transpiration

About 32% of the area of Groenvlei comprises reed and reed swamp vegetation (i.e. is not open water), and this could greatly influence evapotranspiration losses from the lake. Goulden et al. (2007) remarked that after a century of investigation, controls on wetland evapotranspiration remain poorly understood. Some researchers argue evapotranspiration from wetlands is less than that from open water, while others have found it to be greater. Obeysekera (as quoted by McMahon et al., 2013) showed transpiration by macrophytes was greater than open surface water evaporation while Headley et al. (2012) reported transpiration by *Phragmites australis* to be 2.6 times greater than that measured at an A pan. Reporting results of a study in Poland, Gilman (2002) noted transpiration could range between 92% and 127% of evaporative losses from open water. McKenzie and Craig (1999) motivated a factor of 1.0 be used (i.e. reed transpiration was equivalent to evaporation from open water). From lysimeter studies at two sites in Italy, Borin et al. (2011) found the average summer plant coefficient [1] of *Phragmites australis* to be 6.5. Observations about the significance of losses from the riparian zone were also made by Everson et al. (2001), Hughes (2004), Tanner (2013) and others. Consequently it was decided to use the total area of the wetland and lake (359 ha) in this modelling exercise, as opposed to the area of the open water body only (245 ha).

[1] Defined as evapotranspiration of plant divided by the calculated reference crop evapotranspiration using the standardized Penman–Monteith equation.

Everson et al. (2001) found distinct seasonal patterns in transpiration losses from the riparian zone of the Sabie River. In summer, losses were measured to be 9 mm/d while in winter losses reduced to 4 mm/d. They found the modelled transpiration losses of between 10 mm/d and 30 mm/d presented by Birkhead et al. (1999) unrealistically high. Similar seasonal patterns in transpiration from wetlands were measured by Sânchez-Carrillo et al. (2001), Schilling and Kiniry (2007), Dye et al. (2008), Borin et al. (2011) and others.

It is well established some vegetation is dormant in the cooler winter months and transpiration losses are minimal, if any (Carter, 1996; Holt et al. 1998; Acreman, 2003; ITRC, 2009). Goulden et al. (2007) reported summer evapotranspiration from a Typha-dominated marsh in the San Joaquin Valley, California, was in the order of 4 mm/d, but in winter was just a few tenths of a millimetre each day (Figure 2.7). A study of *Phragmites australis* in China by Zhou and Zhou (2009) found evapotranspiration from a marsh accounted for 376 mm losses in the growing season, but only 56 mm outside of the growing season i.e. 15%.

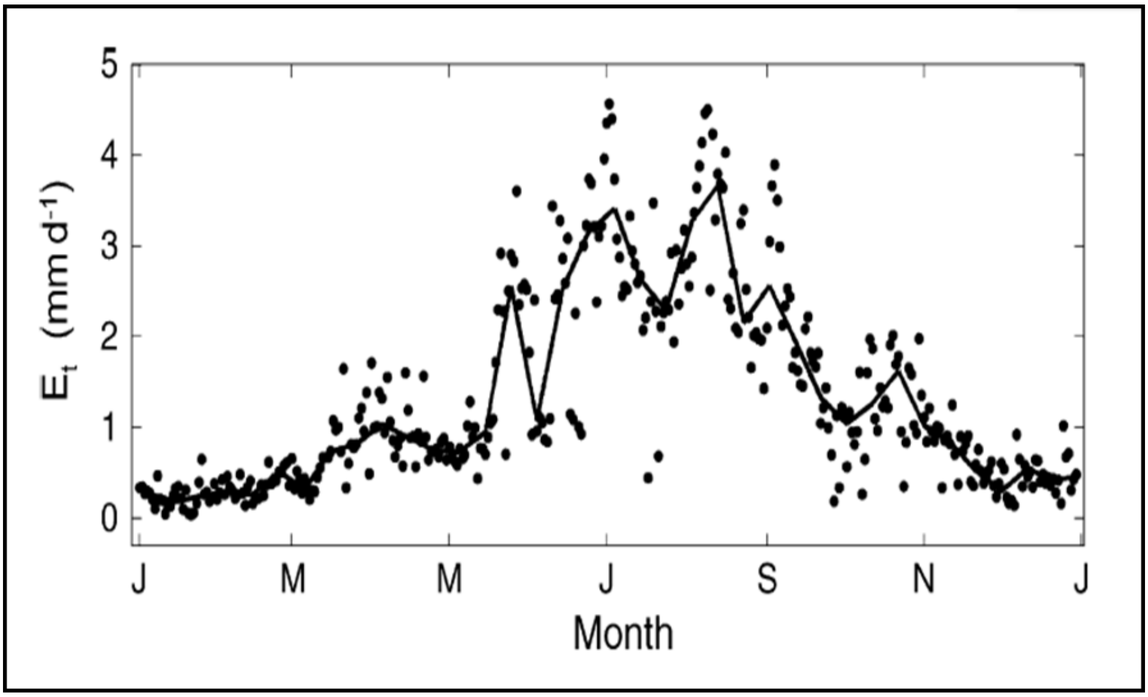

Figure 2.7: Evapotranspiration measured at a Typha-dominated marsh subjected to a Mediterranean climate (Gouldin et al., 2007).

2.5.3 Groundwater flow calculations

In his assessment of 23 water balance studies of lakes in North America, Winter (1981) found groundwater was seldom specifically examined. In three of the studies groundwater was not even mentioned. More recent studies by Becht et al. (2006), Turner and Townley (2006), Virdi et al. (2013) and others recognise the groundwater or potential groundwater component of lake water balances.

Groundwater flow can be quantified using Darcy's Law. Using results of column experiments conducted in 1855 and 1856, Henry Darcy showed discharge of water through sand is proportional to the head and inversely proportional to the thickness of the sand layer that the water passes through. For laminar flow and steady-state conditions, Darcy's Law is written as:

$$\mathbf{Q = K\ i\ A}$$

where	Q	=	discharge (m^3/d)
	K	=	hydraulic conductivity (m/d)
	i	=	hydraulic gradient
	A	=	area (m^2)

As T is a product of K and aquifer thickness (b), the equation can be rewritten as:

$$\mathbf{Q = T\ i\ W}$$

where	T	=	transmissivity (m^2/d)
	W	=	width (m)

The Dupuit assumption assumes groundwater flow in unconfined aquifers is horizontal. Application of Darcy's Law to the Groenvlei case study allowed groundwater flow into and out of the lake to be quantified. Errors and uncertainty in estimating flow emanate from errors in estimating various components of the equation.

2.5.3.1 Quantifying aquifer parameters

Seepage meters

Seepage meters were first developed in 1944 to measure water losses from irrigation canals (Taniguchi and Fukuo, 1993). They have since been used to measure seepage from and into rivers, lakes and the sea. McBride and Pfannkuch (1975) proposed seepage meters used near the shore could quantify the seepage component of the water balance independently from the other components, possibly more accurately than determining seepage as a residual in a water balance.

Seepage meters provide a point sample both in place and time. While inexpensive and simple in design, the method is prone to installation errors, errors in measurement and fluctuation with time. Murdoch and Kelly (2003) found the performance of the device far from simple. For example, the build-up of hydraulic head inside the device could divert flow way from it. More sophisticated (and expensive) devices have been used with success, for example by Sholkovitz et al. (2003) to investigate submarine groundwater discharge and tidal influences. Boyle (1994) used a relatively sophisticated seepage flow meter to successfully measure fluxes both into and out of Alexander Lake in Canada.

Paridaens and Vandenbroucke (2007) took 26 measurements at three locations west and south of Groenvlei, with the influx ranging between 1.01 mm/d and 24.18 mm/d (0.001 m/d – 0.024 m/d). Notwithstanding the errors and difficulties described by them, the hydraulic gradient at the point of measurement was not accounted for and the data failed to reflect that water discharges from the lake along the southern shores. Given the sandy nature of the aquifer, significantly higher K values obtained from other methods and failure to reflect outflow along the southern shores, the data is considered too unreliable to use in assessing the groundwater contribution to Groenvlei.

Pumping tests

Pumping tests are the preferred means of quantifying aquifer parameters as they provide a better spatially averaged value than other techniques such as grain size analysis, double-ring infiltrometer tests, slug tests and seepage meters. Also, pumping tests are *in situ*, and not subject to errors related to sample disturbance. Procedures for conducting pumping tests and methods for interpreting the data are well documented in texts such as Driscoll (1986), Kruseman and de Ridder (1990), Domenico and Schwartz (1997) and Misstear et al. (2006). Because of the predominance of fractured aquifers in South Africa, Murray (1996), van Tonder and Xu (1999) and van Tonder et al. (2002) have addressed the analysis of pumping test data from secondary aquifers.

Step tests [1] are used to assess borehole performance, and entail pumping a borehole for fixed periods (steps) at increasing rates of abstraction. The steps are usually of 60 minutes duration and a step test could entail three or more steps. The test is usually conducted until the groundwater level reaches the test pump inlet level, whereafter the pump is switched off and the recovery of the borehole is monitored. Outside of ensuring boreholes were properly developed, information provided by step tests is of little direct relevance to this study.

Constant discharge tests are used to determine hydraulic properties of aquifers, and sometimes referred to as aquifer tests. Boreholes are pumped at a constant rate for a fixed duration. The pump is then switched off, and the recovery of the borehole then monitored for an equivalent duration. The duration of constant discharge tests is guided by the proposed use of a borehole and the level of reliability required. A duration of 72 hrs of continuous pumping is the industry standard, but tests can range in length from 8 hrs to a few weeks. SANS (2003) recommended a duration of between 48 hrs and 72 hrs if boreholes are to be used for town water supply. An observation borehole some distance away from the pumped borehole is required if storativity is to be determined. If a single borehole is used in the test, only transmissivity can be assessed.

Günther Thiem was one of the first to use two or more piezometers to determine the transmissivity of an aquifer (Kruseman and de Ridder, 1990). He developed his steady-state

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[1] Nomenclature used in the literature is varied, including step drawdown test, multiple discharge test, multiple rate test, well efficiency test and well performance test.

solution in 1906. Charles Theis developed a formula in 1935 that did not require steady-state conditions – conditions seldom attained in confined aquifers. He introduced factors of time and storativity. This early work was expanded on by many others to accommodate different aquifers and responses. However, almost all pumping tests are founded on the following assumptions and conditions (Kruseman and de Ridder, 1990):

- The aquifer has a seemingly infinite areal extent;
- The aquifer is homogeneous, isotropic and of uniform thickness over the area influenced by the test;
- Prior to pumping, the water table or piezometric surface is (nearly) horizontal over the area influenced by the test;
- The aquifer is pumped at a constant rate;
- The pumped borehole penetrates the entire aquifer and thus receives water from the entire thickness of the aquifer;
- The diameter of the pumped well is infinitesimal in relation to the aquifer;
- Groundwater density and viscosity are constant; and
- Groundwater flow can be described by Darcy's Law and is assumed to flow horizontally.

Drawdowns measured during pumping tests are plotted against time on either logarithmic or semilogarithmic graph paper, with the shape of the curve being characteristic of aquifer conditions (Figure 2.8). Unconfined aquifers typically produce straight lines when plotted on semilogarithmic graphs, but delayed yield can cause a flattening of the graph before furtherance of a slope reflective of the prevailing transmissivity and storativity. The delayed yield is caused by gravity drainage replenishment from the pore spaces above the cone of depression. The Boulton or Neuman curve-fitting methods are used to interpret pumping test data from these conditions. Alternatively, the Cooper-Jacob straight line method can also be used if the delayed yield segment is not used to fit the straight line. After analysing data from 628 single-borehole tests with the Cooper-Jacob straight-line method, Halford et al. (2006) found the method performed more reasonably than most alternatives.

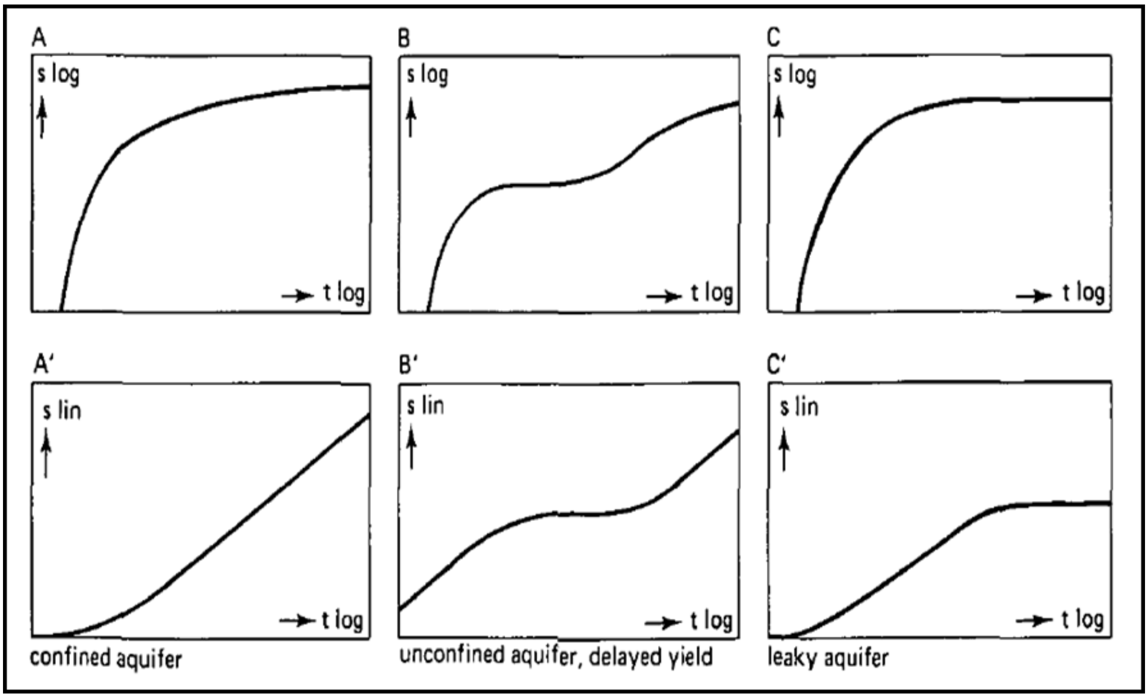

Figure 2.8: Pumping test graphs characteristic of different aquifer types (Kruseman and de Ridder, 1990).

As experienced by Vandoolaeghe (1989) on the Cape Flats, sandy unconfined aquifers can produce pumping test curves characteristic of leakage from the overlying aquifer material. In such instances, the Walton or Hantush methods can be used to interpret T and S from the pumping test data.

2.5.3.2 Quantifying hydraulic gradients

Hydraulic gradients are determined using measured groundwater level data and the position of the point at which it was measured. While the hydraulic gradient is easily determined by dividing the difference in water level by the distance between the points of measurement, the hydraulic gradient must be determined along a line perpendicular to groundwater contours. Compilation of a groundwater contour map is particularly useful for this purpose.

Because depth to groundwater can only be measured in boreholes, the number of measurements is often limited. Also, data is usually collected over a period of time. It is therefore common to use all available groundwater level data in compiling a groundwater level contour map. Errors can be reduced if monitored data are available and measurements are corrected for either seasonal or annual fluctuations. Further, the accuracy of such maps is directly related to (a) the amount and distribution of data available and (b) the method of contouring used.

It is noteworthy that the hydraulic gradient tends to show little seasonal variation, as the mutual rise and fall of groundwater levels do not significantly alter the gradient. Probably the greatest source of error related to determining hydraulic gradients is that associated with the elevation of the point of measurement. Unless determined accurately by a surveyor, elevations can be determined from maps, satellite imagery, geographical information system (GIS) datasets or using handheld global positioning systems (GPS) [1]. While the accuracy of surveyed elevations is in the order of 0.005 m and depth to groundwater some 0.01 m (if depth to water is less than ~ 30 m), the accuracy of elevation data determined from maps is probably in the range of 0.5 m to 2 m. These levels of accuracy are particularly problematic in terrains with either flat topographic or flat hydraulic gradients, or both.

Hydraulic gradients measured in coastal primary aquifers elsewhere in South Africa have generally been in the order of 0.001 to 0.003 (Henzen, 1973; Wright, 1991; Wright and Conrad, 1995; Parsons, 2002; Parsons 2009a).

2.5.3.3 Groundwater Fluxes

Earlier studies of coastal primary aquifers in South Africa have estimated groundwater flow through an aquifer or groundwater discharge to sea. This information can be presented as a flux (Table 2.8). A one-dimensional form of Darcy's Law can be written as:

$$\mathbf{Q = K\ dh/dL}$$

where	q	=	Darcian flux i.e. groundwater discharge volume per unit area per unit time
	K	=	hydraulic conductivity
	dh/dL	=	hydraulic gradient

[1] Use of inexpensive handheld GPS units is not advised as the z-orientation accuracy is too poor.

An analysis of this could be useful for quantifying the groundwater contribution to Groenvlei as it represents the collective knowledge and assumptions pertaining to flow in such systems. Use of information in this way would equate to van Tonder's (2004) concept of a "qualified guess" applied by him in estimating recharge. Excluding the high flux deduced from Henzen (1973), the average flux was calculated to be 1.11 m^3/d/m.

Table 2.8: Estimates of groundwater fluxes in coastal primary aquifers

Area	**Source**	**Flux (m^3/d/m)**
Atlantis Aquifer	DWAF (1999)	1.00
Atlantis Aquifer	van der Voort *pers.comm.* (2000)	0.90
Cape Flats Aquifer	Henzen (1973)	4.00
Cape Flats Aquifer	Gerber (1981)	0.90
Alexander Dune Field	Campbell and Bate (1998)	1.00
Alexander Dune Field	Campbell and Bate (1998)	1.50
Lake St Lucia	Rawlins (1981)	1.23
Lake St Lucia	Kelbe and Rawlins (1992)	0.78
Lake St Lucia	Kelbe and Rawlins (1992)	2.99
Mfabeni swamp	Kelbe and Rawlins (1992)	0.85
Lake St Lucia	Kelbe et al. (1995)	0.75
Lake Sibaya	Pitman and Hutchinson (1975)	0.68
Zululand Coastal Aquifer	Meyer et al. (2000)	0.90
Zululand Coastal Aquifer	Meyer et al. (2000)	1.01

2.5.4 Chemical tools

2.5.4.1 Theoretical considerations

Consideration of the hydrochemistry of the component waters provides additional tools for understanding groundwater – lake interaction. Use of the chemical mass balance equation builds on the water balance approach and allows for the disaggregation of nett groundwater contribution and quantification of groundwater inflows and outflows (Sacks et al. 1998). Piper and Durov diagrams are used to graphically represent the inorganic chemistry of water, and have long been used to "fingerprint" groundwaters. Turner and Townley (2006) were able to delineate lake release zones using hydrogeochemical signatures of chloride and stable isotopes. Evaporation from open water is the only process that can cause isotopic enrichment, allowing for the deduction that isotopically enriched groundwater originated from an adjacent open body of water. Particularly useful was a plot of Cl vs δ^2H for identifying groundwater capture and release zones. Similarly, isotopes of radon (Rn) have been particularly useful in understanding submarine groundwater discharges (Burnett and Dulaiova, 2003, Muligan and Charette, 2005, Sanford et al., 2007; Ono et al., 2013).

2.5.4.2 Chemical mass balance

The water balance described in Section 6.1.1 only allowed the nett groundwater contribution to be determined. Also, the water balance can be subject to large errors that arise from uncertainty associated with individual components. Coupling the water balance with balances for conservative solutes and environmental isotopes can assist in reducing the size of the errors. Using steady-state assumptions, a chemical mass balance approach allows for long-term groundwater inflows and outflows to be quantified. Such an approach was successfully employed by Sacks et al. (1998), Barr et al. (2000), Turner and Townley (2006) and others.

The chemical mass balance approach uses a naturally occurring tracer combined with the water balance equation to compute groundwater inflow and outflows. The tracer can be a conservative inorganic solute such as chloride or an isotope of the water molecule. The water balance equation presented in Section 2.5.2.1 is rewritten as follows:

$$\mathbf{(P.C_P + R_i.C_{Ri} + G_i.C_{Gi}) - (E.C_E + R_o.C_{Ro} + G_o.C_L + A.C_A) = \Delta V.C_L}$$

where

	$\Delta V.C_L$	=	mass of tracer in the lake (load)
	C_P	=	concentration of tracer in precipitation
	C_{Ri}	=	concentration of tracer in run-off (in)
	C_{Gi}	=	concentration of tracer in groundwater (in)
	C_E	=	concentration of tracer in evaporation
	C_{Ro}	=	concentration of tracer in run-off (out)
	C_L	=	concentration of tracer in lake
	C_A	=	concentration of tracer in abstracted water

2.5.4.3 Inorganic chemistry

The hydrochemical character of water is an expression of the history of that water and is controlled by climate, geology and flowpath through the environment. General properties (such as EC) and concentration of specific ions combine to define hydrochemical character.

Graphic techniques are widely used to display and interpret the chemistry of water, with the Piper diagram and Durov diagram being well known (Johnson, 1975; Hem, 1985; Tredoux, 1987). They are relatively simple and can be used without extensive knowledge of chemistry (Zaporozec, 1972).

2.5.4.4 Stable isotope tracers

Environmental isotopes are those naturally occurring isotopes of elements found in abundance in our environment (e.g. H, C, N, O and S), and now routinely contribute to hydrological and geohydrological investigations, complementing geochemistry and physical studies (Clark and Fritz, 1997). Marimuthu et al. (2005) observed isotope data often provides important supplemental information to more traditional hydraulic data. Stable isotopic composition, for example, is modified by meteoric processes, allowing for identification of waters from different parts of the hydrological system by their isotopic signatures. Similarly, radioactive decay provides a measure of circulation times and renewability. The stable isotopes used in water studies are generally of light elements with the relative mass difference between their isotopes being large e.g. Deuterium (^{2}H) has 100% more mass than ^{1}H. This allows for measurable fractionations during physical and chemical reactions.

Clark and Fritz (1997, pg 2) succinctly describe the structure of a nuclide. "The nuclear structure of a nuclide (an isotope-specific atom) is classically defined by its number of protons (Z) which defines the element, and the number of neutrons (N) which defines the isotope of that element. For a given nuclide, the sum of protons and neutrons gives the atomic weight (A), expressed by the notation $^{A}_{Z}Nu_{N}$. For example, most oxygen has 8 protons and 8 neutrons, giving a nuclide with 16 atomic mass units ($^{16}_{8}O_{8}$) while about 0.2% of oxygen has 10 neutrons ($^{18}_{8}O_{10}$). In reality, the mass of a nuclide is slightly less than the combined mass of its neutrons and protons. Conventional notation for a nuclide uses only the elemental symbol and atomic weight (e.g. ^{18}O or ^{34}S)." Unstable isotopes or radioactive nuclides have a probability of decay, but stable isotopes do not spontaneously disintegrate.

Stable isotope concentrations are expressed as the difference between the measured ratio of a sample and the reference or standard ratio divided by the measured ratio of the reference:

$$\delta^{18}O = \left(\frac{\left(\frac{^{18}O}{^{16}O}\right)_{sample}}{\left(\frac{^{18}O}{^{16}O}\right)_{standard}} - 1 \right) * 1000\ \text{‰}$$

The standard used is the Vienna Standard Mean Ocean Water (VSMOW). **δ**-values are expressed as parts per thousand or permil (‰).

Evaporation and cycles of condensation significantly change the isotopic composition of water through fractionation. Lighter isotopes evaporate first, resulting in the remaining water becoming enriched with heavier isotopes i.e. the $\delta^{18}O$ value increases. Given the composition of water, deuterium (^{2}H or D) and oxygen-18 (^{18}O) are two stable isotopes widely used in hydrological studies. Water with an isotopic composition falling on the meteoric water line is assumed to have originated from the atmosphere and to be unaffected by other isotopic processes. Deviation from the meteoric water line indicates evaporation from open water, exchange with rock minerals, condensation and other processes (Domenico and Schwartz, 1997).

The use of stable isotopes in groundwater studies is well documented, and range from studying recharge (Bredenkamp et al., 1995), baseflow separation (Saayman et al., 2003) and leakage of water pipes (Verhagen and Butler, 1995). In regard of groundwater – lake interaction, Turner et al. (2000) used stable isotope data to confirm groundwater flow patterns through Lake Jasper, a shallow coastal freshwater lake in the southwest of Western Australia. The lake has an open-water surface area of 4.5 km^2 and an average depth of 3 m. Water in the lake is enriched with heavy isotopes of deuterium and oxygen-18, allowing for the lake release zone (discharge zone) to be mapped by comparison of groundwater isotopic compositions to lake compositions. They found the capture zone to be much larger than the release zone.

2.5.4.5 Radon

Radon (^{222}Rn) is a naturally occurring radioactive nuclide that has a half-life of 3.8 days. The noble gas is an indirect decay product of uranium or thorium natural decay series, and colourless, odourless and tasteless. ^{222}Rn is soluble in water, relatively easy to measure and behaves conservatively in coastal ocean waters (Dulaiova et al., 2006). Radium-226 (^{226}Ra) – from which ^{222}Rn is produced – is ubiquitous in rocks, soils and sediments, and hence found in virtually every aquifer matrix (Schmidt et al., 2008). In contrast to groundwater, surface water lacks major contact with radon-emanating mineral material.

The use of ^{222}Rn as a tracer has not been widely applied in South Africa. Hobbs et al. (2010) recognised the tool has potential, but did not yield conclusive results in their applications. Internationally, ^{222}Rn has proved to be a particularly useful tracer in the study of surface – groundwater interaction and is widely used in the study of submarine groundwater discharge (Burnett and Dulaiova, 2003, Mulligan and Charette, 2005, Sanford et al., 2007). Dulaiova et al. (2006) determined a seaward flux of groundwater discharge of 6 m^3/d per metre width of coastline of West Neck Bay, New York. A similar flux was determined from a sandy unconfined aquifer on Cape Cod Massachusetts by Mulligan and Charette (2005) using a variety of methods, including radon.

The use of ^{222}Rn as a tracer has also been applied to groundwater – lake studies. Schmidt et al. (2008) reported the radon method generally allowed reliable quantitative estimation of groundwater discharge rates into small lakes in Germany, even with a few samples. Ono et al. (2013) used a continuous measurement device to identify groundwater discharge into Lake Ezu in Japan, with the results being verified by divers conducting visual inspections.

2.6 Recharge

Accurate estimation of recharge remains one of the biggest challenges for groundwater investigators (Weaver and Talma, 2005). The addition of precipitation via the vadose zone to the aquifer is a complex process influenced by rainfall duration and intensity, antecedent moisture conditions, soil and geology, hydraulic conductivity and depth to water, amongst others. The topic has been widely studied both internationally (Simmers, 1987; Gieske, 1992;

Scanlon et al., 2002, Healy, 2010 and others) and locally (Bredenkamp et al., 1995; Xu and Beekman, 2003a; DWAF, 2006; van Wyk, 2010 and others). The role of recharge in understanding sustainability is also hotly debated (Bredehoeft, 2002, Devlin and Sophocleous, 2005). From the perspective of this research, a reasonable assessment of recharge is required to assess whether enough groundwater is available to satisfy the groundwater input into Groenvlei predicted in Section 6. All methods used to estimate recharge have limitations, prompting van Tonder and Xu (2000) to advocate an integrated approach. The availability of data and time are key issues that control which methods can be used. Six methods are potentially available with which to estimate recharge to the Eden Primary Aquifer:

- Estimates of recharge to other coastal primary aquifers;
- Chloride Mass Balance (CMB) method;
- Water balance methods (EARTH model; CRD method; RIB method and SVF method).

2.6.1 Recharge estimates at other primary aquifers

Most major groundwater studies address recharge to one degree or another, with estimates for South African primary aquifers sourced from the literature summarised in Table 2.9. The estimates are based on a variety of methods, including the assumption that a fraction or percentage of mean annual precipitation infiltrates the subsurface and percolates down to the water table. Ignoring the quantum and seasonality of annual rainfall, estimates of recharge to primary aquifers along the South African coast are in the order of 20% MAP.

The concept of a rainfall threshold having to be exceeded before recharge takes place is reported in the literature (Bredenkamp et al., 1995; van Wyk, 2010) and is included in the CRD method for estimating recharge described by Xu and van Tonder (2001). Rawlins (1991) observed the primary aquifer near St Lucia is only recharged by those rainfall events that exceed 10mm. The same threshold was proposed by Owor et al. (2009) for the Upper Nile Basin and Rajendra Prasad et al. (2009) for the Narava basin, India. When rainfall is less than 10 mm, rain is either intercepted by plants or retained in the vadose zone.

Table 2.9: Estimates of recharge to South African coastal primary aquifers expressed as a percentage of mean annual precipitation.

Aquifer	Source	Recharge (% MAP)	Comments
Atlantis	Bredenkamp & Vandoolaeghe (1982)	25	Water balance method
	Dyke (1992)	15	
	Fleisher (1990)	16	
	Fleisher & Eskes (1992)	23	Hydrograph method
		42	CMB method
		23	Average, dune area 42%, vegetated areas 23%
Cape Flats	Henzen (1973)	17	
	Gerber (1981)	33	
	Vandoolaeghe (1989)	15 – 35	
	Vandoolaeghe (1990)	33	
	Cave and Weaver (2000)	25	Philippi – assumed
Eden	Vivier (2009)	45	Assumed
Lower Berg	Timmerman (1985b)	15	
	Baron (1990)	10	
	Du Toit & Weaver (1995)	8	Water level data near Saldanha Steel
	Du Toit & Weaver (1995)	12	Modelling for Saldanha Steel
	Weaver & Talma (2005)	12	Langebaan Road, CMB method
Zululand	Worthington (1978)	24	
	Meyer & Kruger (1987)	21	
	Bredenkamp et al. (1992)	12 - 15	
	Meyer & Godfrey (1995)	18	CMB method
	Meyer et al. (2001)	19	
	Vœret et al. (2007)	20 – 40	
	Dennis & Dennis (2009)	8.5	Coastal area north St Lucia

2.6.2 Chloride Mass Balance method

The CMB method was described by Dettinger (1989), Bredenkamp et al. (1995), Wood (1999), van Tonder and Xu (2000), Weaver and Talma (2005), Healy (2010), van Wyk (2010) and others. It can be applied to both the unsaturated zone and the saturated zone, depending whether Cl concentration soil profile data or groundwater quality data are available. The steady-state method is attractive in that it is simple to use, but this is balanced by a lack of information on Cl concentration in rainfall, the difficulty of accurately measuring Cl at such low concentrations, and determining a representative Cl concentration in groundwater.

The method was proposed by Eriksson and Khunakasem in 1969, and is founded on Cl being a conservative tracer generally not absorbed or desorbed during transport through the soil. It is assumed all chloride in groundwater is provided by rainfall and there is no recycling or concentration of chloride in the aquifer (Wood, 1999). Also, the chloride mass balance flux must have remained constant over time. Recharge is calculated using the following equation:

$$\mathbf{R_e = (Cl_r / Cl_{gr}) \times 100}$$

where	R_e	=	recharge expressed as % MAP
	Cl_r	=	Cl concentration in rain
	Cl_{gr}	=	Cl concentration in groundwater

Because of high Cl concentrations at the coast, seasonal fluctuations and dry season deposition, van Wyk et al. (2011) warned application of the CMB method in coastal regions should be done with caution. Seasonality of Cl concentration in rain, spatial variation and the ability of laboratories to accurately measure such low Cl concentrations are concerning factors raised by van Tonder and Bean (2003).

Guan et al. (2010) (quoting Eriksson, 1959, 1960) observed about 10% of total chloride in the sea salt aerosols moves into the continents, and the majority of this chloride is deposited within 100 km of the coastal area. They also observed it is the bulk chloride deposition that is of interest, including both wet deposition (with rainfall) and dry deposition.

Where applied with discretion in South Africa, reasonable results have been obtained. Recharge to the Atlantis Aquifer was determined to be 12.5% MAP while that in the vicinity of St Lucia ranged between 22% and 37% MAP (Bredenkamp et al., 1995). Meyer and Godfrey (1995) reported a slightly lower estimate of 18% for the St Lucia area. Weaver and Talma (2005) estimated recharge in the Struisbaai area to be between 9.7% and 13.5% MAP.

Some data pertaining to the Cl concentration of rain is presented by Weaver et al. (1999), Weaver and Bean (2005), Weaver and Talma (2005) and van Wyk (2010). Measurements at Struisbaai – located on the coast 285 km west of Groenvlei and in the same rainfall region – between September 1999 and September 2000 indicated the Cl concentration in rain to be 25.3 mg/L. This compared well to earlier measurements using two collectors installed by Weaver et al. (1999). A Cl concentration of 14 mg/L was reported at George by Weaver and Bean (2005), but being 10 km inland from the coast and at an elevation of 200 mamsl this station would have been subjected to rainout and continental effect. Extrapolating the Weaver and Bean data back to the coast indicates a Cl concentration of about 20 mg/L. This estimate is slightly higher than that reported by Guan et al. (2010) for coastal sites near Adelaide, Australia that experience a similar rainfall.

2.6.3 Water balance methods

A suite of water balance methods used to estimate recharge was included in the Excel-based RECHARGE software package (van Tonder and Xu, 2000). Using monthly rainfall, groundwater level and groundwater abstraction data with specific yield allows recharge to be determined by comparing observed groundwater levels with simulated levels. The accuracy of these methods is largely dependent on the quantity and quality of data available for interpretation (van Tonder and Bean, 2003). It is generally assumed recharge is uniform over the model area.

EARTH model [1]

Introduced by Van der Lee and Gehrels (1990) (as quoted by Gieske, 1992) the EARTH model is a physically based lumped distributed model used for simulating groundwater level fluctuations by coupling climatic, soil moisture and groundwater level data. The storage coefficient or specific yield of the saturated zone needs to be known for recharge to be estimated. The general equation used in the model is as follows (van Tonder and Xu, 2000):

[1] Acronym for "Extended model for Aquifer Recharge and soil moisture Transport through the unsaturated Hardrock"

$$\mathbf{S\ dh/dt = (R - h) / DR}$$

where	R	=	recharge
	S	=	specific yield
	dh/dt	=	change in water level head during one month
	DR	=	drainage resistance (a site specific parameter)
	H	=	groundwater level

Cumulative Rainfall Departure (CRD) method

The CRD method was first introduced more than 70 years ago and has been described by Bredenkamp et al. (1995), van Tonder and Xu (2000) and Xu and Beekman (2003a). It is based on the observation that groundwater levels correspond with patterns of rainfall. The method is based on the rationale that equilibrium exists between the average rainfall and the hydrological response (run-off, recharge, evapotranspiration and other losses). The CRD method can also be applied when natural conditions no longer prevail because of pumping, which induces a response of groundwater level. The method has proved to be particularly applicable in dolomitic aquifers, but can be used in other geohydrological environments. Beekman and Xu (2003a) reported the CRD method to be relatively accurate, inexpensive and easy to use. Xu and van Tonder (2001) found it to be a simple but powerful tool with which to estimate recharge:

$$R_T = rCRD_i = S_y\,[\Delta h_i + (Q_{pi} + Q_{outi}) / A\,S_y)] \quad \textit{with}$$

$$CRD_i = \sum_{i=1}^{N} R_i - \left(2 - \frac{1}{R_{av} i} \sum_{i=1}^{N} R_i\right) iR_t$$

where	r	=	fraction of a CRD that contributes to recharge
	S_y	=	specific yield
	Δh_i	=	water level during month i
	Q_p	=	groundwater abstraction
	Q_{out}	=	natural outflow
	A	=	recharge area

R_i = rainfall for month i

R_t = threshold value representing aquifer boundary conditions

R_t may range from 0 to R_{av}, with 0 representing a closed aquifer with no outflow and R_{av} representing an open aquifer with some outflow e.g. spring discharge. The method can only be applied where groundwater level fluctuations are observed. It is best suited to those areas with a small storativity (e.g. fractured aquifers) and sensitive to recharge with rainfall. The accuracy of recharge estimation decreases with increasing depth to water table.

Rainfall Infiltration Breakthrough (RIB) model

The RIB model grew from the CRD method (Xu and van Tonder, 2001, Xu and Beekman, 2003b) and was modified by Sun et al. (2013) to consider both monthly and daily data. Sun et al. (2013) described the mathematics of the model and the use of the Excel-based interface. They found the method best suited to assessments based on monthly data and shallow unconfined aquifers with relatively low transmissivity. They also found estimates of recharge sensitive to estimates of specific yield.

Saturated Volume Fluctuation (SVF) method

Also known as the water-table fluctuation method, this water balance approach considers a change in storage (as represented by a change in groundwater level) against recharge and abstraction. It is assumed the base of the aquifer is impervious, preventing losses or inflow through the base. If evapotranspiration losses are relevant, they can be included in the abstraction term (Bredenkamp et al., 1995). Lateral inflow and outflow can be assumed to be equal, and thus not sometimes not used in the water balance. The SVF method can be applied to a single borehole or to an area with a number of boreholes with monitored groundwater level data; however having water level fluctuation data representative of the aquifer as a whole is a key issue when applying this method. The general equation for the SVF method is:

$$\mathbf{R + (I\text{-}O) - Q = S\ \Delta V}$$

where	R	=	recharge
	(I-O)	=	groundwater inflow less outflow
	Q	=	groundwater abstraction (including evapotranspiration)
	S	=	specific yield
	ΔV	=	change in saturated volume

If two points on a hydrograph are selected that have the same level, then ΔV is equal to zero and the general equation can be rewritten as:

$$\mathbf{R + (I\text{-}O) = Q}$$

This is known as the Equal Volume method. Assuming groundwater inflow to be equivalent to outflow, the equation can be further simplified to show recharge is equivalent to abstraction.

3 METHODOLOGY

3.1 Preamble

The research presented in this thesis is based on data and information gathered during the past decade from groundwater consulting appointments in and around Groenvlei. This approach prevented implementation of a focused research effort, but rather relied on geohydrological data collected during the course of various projects. Only the collection of stable isotope and radon samples by Parsons and Bugan (2008) had a research focus i.e. evaluating radon (^{222}Rn) as a tool for assessing the groundwater contribution to wetlands.

An extensive literature review was undertaken as part of this research effort, with the following key objectives:

- Collect all available information pertaining to the hydrology of Groenvlei;
- Collect available information pertaining to similar systems elsewhere;
- Understand the state-of-the-art regarding groundwater – lake interaction; and
- Research tools that could be used to quantify the role of groundwater in sustaining Groenvlei.

Chapter 2 of this thesis is an outcome of the literature review.

3.2 Collection of Data for Water Balance

The water balance equation for Groenvlei dictated climatic (rainfall, evaporation) and lake water level data be collected (Appendix B). Rainfall and evaporation have been monitored on a daily basis by DWA at station K3E003 since 1976. Evaporation is measured using a S pan (Symons's Pan or S class pan). This station is located at Rondevlei some 10 km west of Groenvlei, with both locations having similar physiographic settings. Lake levels are also monitored by DWA at KR4001, established in 1977.

3.2.1 Rainfall data

Inspection of the rainfall record indicated data were missing. For example, a rainfall event of almost 80 mm was recorded at K4E001 on 13 June 1983, but no rainfall was recorded at Station K3E003 less than 5 km away. Similarly, rapid rises in water levels in Groenvlei are sometimes not accompanied by any measured rainfall (e.g. 2 Dec. 1993, 21 Apr. 1994). Instances were also found where high rainfall did not reflect in the water level of the lake (1 Jan. 2007).

Rainfall has also been monitored by the Knysna Municipality at the Sedgefield fire station since 1997. This data was sourced from the municipality to confirm the rainfall data collected from K3E003. Daily rainfall data from the two stations were plotted on Figure 3.1. The data displayed poor correlation (R^2 = 0.27). Rainfall on 24 July 2011, for example, caused the water level in Groenvlei to rise 100 mm, suggesting a significant event. 144 mm was measured at the Fire Station, but only 53 mm at K3E0003. The lack of correlation is further exemplified by the number of times rainfall is recorded at the Fire Station, but not at K3E003.

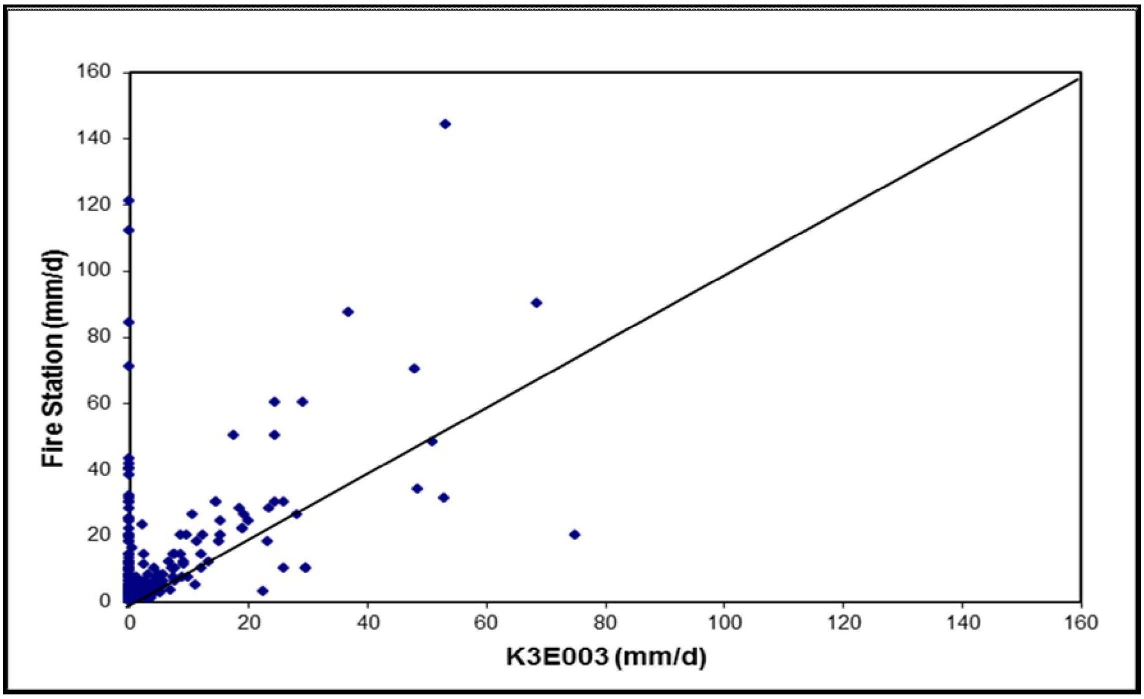

Figure 3.1: A comparison of rainfall measured at K3E003 and the Sedgefield fire station between 1 January 2009 and 31 December 2012.

Further comparison of the two data sets indicated annual differences ranging between 3.7% and 98%. Excluding data measured during 2011 and 2012 which had differences of 98% and 63% respectively, the median difference in the two annual rainfall data sets was 14.1%. The differences could arise from either different amounts of rainfall at the two locations and / or errors in measurement and data storage. It is far more difficult to identify missing rainfall data than it is to identify missing evaporation data, but noting about 17% of the evaporation record was missing (see below) and the differences in rainfall measured at Rondevlei and the fire station, it is apparent the rainfall record introduces a degree of uncertainty and possible error into the water balance.

3.2.2 Evaporation data

Analysis of the evaporation data from K3E003 indicated about 17% of the record measured between 1981 and 2012 was missing. These gaps are identifiable from Figure 3.2. Because evaporation is relatively predictable, it was possible to patch the record using average evaporation of each month. The monthly distribution of evaporation from zone 24B presented in the WR90 data set (Midgley et al., 1994) could have been used to do this (Table 3.1). However, it was decided to determine the average evaporation for each month using the measured S pan data from K3E003.

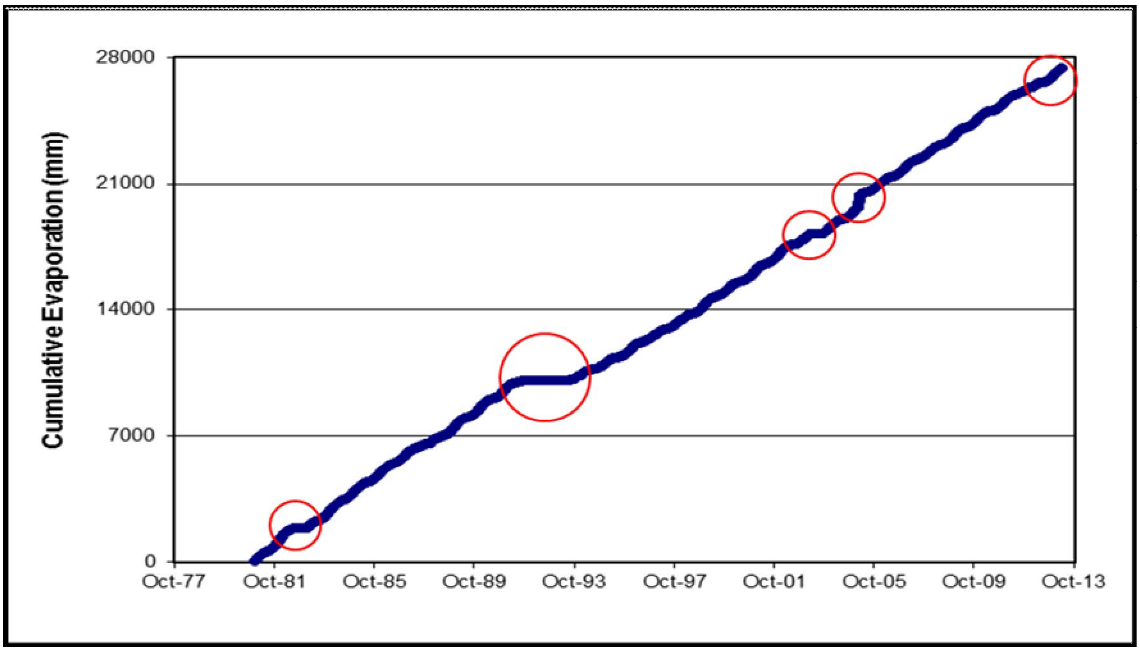

Figure 3.2: Identification of periods of missing data in the evapotranspiration record of K3E003

Table 3.1: Monthly S pan evaporation determined using the distribution provided by Midgley et al. (1994).

Month	Monthly Distribution (%)	(mm/d)
Oct	9.21	3.9
Nov	10.91	4.7
Dec	13.29	5.7
Jan	12.96	5.5
Feb	10.16	4.3
Mar	8.77	3.8
Apr	6.59	2.8
May	5.41	2.3
Jun	4.95	2.1
Jul	5.06	2.2
Aug	5.92	2.5
Sep	6.77	2.9

In determining monthly averages, 32 daily measurements of evaporation (i.e. 0.2% of the record) were interpreted to be erroneously high, ranging between 10.0 mm/d to 55.7 mm/d. Nineteen of these measurements were recorded in March 2005. Clulow et al. (2012), working on a coastal wetland in Maputoland (i.e. a similar setting as that of Groenvlei), measured maximum evaporation rates of 6.0 mm/d. Inspection of the data from K3E003 showed few peak summer measurements were in excess 7.0 mm/d. Consequently, the 32 records were removed and the evaporation data set corrected using the monthly averages presented in Table 3.2.

Table 3.2: Monthly S pan evaporation measured at K3E003.

Month	Average (mm/d)	Std deviation (mm/d)	Min (mm/d)	Max (mm/d)	Values above 10	Missing data (%)
Oct	3.04	1.04	0.0	7.6	0	16
Nov	3.75	1.10	0.0	7.7	0	18
Dec	4.34	1.33	0.0	9.0	4	23
Jan	4.34	1.29	0.0	8.8	4	19
Feb	3.84	1.17	0.0	8.7	4	12
Mar	3.05	1.14	0.0	10.0	19	14
Apr	2.23	0.83	0.0	7.0	1	19
May	1.59	0.76	0.0	5.6	0	17
Jun	1.36	0.64	0.0	4.7	0	26
Jul	1.28	0.66	0.0	4.3	0	17
Aug	1.70	0.70	0.0	5.0	0	18
Sept	2.25	0.87	0.2	6.9	0	18

Using the patched data, average S pan evaporation between 1981 and 2012 was determined to be 1 028 mm/a [1]. This is lower than the 1 326 mm/a given in WR90 for K3E003, but similar to the 1 096 mm/a given for the nearby, but no longer functioning weather station at Swartvlei. These two weather stations were 2.5 km apart. Fijen (1995) presented a 14 year average of 1 187 mm/a. Considering the 17% missing record and the variance of estimates of annual evaporation presented by Fijen (1995), Midgeley et al. (1994) and this research, the evaporation data is likely to add further uncertainty to that introduced by the rainfall data.

3.2.3 Groenvlei water level and water chemistry data

Monitoring station K4R001 is managed and maintained by DWA. It was established in 1977 at 34.03444°S and 22.85194°E and is used to monitor both water level and chemistry of Groenvlei. Two loggers are used at all times to monitor the water level of Groenvlei (van Wyk, *pers.comm.*, 2013). At times, a third logger is also used. During each monthly site visit, data are checked and if the logger data differs by more than 5 mm from the manual plate measurement, then the loggers are recalibrated. Comparison of the data collected by the two loggers and comparison with the manual water level measurements are presented in Figures 3.3, 3.4 and 3.5.

Plate 3.1: A researcher at the K4R001 gauging station at Groenvlei (Photo: E De Roeck).

[1] The synthetically generated S pan record amounted to 996 mm/a.

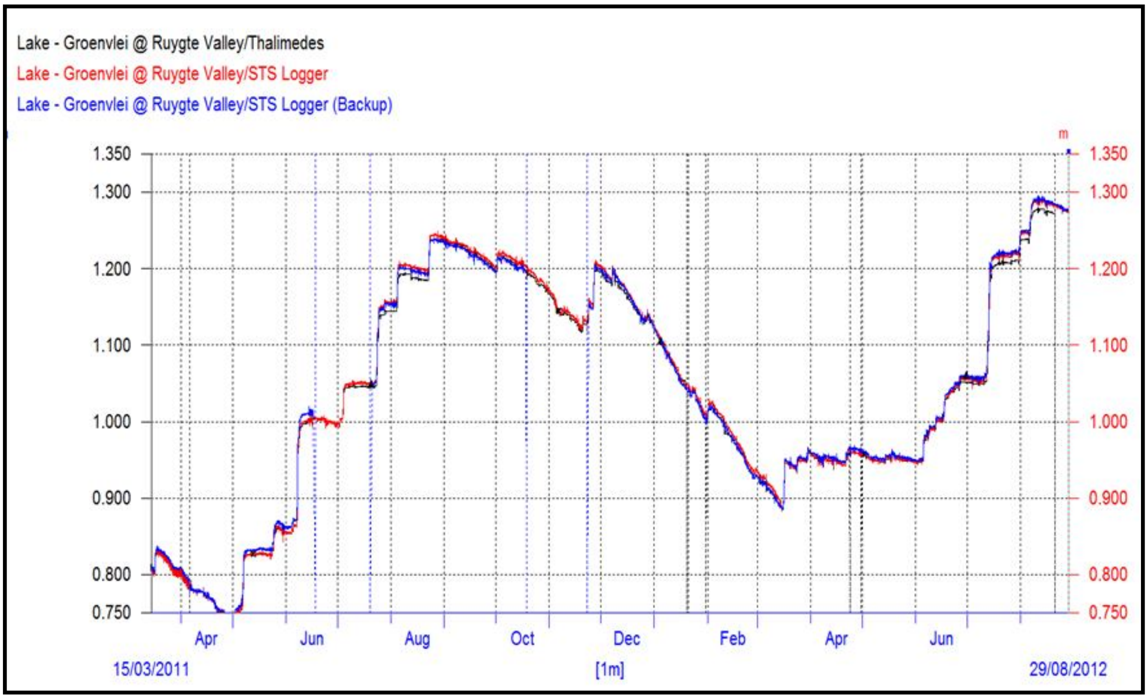

Figure 3.3: Comparison of water levels monitored at Groenvlei using three different loggers. Graph provided by Neil van Wyk, DWA George Office, October 2013.

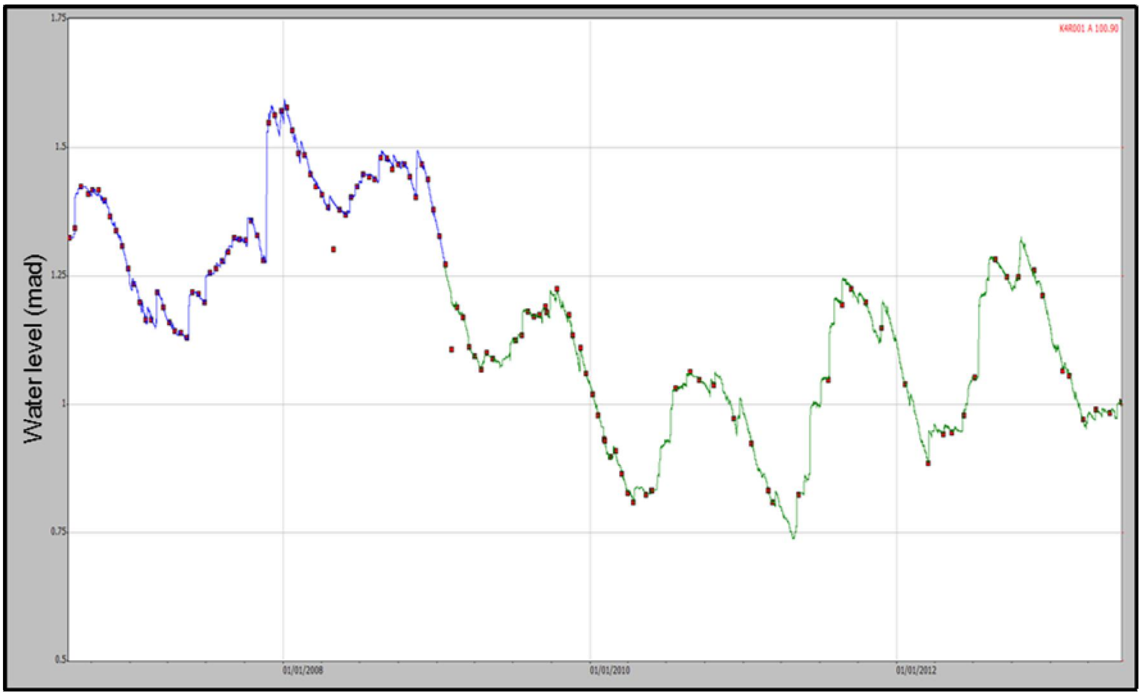

Figure 3.4: Comparison of water levels monitored at Groenvlei using data loggers (blue and green lines) and control water levels recorded at the gauge plate (red squares). Graph provided by Neil van Wyk, DWA George Office, October 2013.

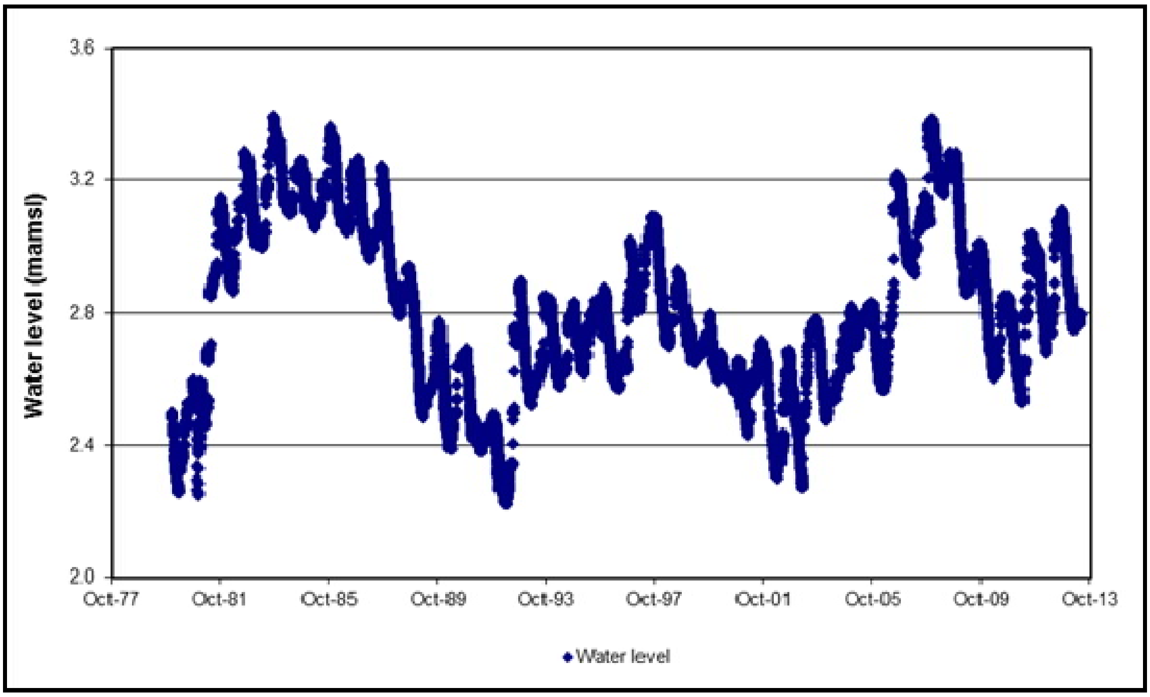

Figure 3.5: Graph of water level of Groenvlei produced using data provided by DWA.

3.3 Groundwater Data

A WISH [1] database was established to store groundwater data collected during the consultancy projects (Appendix B). Of particular importance was data pertaining to borehole location, elevation, groundwater level and groundwater chemistry. Pumping test data and groundwater level monitoring data were sourced from specific projects, and are referenced in the text. The location of boreholes in the immediate vicinity of the lake are shown in Figure 3.6.

Private boreholes drilled in the vicinity of Groenvlei (DOD1, GRU1, GRU2, PGB1) were drilled for water supply purposes. Depth of drilling and borehole construction was guided by this consideration and not by research objectives. This resulted, for example, in the thickness of the sand in the vicinity of the lake not being established. Typically the boreholes were drilled about 20 m below the water table. The boreholes were drilled at a diameter of about 250 mm, and fitted with 125 mm / 144 mm PVC screen and casing. The screen extended

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[1] Windows Information System for Hydrogeologists (WISH) was developed at the IGS, Bloemfontein (Lukas, 2012). It evolved from the National Groundwater Database (NGDB) and Hydrocom.

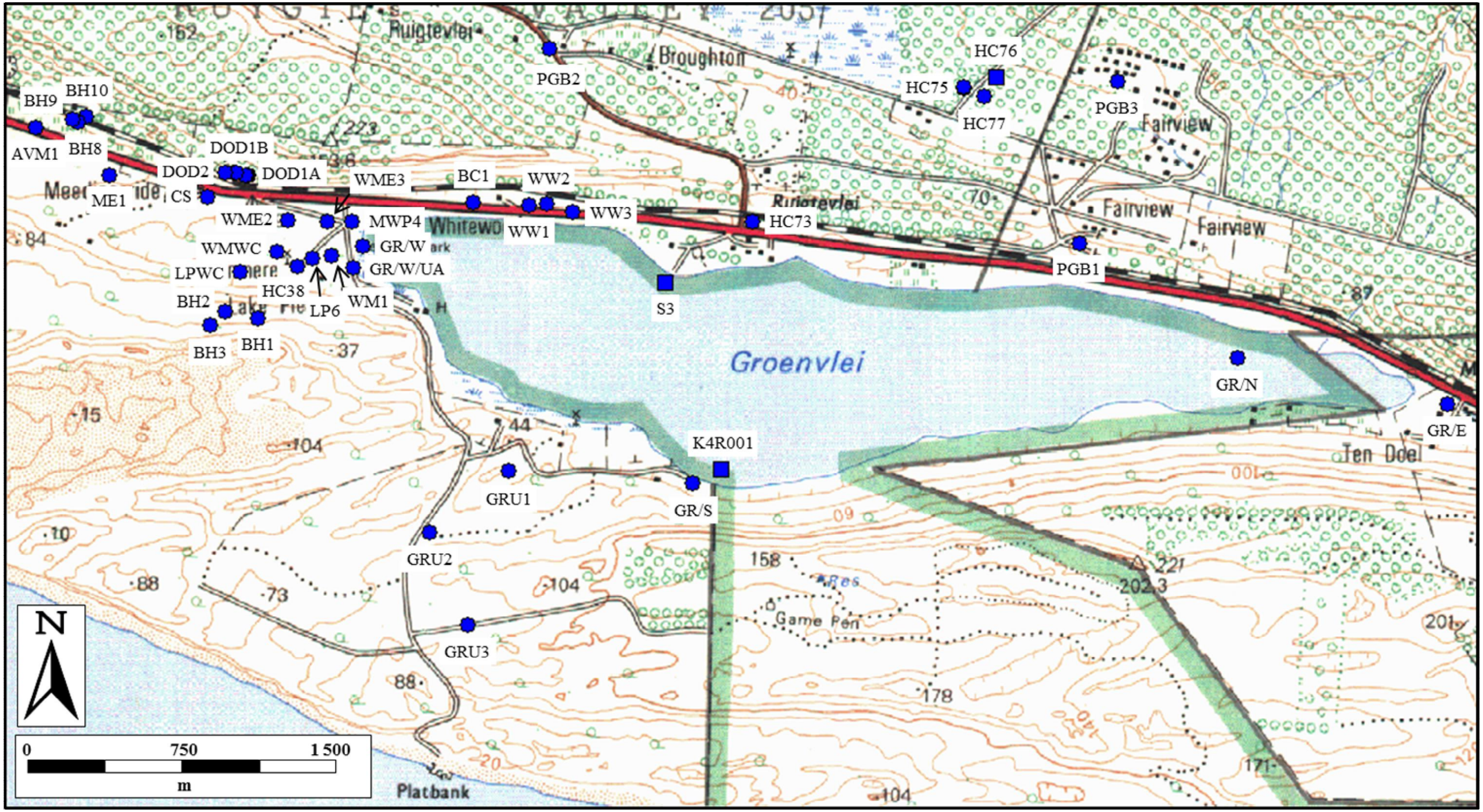

Figure 3.6: Boreholes located in the immediate vicinity of Groenvlei.

from some 6 m above the water table to the bottom of the borehole, and the annulus was backfilled with 16 / 30 graded sand. The boreholes were initially developed by airflushing and later by pumping.

Groundwater samples were routinely collected during groundwater investigations of the area, mainly for the purpose of assessing water potability. Samples were submitted to laboratories for analysis, which typically included major anions and cations, EC and pH. The analytical results were added to the WISH database. EC field measurements were regularly taken during pumping tests to monitor any groundwater quality changes induced by pumping.

3.4 Isotope study

Parsons and Bugan (2008) collected water samples from Groenvlei and groundwater for stable isotope and ^{222}Rn analysis with a view to testing ^{222}Rn as a tool in surface – groundwater interaction studies. Water samples were taken from the lake and from groundwater, with the expectation that differences between the two waters would allow discharge into and from the lake to be detected. Of particular importance was investigating the stratified hydrochemistry at GRU1 (see Section 5.7.2).

Analysis of the ^{222}Rn data showed groundwater samples collected using a bailer yielded low radon concentrations, suggesting the groundwater to have been resident in the borehole for sufficient time for the radon levels to have decayed. Using an average of 0.8 Bq/L in bailed samples from DOD1B and GRU1, it was calculated groundwater was in the borehole and with limited contact with the sediments for 11 days. Consequently bailed groundwater ^{222}Rn samples were excluded from the analysis of the data.

3.5 Quantification Process

Once all the required data had been collected, reviewed and edited, a daily water balance model for Groenvlei was developed (Section 6.1). The output of the model was the nett groundwater contribution to Groenvlei. Initial work resulted in good results being obtained for summer, but not for winter. Realisation that the open water and vegetative fringe behaved differently in winter yielded better results.

Using steady-state calculations based on Darcy's Law, it was possible to disaggregate nett groundwater contribution into groundwater inflows and outflows. This resulted in all of the terms of the water balance being known, and allowed for a review of the evaporation data so evaporation from open water and transpiration losses from the vegetative fringe could be quantified using S pan data.

A chemical mass balance model was then developed based on a water balance model coupled to hydrochemical data. The quality of groundwater was defined using data stored in the database, while the Cl concentration of rain and evaporated water was based on the literature. The quality of water leaving the lake through discharge into the aquifer was set using the monitored lake quality data. This iterative process allowed for refinement of the input parameters such that a better quantification of the role of groundwater in sustaining Groenvlei was achieved.

4 DESCRIPTION OF GROENVLEI

4.1 Preamble

Groenvlei, literally translated from Afrikaans as “Green Lake”, is an isolated shallow eutrophic coastal lake surrounded by reeds with small beaches on the southern, western and northern shores. Jones and Day (2003) classified the water body as a permanently indurated endorheic basin that changes only slightly from season to season. Groenvlei is somewhat unique in South Africa in that it is a relatively large permanent open body of water (359 ha) not fed by river inflow or connected to the sea. This means it is not subject to sedimentation, channel diversion and other anthropogenically-induced impacts that influence and control floodplain wetlands. The lake is fed by direct rainfall and groundwater inflow, offset by evaporation losses from the lake surface, transpiration losses from vegetation in and peripheral to the water body, and subsurface discharge along the southern shores.

4.2 Climate

Climatically, the southern Cape is transitional between the winter rainfall Mediterranean climate of the Western Cape and the subtropical summer rainfall characteristic of much of the rest of the country. Groenvlei experiences a mild and temperate climate. Average daily maximum temperatures range between 23.8^{o}C (February) and 18.2^{o}C (August). In the same months, average daily minimum temperatures are 19.7^{o}C and 8.0^{o}C respectively.

The measured mean annual rainfall at Groenvlei amounted to 653 mm/a (1981 – 2012) [1] (Figure 4.1), most of which is orographic in nature. Thunderstorms are rare. Rain falls throughout the year, with average monthly rainfall ranging between 37 mm/month (May) and 68 mm/month (October) (Figure 4.2). The bimodal rainfall pattern described by Allanson (2006) is not apparent. However, rainfall in the area can be variable, with both floods and droughts being periodically experienced. In the past decade, for example, Sedgefield

[1] Rainfall data from 1991, 1992 and 1993 were removed from the rainfall record used to determine this average as both evaporation and Groenvlei water level data indicated missing rainfall data during this period.

experienced below-average rainfall for four consecutive years (2008 – 2011), while floods were experienced in March 2003 (175 mm in 3 days), December 2004 (166 mm in 2 days), August 2006 (275 mm in 2 days) and November 2007 (257 mm in 2 days).

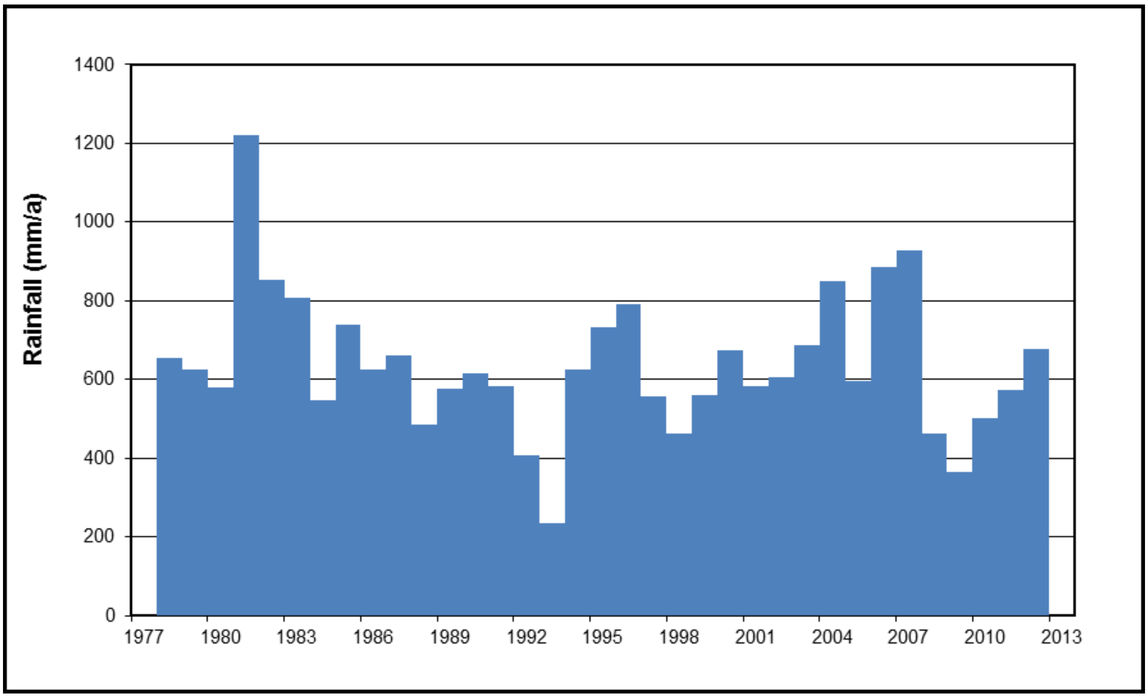

Figure 4.1: Annual rainfall measured at Station K3E003, Rondevlei, between 1977 and 2013.

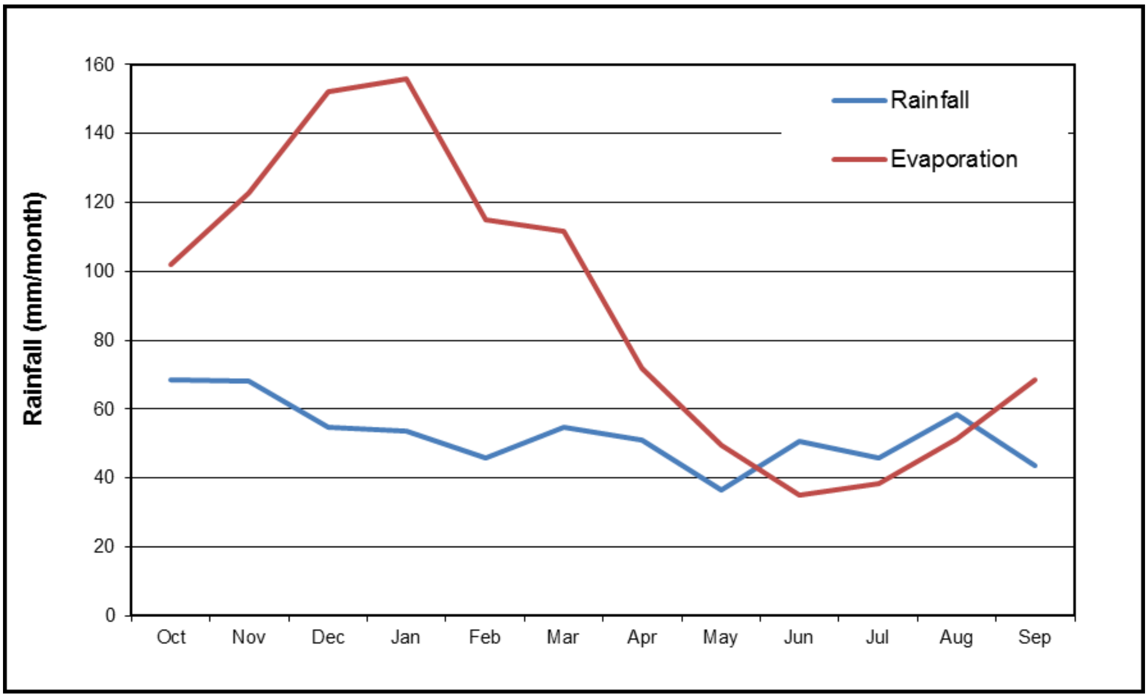

Figure 4.2: Average monthly rainfall and S pan evaporation measured at Station K3E003, Rondevlei.

Mean annual S pan evaporation measured at station K3E003 - located 10 km north west of Groenvlei - amounted to 1 028 mm/a (1981 – 2012 using patched data). Monthly S pan evaporation ranged from 175 mm/month in summer (December, January) to 65 mm/month (June, July) (Figure 4.2).

Winds are mainly coast-parallel, with east to south-east winds being dominant in summer. Westerly to north-westerly winds dominate in winter. Strong winds are not common, but Coetzee (1980, quoting Manley, 1972) reported a mean annual calm of 34.2% for George. Fijen (1995) noted dry bergwinds can cause a severe lowering in humidity and increased fire hazards, particularly in winter.

Age dating using various isotopes, stratigraphic investigation and pollen by Holmgren et al. (2003), Marker and Holmes (2005), Chase et al. (2009) and others has shown climatic conditions changed significantly in Southern Africa during the Holocene Epoch (last 11.7 ka). In addition to the end of the last glacial period 10 ka, relatively humid conditions prevailed in southwestern Africa between 8.7 ka – 7.5 ka, 6.9 ka – 6.7 ka, 5.6 ka – 4.9 ka and 4.2 – 3.5 ka. A period of marked aridity occurred between 3.5 ka and 0.3 ka. Initially, conditions were quite warm before the onset of mid-Holocene cooling, accompanied by variable but mostly dry conditions. The Little Ice Age covered the four centuries between AD 1500 and 1800, with temperatures steadily rising since then. These works, together with observations in recent rainfall records, highlight climatic changes and variability in the past 10 ka, and the degree of uncertainty when trying to model the water balance of Groenvlei in this time span.

4.3 Geology

4.3.1 General description

The geology and geological history of South Africa is well documented in Truswell (1977), Johnson et al. (2006) and others. Generally these works provide little detail on recent coastal sand deposits characteristic of the study area.

Marker and Holmes (2005) reported that the southern Cape shares its geological foundation with Gondwana. Deposition of the Kaaimans Group and intrusion of the George granite occurred during the later stages of the Proterozoic era (610 – 540 Ma) (Figure 4.3, Table 4.1).

This was followed by the deposition of the Cape Supergroup and the early part of the Karoo Supergroup during the Paleozoic era (540 – 254 Ma). The Cape Orogeny occurred during the early stages of the Mesozoic era (230 – 210 Ma), forming the prominent Cape Fold Belt. The breakup of Gondwana during the Jurassic period (180 – 135 Ma) and planation that occurred during the Cretaceous period (135 – 65 Ma) were instrumental in shaping the coast and morphology of southern Cape. The subsequent Cenozoic era saw continuing erosion and moulding of the present day landscape. This era included widespread laterisation, two major periods of uplift (18 Ma and 2.5 Ma) and significant fluctuations of sea level. The prominent dunes in the study area were formed at this time. The rifting responsible for the breakup of Gondwana continues at a rate of about 2 cm/yr.

Calcified sands formed the aeolianite observed at the mouth of Swartvlei and east thereof. The three dune cordons have been aged at 120 – 150 ka (seaward cordon), 120 – 230 ka (middle cordon) and more than 500 ka (landward cordon) (Bateman et al, 2011). These are covered by sands which have no dune structure, are entirely decalcified and comprise rounded, medium to fine-grained sand (Marker and Holmes, 2005). The cover sands are significantly younger than the aeolianite, ranging between 0.5 ka and 3.1 ka, and were deposited after the mid-Holocene high seawater level. Tinley (1985) considered dunes directly west of Groenvlei as the best example of compound imbricate parabolic dunes in South Africa.

The lake is located on unconsolidated aeolian sands of late Pleistocene and Holocene age (Figure 4.3). Little information is available regarding the thickness of the sands or the nature of the underlying geology, with neither Birch et al. (1978) nor Martin and Fleming (1986) providing insight into depth to bedrock [2]. Drilling of boreholes directly west and south of Groenvlei revealed the sand extends to at least -10 mamsl (Parsons, 1997, 2006c, 2009a), thereby indicating the aquifer to be at least 15 m thick. Data presented by Bateman et al. (2011) indicates the aeolianite extends to depths of at least -35 mamsl. Boreholes drilled in the car park at Myoli Beach – 3 km west of Groenvlei – revealed the unconsolidated sand extends to -25 mamsl and the aeolianite extends to at least -63 mamsl (Parsons, 2011). Given

2 Extrapolation of pre-Quaternary basement levels in Fig.10b of Martin and Fleming (1986) indicated sands could be in the order of 50 m thick, extending to an elevation of -50 mamsl. However, this observation must be treated with caution.

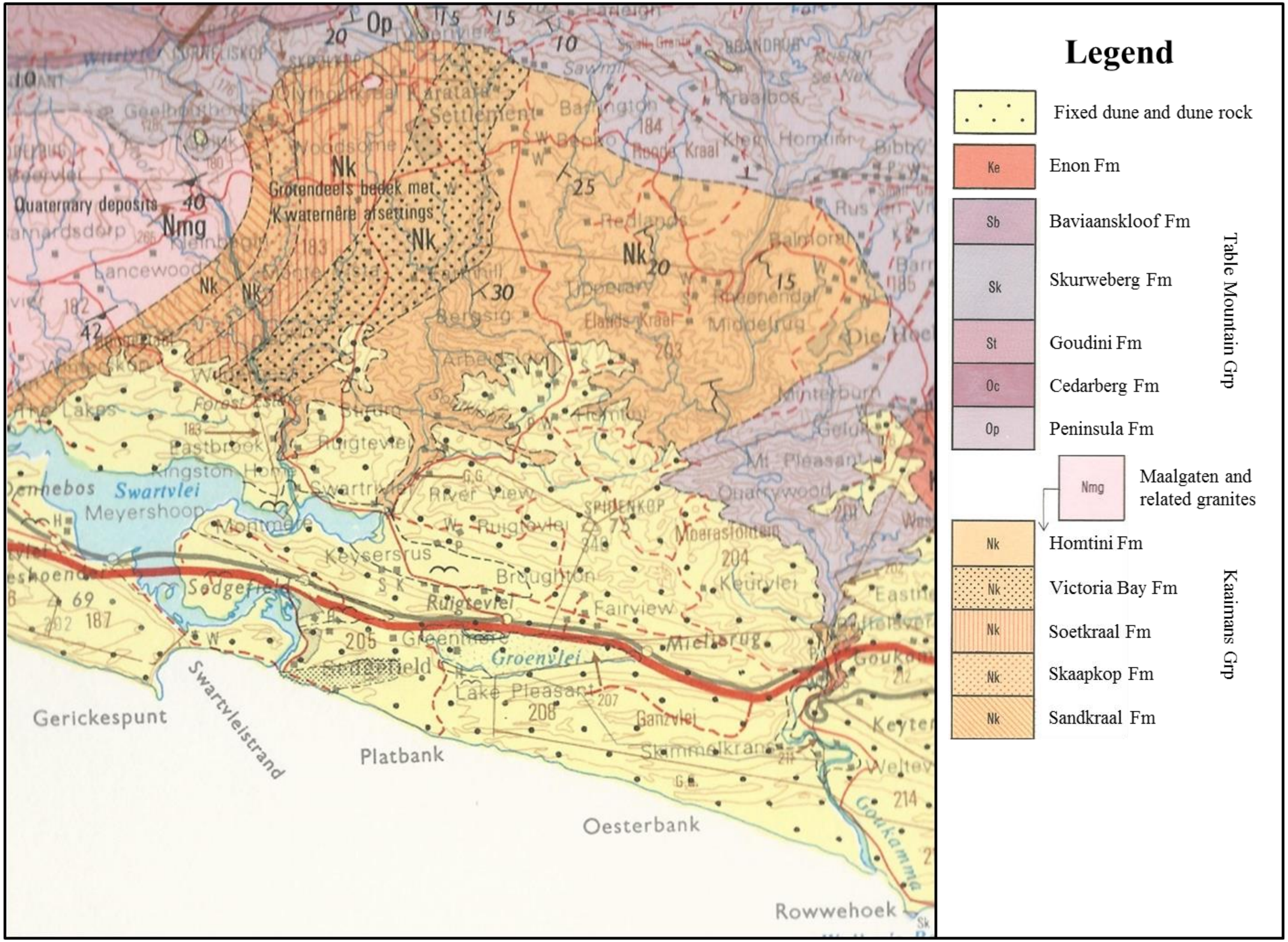

Figure 4.3: Lithostratigraphy of the study area (from Coetzee, 1979).

Table 4.1: Abridged geological succession of the study area.

Era		Geological epoch/period (ages in Ma)			Geological group, formation, etc.	Dominant rock or material type
CENOZOIC	Quaternary	Holocene	0.01	Algoa Group	Schelm Hoek Fm*	Unconsolidated sand dunes
		Pleistocene	1.8		Nahoon Fm	Cemented dune rock
					Salnova Fm	Calcareous sandstone, beach deposits
	Tertiary	Pliocene	5		Nanaga Fm	Calcareous sandstone with aeolian cross-bedding
		Miocene	23		Alexandria Fm	Limestone, calcareous sandstone, beach deposits
		Oligocene	34			
		Eocene	56		Bathurst Fm	Limestone
		Palaeocene	65		Grahamstown Fm	Silcrete
MESOZOIC		Cretaceous	145		Mzamba Formation	Marine siltstone
				Uitenhage Group	Sundays River Formation	Marine mudstone, siltstone
					Kirkwood Formation	Fluvial sandstone, mudstone
					Enon Formation	Conglomerate
					Mngazana Formation	Conglomerate
		Jurassic	200	Karoo Supergroup	Drakensberg Group	Basalt, dolerite
		Triassic	250		Clarens, Elliot and Molteno Formations	Sandstone, siltstone, mudstone
					Beaufort Group	Shale, mudstone
PALAEOZOIC		Permian	300		Ecca Group	Shale
		Carboniferous	360		Dwyka Group	Tillite (diamictite), shale
		Devonian	416	Cape Supergroup	Witteberg Group	Sandstone, shale
					Bokkeveld Group	Shale, sandstone
		Silurian	444		Table Mountain Group	Sandstone
		Ordovician	495			
		Cambrian	545		Cape Granite Suite	Granite
		Late Precambrian	800		Kaaimans and Gamtoos Groups	Quartzite, quartz-schist, phyllite, limestone

* Fm = Formation

(Maud, 2008)

the geological history of the area – particularly that related to sea level regression and erosion during the Quaternary period – it is unwise to extrapolate sand (and aquifer) thickness from Myoli Beach to Groenvlei.

Based on the 1 : 250 000 scale regional geological maps of the area (Coetzee, 1979) [1], basement geology is expected to comprise sandstone and quartzite of the Peninsular Formation of the Table Mountain Group (Figure 4.3). These rocks are of Ordovician age (495 – 443 Ma). The contact with older shale and quartzite of the Kaaimans Group is interpreted to be in the vicinity of the western edge of Groenvlei.

4.3.2 Formation of Groenvlei

Groenvlei started to form during the last glacial period in the late Pleistocene. As sea levels started to fall some 120 ka in response to the onset of the last period of glaciation (Figure 4.4), rivers extended into the newly exposed coastal areas and cut deep valleys into them (Martin, 1959, Hill, 1974, Norman and Whitfield, 2006). This reached a peak 17 ka when the sea level was 130 m lower than present (Miller, 1990; Ramsay, 1997; Davies, 2007), referred to as the Last Glacial Maximum.

Sea levels then rose relatively rapidly with the onset of global warming and melting of the ice caps and reached levels similar to current levels about 7 ka years ago (Figure 4.4). This rapid rise in sea levels resulted in the landward transgression of the shoreline and drowning of the eroded valleys. The maximum height of sea level above present levels since the last glacial period was about 2.5 m, reached 5 ka (Miller, 1990; Marker and Holmes, 2005). After a series of smaller rises and falls in sea levels, the level of the sea receded to its present level about 4 ka years ago.

Wind-blown sand deposits covered the area between Groenvlei and the sea, effectively covering evidence of the lake's earlier connectivity to the sea and Swartvlei. Some debate exists whether Groenvlei owes its origin to the rise of sea level and drowning of a valley or to a dune barrier, as proposed by Allanson (2006). Stratigraphic evidence indicated Groenvlei was fresh 8 ka and peat had deposited over lake mud (Marker and Holmes, 2005). Forest expansion was recorded at 6.9 ka, suggesting a wetter climate.

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1 The geological map of the area uses previous nomenclature of units of the Table Mountain Group, while current nomenclature of Johnson et al. (2006) is used in this thesis.

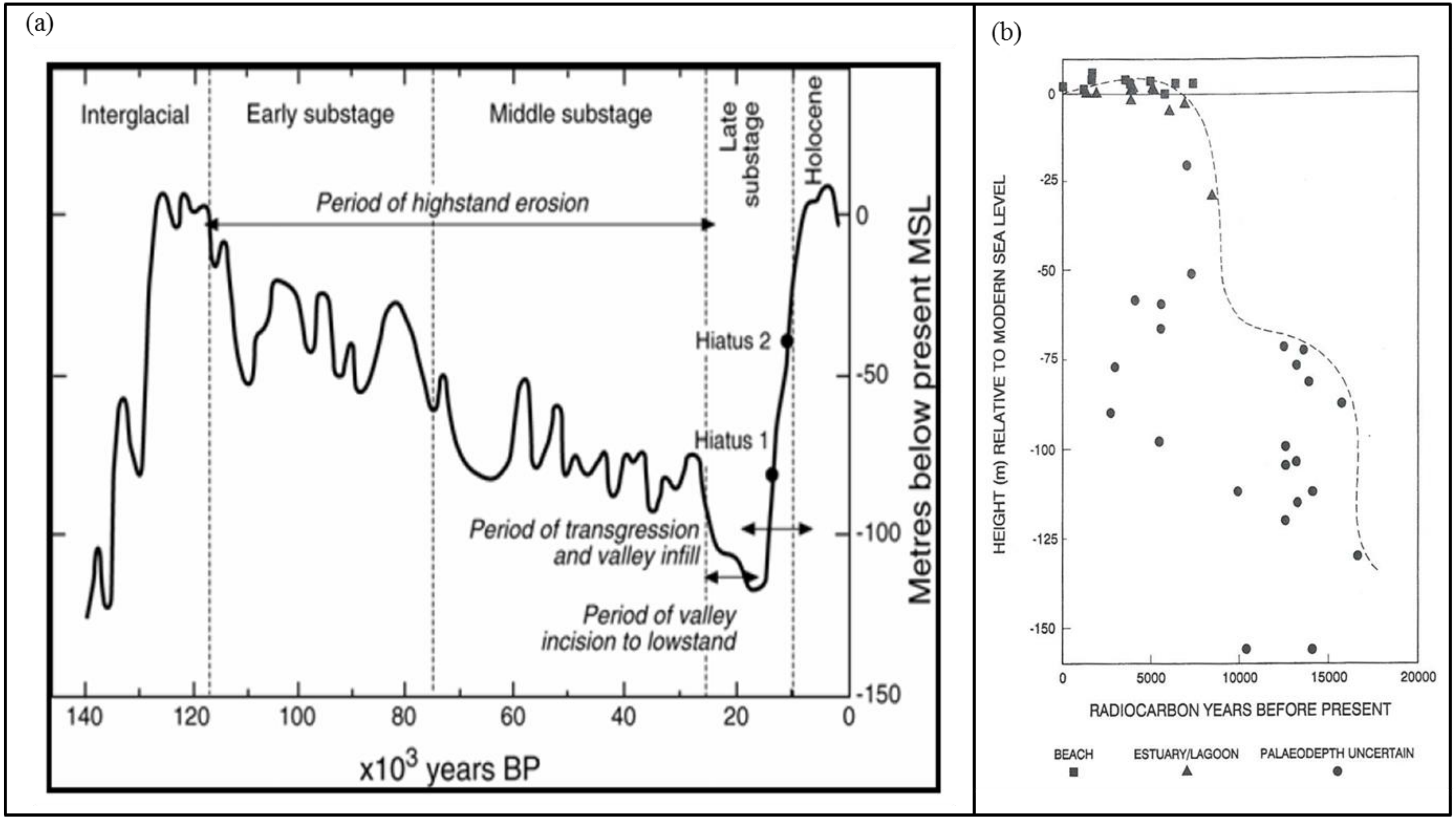

Figure 4.4: Changes in sea level during the Quaternary period presented by (a) Davies (2007) and (b) Miller (1990).

The Holocene high sea level appears to have caused marine incursion into the lake, with Groenvlei then being cut off from the sea as the sea level fell. Cover sands directly west of Groenvlei were dated by Bateman et al. (2011) to range between 3.1 ka and 0.6 ka. Between 2.6 ka and 1.4 ka, forests diminished and were replaced by grass and shrubs, indicating drier and perhaps cooler conditions. Forest expansion occurred again after 1.4 ka.

Age dating of sediments in the vicinity of Groenvlei supports the evolutionary model described above. Based on sediment dynamic calculations, Illenberger (1996) tentatively set the age of the seaward dune cordon at 6 ka and the middle dune cordon at between 200 ka and 500 ka. The landward dune cordon was calculated at being older than 600 ka. Using pollen and charcoal dating techniques, Duncan (2006) aged sands at the edge of Groenvlei at 7.4 ka while Kirsten (2008) used radiocarbon and luminescence dating of diatoms to date lake sediments at about 4 ka, with a thin layer of younger sediments (0.7 – 0.3 ka) overlying these. Irving and Meadow (1997) used radiocarbon dating to date the sediment-filled depression of Vankervelsvlei. Ages ranged from 3.2 ka at the top of the profile to almost 40 ka some 0.8 m below ground level.

The estuarine character of an early Groenvlei is supported by the presence of relict estuarine fauna in the lake (Coetzee, 1980; Hart, 1995; Allanson, 2006). Presence of the isopod *Pseudoshaperoma barnardi*, amphiphod *Grandidierella lignorum*, fish Cape silverside *Atherina breviceps* and estuarine round-herring *Gilchrestella estuaria* – suggesting a mixing of freshwater and estuarine elements –is put forward as proof of the estuarine origin of the lake. Interestingly, the South African National Biodiversity Institute (SANBI, 2011) classified Groenvlei as an estuarine wetland.

4.4 Physiography

The study area is located between the west – east trending Outeniqua Mountains in the north and the sea. The physiography of the study area is dominated by the three west-east trending dune cordons described in Section 4.3.1, with Groenvlei located in the dune slack between the seaward and middle dune cordons (Figure 2.1). The morphology of the area is illustrated in Figures 4.5 to 4.7. The seaward dune cordon attains a maximum elevation of 207 mamsl with a steep north-facing slope. The water level in Groenvlei is in the order of 3 mamsl.

Figure 4.5: Contours draped over a Google Earth image of the study area, illustrating the topography near Groenvlei.

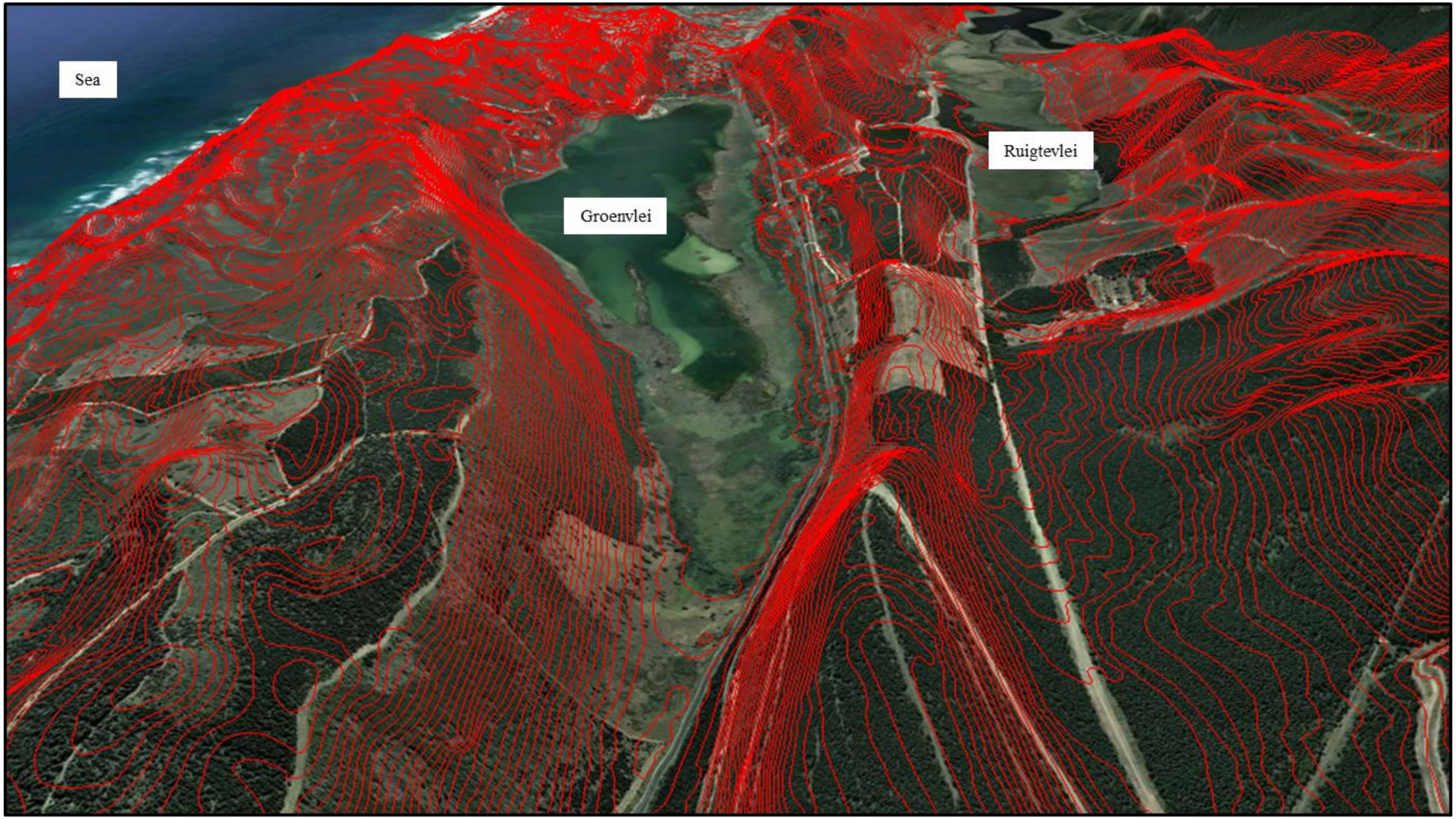

Figure 4.6: Westward view of contours draped over a Google Earth image of the study area. The position of the lake between the seaward and middle dune cordon is well illustrated, as is the proximity to the now-dry Ruigtevlei.

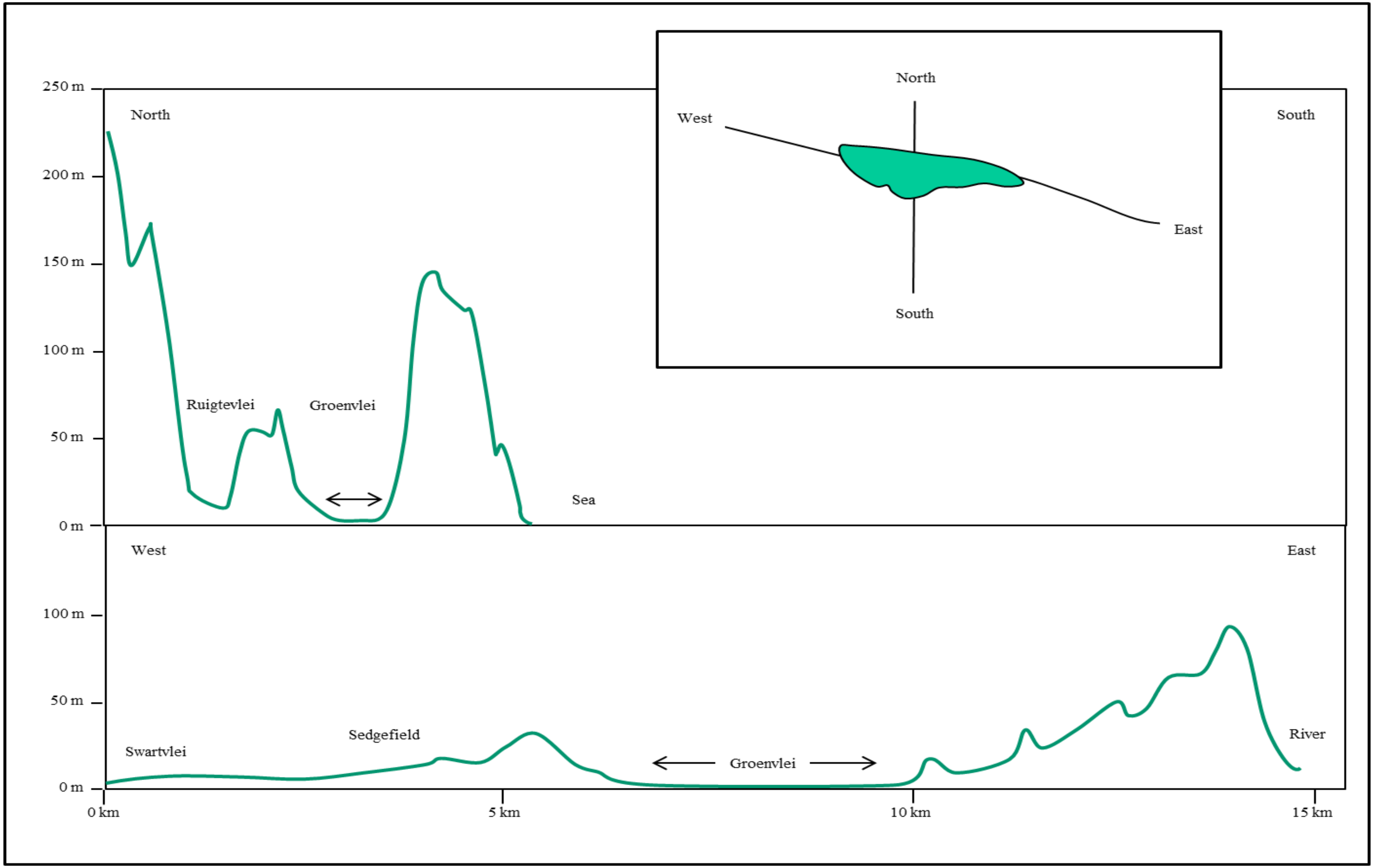

Figure 4.7: North – south and west – east profiles of the study area (vertical exaggeration ~25).

The wetland was delineated by the outer reed fringe (Figure 4.8), with topography being considered in the easternmost corner of the lake where the change in vegetation was less marked. Previously, Fijen (1995) only considered the open water body, but Parsons (2008a) assumed the wetland to include both the open water and reed collar as Allanson (2006, personal communication) was of the opinion significant water losses occur through evapotranspiration from the wetlands immediately surrounding the lake. Such an approach is supported by the literature (Section 2.5.2.6). Groenvlei has a west – east elongated shape, being some 4.8 km long and 1.2 km wide. The surface area of the open water body is 2.45 km^2 while the surrounding vegetation in and peripheral to the water body covers a further 1.14 km^2. The total area of the lake is thus 3.59 km^2, with a perimeter of 10.66 km. The estimated volume of water stored in Groenvlei is 9.1 Mm^3.

Martin (1960a) observed a platform along the northern shore some 0.7 m higher than the average water level of the lake. This, together with wetland at the eastern end of the vlei, indicates the lake was previously larger than presently the case.

In the absence of any influent rivers and the vegetated character of the surroundings, it is expected little sedimentation has taken place in the lake during the past 60 yrs. The maximum depth of Groenvlei is about 5.6 m, but much of the lake is less than 3.7 m deep. The reed fringe is generally less than 1 m deep.

4.5 Hydrology

Groenvlei is located in the Gouritz Water Management Area and in quaternary catchment K40D. The study area is bounded by Swartvlei in the west and the Goukamma River in the east. Ruigtervlei, described by Bateman et al. (2011) as a former lake, is located north of the middle dune cordon while the sea is located in the south (Figure 4.9).

The absence of any discernible seasonal pattern of rainfall ensures rivers feeding the coastal lakes of the southern Cape are perennial in character and river inflows are considerably greater than direct precipitation (Hart, 1995). Groenvlei is the exception as no rivers flow into the lake, and its only source of water is direct rainfall and groundwater discharge. The absence of any rivers also means periodic flooding is not a hydrological driver, as it is in the case of other wetlands such as the Okavango Delta or Nylsvlei.

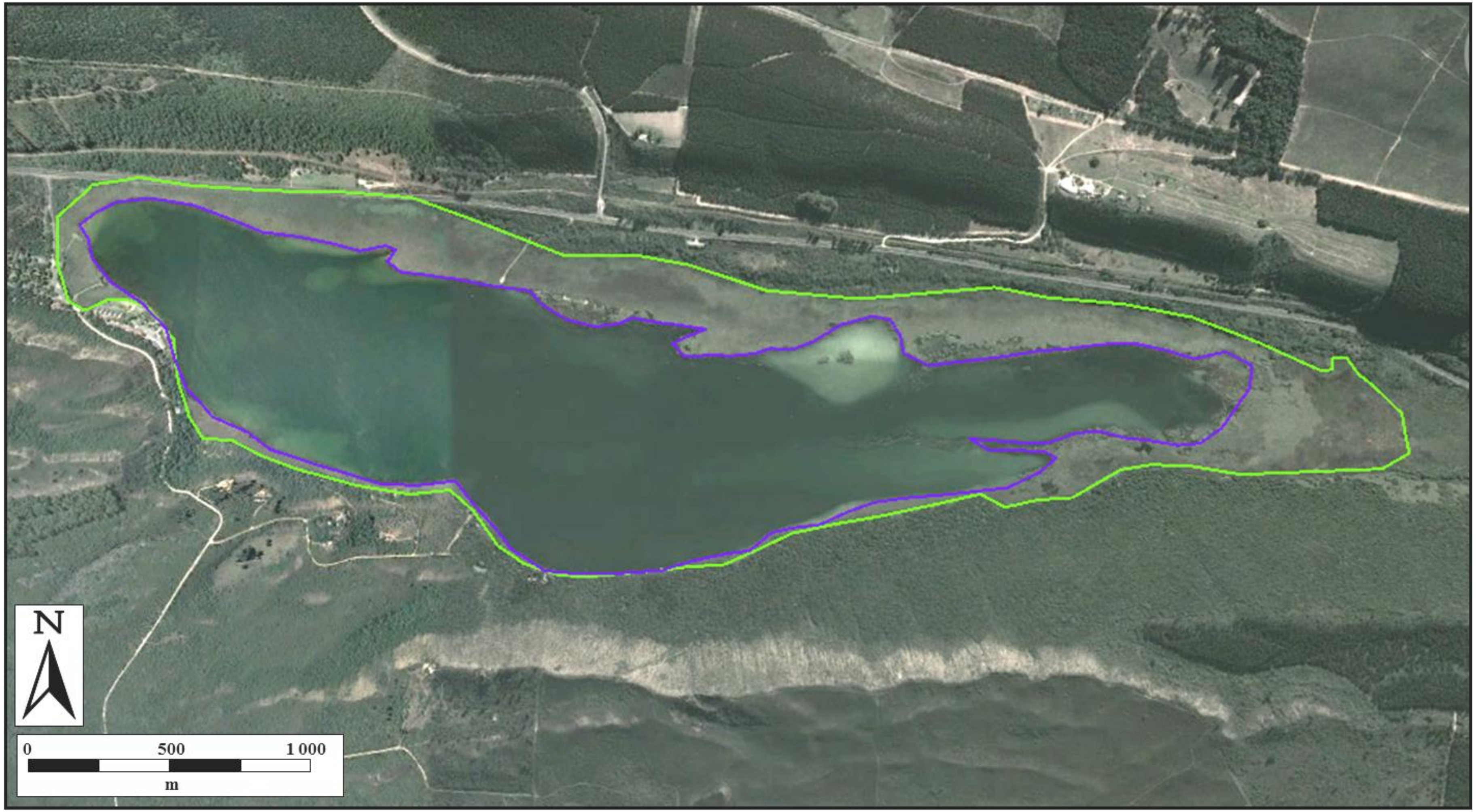

Figure 4.8: Delineation of the lake, showing the total area of Groenvlei (green line) and the extent of the open body of water (blue line).

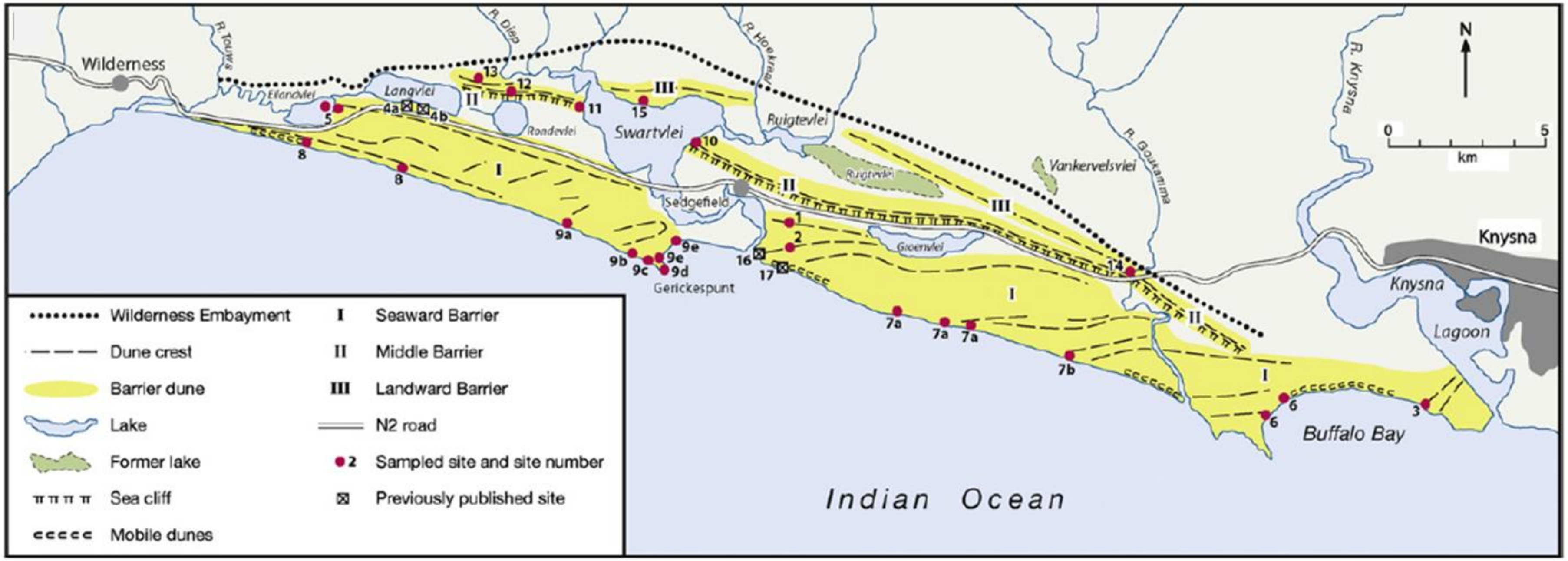

Figure 4.9: Map showing the position of coastal dunes in the Wilderness embayment, with Ruigtevlei positioned directly north of the middle dune cordon and Groenvlei directly south (from Bateman et al., 2011).

Topography in the immediate vicinity of Groenvlei slopes inward toward the lake, and the lake has no connection to the sea. The lake is bounded to the west by Sedgefield and steep, vegetated semi-consolidated dunes to the north and south. The lake is at an elevation of 3 mamsl, while the dunes peak at about 200 mamsl. Martin (1959) and Watling (1979) reported the catchment of Groenvlei to be 9.5 km^2 in extent while Fijen (1995) set the catchment area of Groenvlei at 13.8 km^2. Vivier (2009) set the catchment area at 10.5 km^2.

Martin (1960a) reported the level of the lake ranged between a maximum of 2.84 mamsl and a minimum of 1.98 mamsl. Long-term monitoring of the water level of the lake at K4R001 since 1977 showed levels range between 2.22 mamsl to 3.39 mamsl (Figure 3.5). The average water level during this time was 2.80 mamsl. The maximum range of water level was 1.17 m, while seasonal fluctuations between summer and winter are in the order of 0.3 m. Water levels decline in the summer months when evaporation losses are greater than rainfall. Levels rise in the winter months when evaporation losses are greatly reduced. However, the rise in water level cannot only be explained by the evaporation – rainfall relationship as average evaporation and rainfall during the winter months is similar (Figure 4.2). As shown in Chapter 6, the contribution of groundwater and seasonal transpiration patterns also play a role.

Long term cycles in water level are evident in Figure 4.10. Water levels are elevated directly after periods of above-average rain (e.g. 1981, 2007). Inspection of the rainfall records indicated major individual events have less control on the water level than sustained periods of above-average rainfall [1]. Periods of below-average rainfall result in the water level of the lake declining, as observed in the late 1980s and late 2000s.

Using estimates of inflows and outflows provided by Parsons (2008a), the turnover or flushing rate of Groenvlei was estimated to be once every 1.4 years. Born et al. (1979) described flushing rate as the time required to replace the equivalent of an entire lake volume.

[1] Coincidence of low annual rainfall in 1992 and 1993 and recovery of the water level is not a hydrological phenomenon, but rather reflects missing data in the rainfall record. No evaporation data was measured at station K3E003 between December 1991 and April 1993.

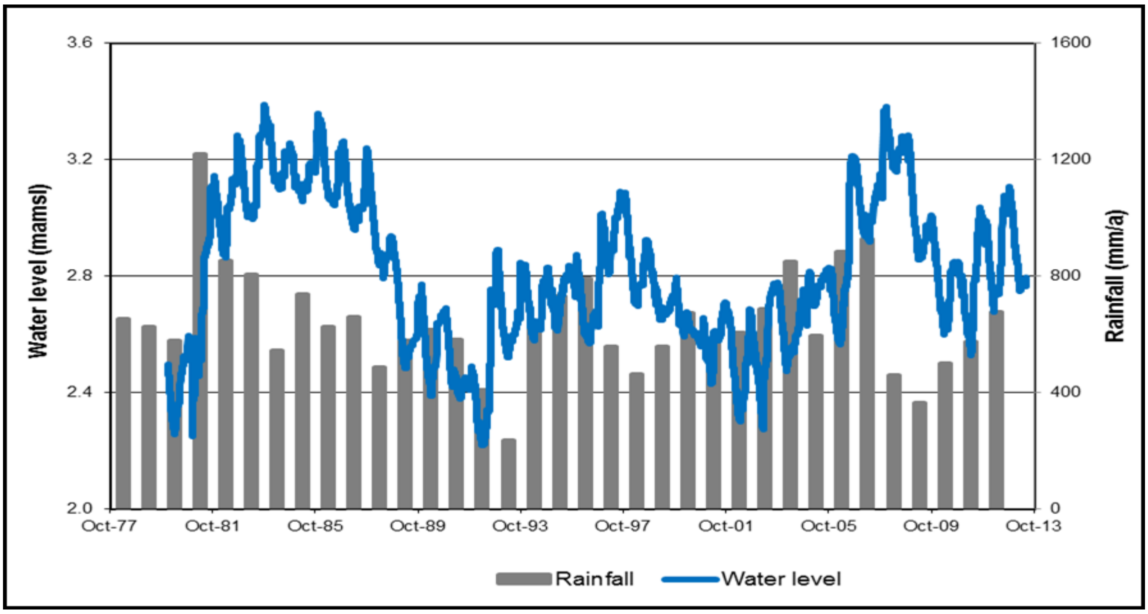

Figure 4.10: The long-term trend of water level of Groenvlei is driven by rainfall, with rises correlating to high annual rainfall. Shorter term or intra-seasonal fluctuations are also apparent, but are governed by evaporation losses.

4.6 Water Quality

Groenvlei has been described as both a freshwater and brackish lake, with confusion stemming from the salinity classification used. As Groenvlei is an inland water system and not a marine or estuarine system, it is proposed the classification for inland water systems used by Olis et al. (2009) be adopted (Table 4.2). Consequently, Groenvlei should be referred to as a brackish lake.

Table 4.2: Definition of water salinity of inland waters

Class	Salinity range (mg/L)	EC (mS/m)
Freshwater	< 2 000	300
Brackish water	2 000 – 12 000	300 – 1 800
Saline water	12 000 – 40 000	1 800 – 6 000
Hypersaline water	> 40 000	> 6 000

(Olis et al., 2009)

The water is green in colour and had a vanishing depth of a standard Secchi disk of 3.25 m in January 1951. Allanson (2006) reported Secchi disc transparency varies around 1.5 m due to both wind-induced turbulence and phytoplankton growth. He also observed the green colour of the water is governed by the wind-induced suspensions that scatter light. The water contains very little organic matter in suspension or colloidal solution. Martin (1960a) and Coetzee (1980) observed Groenvlei is shallow enough to permit complete mixing, particularly during and after strong winds. Temporary thermal stratification develops during periods of warm, still conditions.

Coetzee (1980) noted the salinity of Groenvlei had not changed noticeably since the studies of Le Roi le Riche and Hey in 1947 and Martin in 1960. Monitoring of water quality at station K4R001 since 1977 indicated EC ranged between 375 mS/m and 525 mS/m, with a median of 440 mS/m (Figure 4.11). The salinity has an inverse relationship with water level, with salinity being lowest when water levels are elevated. As the water level in the lake starts to drop the salinity starts to increase. Summary statistics of other major anions and cations are presented in Table 4.3

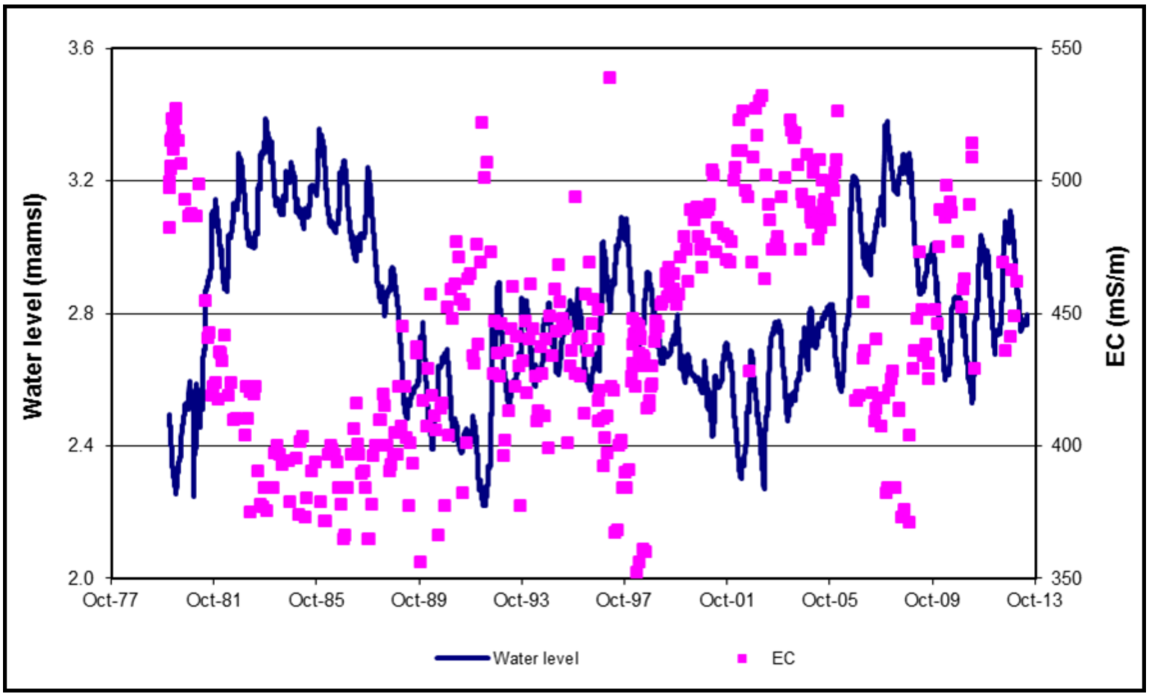

Figure 4.11: Fluctuations of the water quality of Groenvlei in relation to changes in water level.

Table 4.3: Summary of major anion and cation concentrations in Groenvlei, as measured at K3R001 between 1977 and 2013.

	EC (mS/m)	**pH**	**K (mg/L)**	**Na (mg/L)**	**Ca (mg/L)**	**Mg (mg/L)**	**SO4 (mg/L)**	**Cl (mg/L)**	**TAL (mg/L)**	**NH4-N (mg/L)**	**N0x-N (mg/L)**	**P-Tot (mg/L)**	**PO4-P (mg/L)**	**F (mg/L)**
Median	440	8.7	18.8	761.4	38.2	92.6	139.8	1227.8	297.1	0.04	0.03	0.03	0.02	0.67
Minimum	340	6.4	1.8	476.3	7.3	56.1	16.4	768.6	90.8	0.02	0.02	0.01	0.00	0.05
2nd percentile	365	7.5	13.5	609.7	23.6	72.2	90.8	1009.4	192.7	0.02	0.02	0.01	0.00	0.48
98th percentile	524	9.6	27.2	926.1	51.7	117.6	196.0	1606.8	356.5	0.14	0.30	0.08	0.08	0.91
Maximum	539	10.0	100.5	1051.8	187.2	132.5	546.8	1833.6	440.0	0.29	0.86	0.30	1.43	1.62
Samples	415	417	414	409	413	414	410	413	417	414	411	72	354	393

The lake water has a distinct NaCl character and different from that sampled from boreholes around the lake (Figure 4.12). The pH of the lake water varies between 7.3 and 10.0, while that of the groundwater falls in a narrower range of 6.3 and 8.3. These differences are useful in that they allow different waters to be "fingerprinted" and identified.

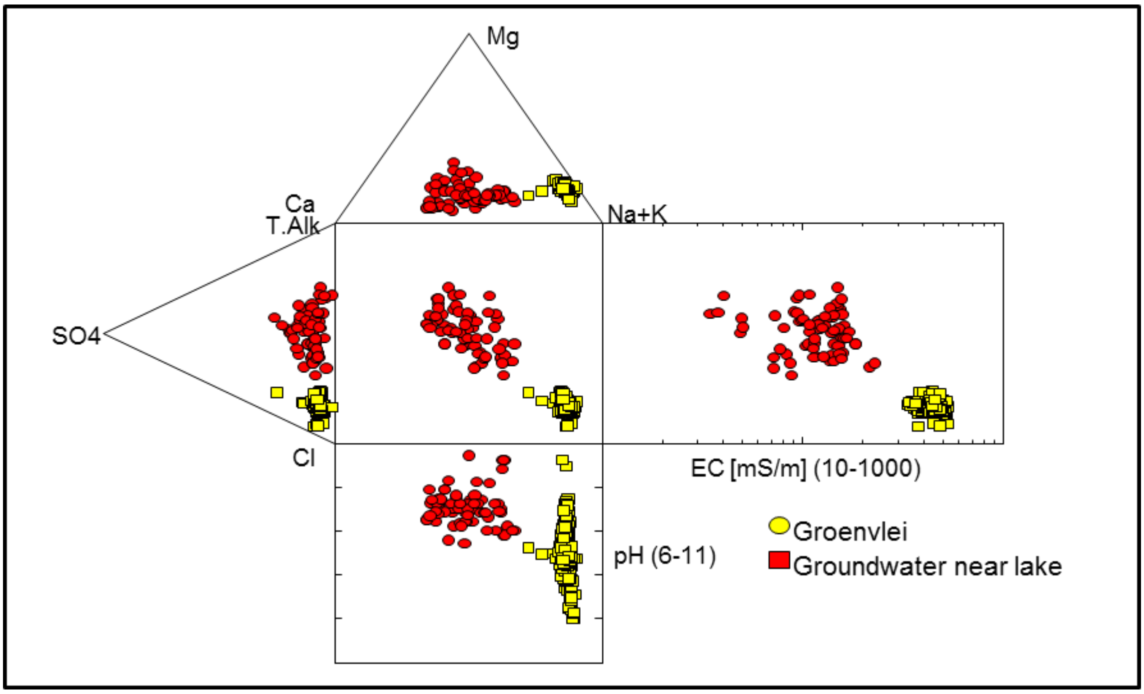

Figure 4.12: Durov diagram showing the hydrochemical character of water monitored at Groenvlei and groundwater sampled in the near vicinity of the lake.

In his three-dimensional mapping of Groenvlei, and using a handheld EC meter, Roets (2008) found EC to range between 444 mS/m and 470 mS/m, a range of 5.5%. He argued the lowest EC values recorded at the eastern end of the lake supported the folk belief of the existence of a spring in the area. An alternative possibility is there was incomplete mixing of groundwater and lake water in the shallow eastern extremities on the day of measurement.

Both Watling (1979) and Coetzee (1980) found significant differences in the water quality of Groenvlei and the other Wilderness Lakes, with differences being attributed to the isolated character of Groenvlei and the river and estuarine inflows into the other lakes. Sampling of both water and sediments in Groenvlei led Watling (1979) to conclude the lake is unpolluted.

The water quality of Groenvlei (440 mS/m) is distinctly different from that of Lake Sibaya (60 mS/m), in spite of both having similar hydrological drivers. Using water quality data collected between July 1988 and February 1990 (Meyer and Godfrey, 2003) it is apparent Lake Sibaya water chemistry has a mixed Cl–Alk character when compared to the NaCl dominated Groenvlei water (Figure 4.13). Reasons for Groenvlei having brackish water are discussed in Section 6.3.

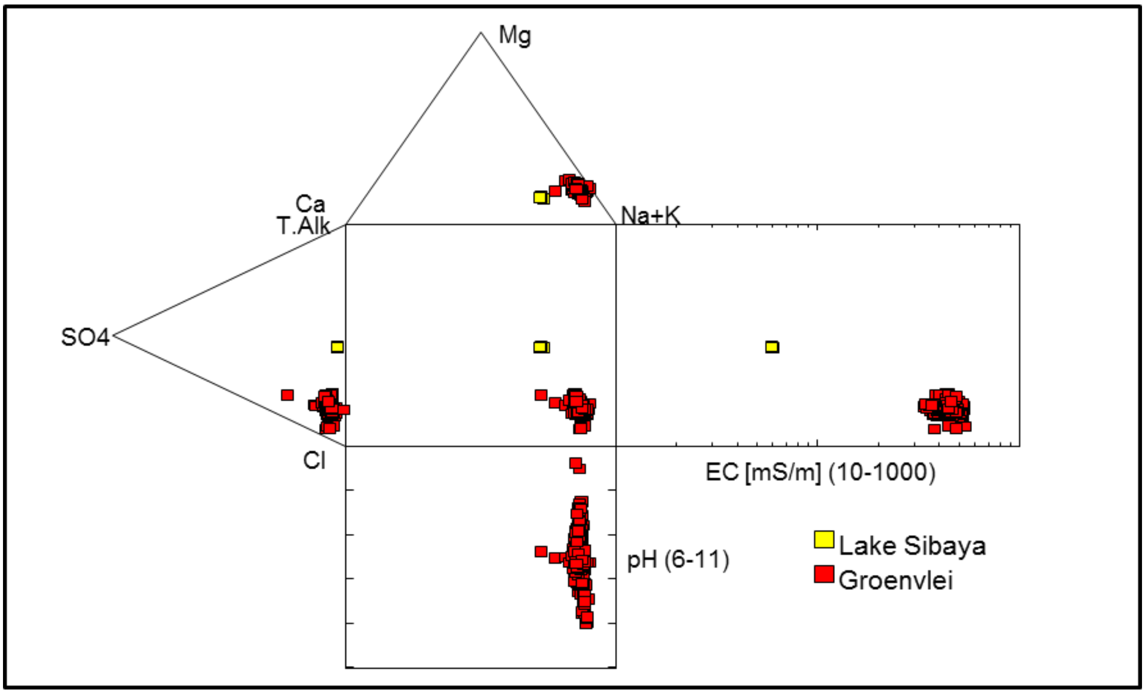

Figure 4.13: Comparison of the hydrochemical character of Groenvlei and Lake Sibaya.

4.7 Vegetation

Groenvlei is characterised by extensive sedgemarch and reedmarsh vegetation along its edges, with surrounding grassland and forest. The vegetation fringing the wetland includes reed (*Typha capensis* and *Phragmites australis*) and reed swamp (*Cladium mariscus*) communities (van der Merwe, 1976). These are particularly abundant along the northern and eastern shores. The vegetative fringe accounts for 31.6% of the total area of the lake. The south-shore woodland is dominated by milkwood trees (*Sideroxyon inerme*). Except for the plantations, few other trees of any size are found in the study area. Fijen (1995) presented a map and section illustrating the distribution of vegetation in and around Groenvlei (Figure 4.14).

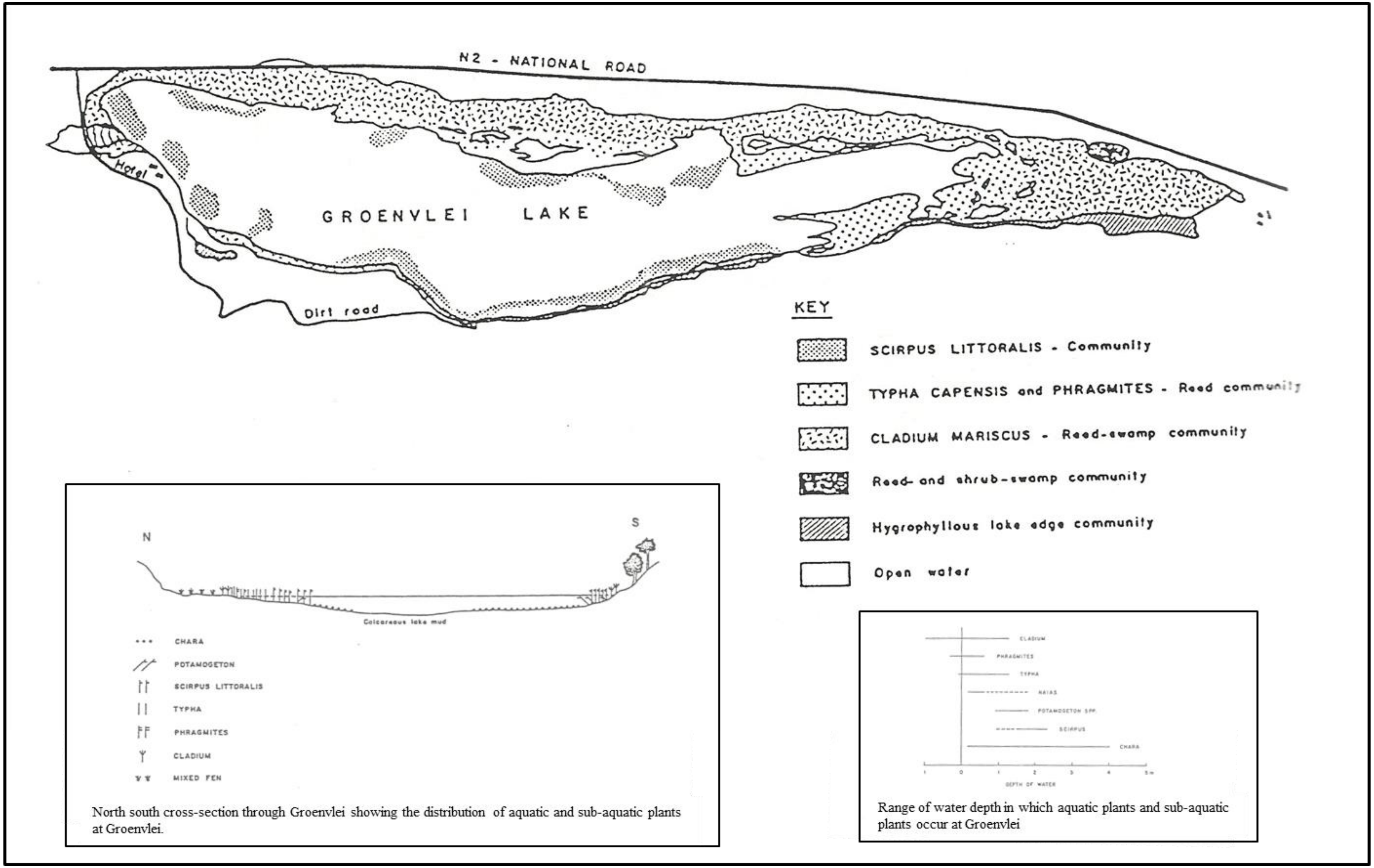

Figure 4.14: Vegetation in and around Groenvlei (after Fijen, 1995).

Plate 4.1: Inspection of the thick reed collar comprising phragmites, typha and cladium (Photo: E Murray).

Groenvlei is located in the Fynbos biome dominated by fynbos and renosterveld vegetation types. The lake straddles the Eastern fynbos – renosterveld bioregion in the north and South Strandveld bioregion in the south. Jones (2006) noted no proteas were found south of Groenvlei, with Rebelo (2006) suggesting this is a result of the young dunes in the south being too rich in calcium. As the dunes mature, leaching occurs and the pH of the soils decreases. Once the soils are acidic enough, proteas will move onto them.

Fynbos and kaffrarian thicket prevail as the natural vegetation. The coastal dune fynbos is dominated by elements of *Ericaceae* (Erica family) and *Restionaceae* (grass-like plants belonging to the Restio plant family). *Protea cynardoides* and *Leucadenrum saligum* are found on the older dune sands north of Groenvlei. A strong component of trees and shrubs characteristic of Kaffrarian thicket is also present. This vegetation type is highly susceptible to infestation of acacia and pine trees.

Fijen (1995) observed only pine trees (*Pinus radiata*) are planted in the study area. Five private timber growers operated within the catchment demarcated by him, covering an area of 270 ha (19.6%). Mapping of a property north-west of Groenvlei by Steyn and Bornman

(2006) showed that 57% of the property was covered by dense stands of Eucalyptus trees (*Eucalyptus grandis*), Pine (*Pinys spp*.), Blackwood (*Acacia melanoxylon*) Black wattle (*Acacia mearnsii*) and Rooikrans (*Acacia cyclops*). The exotic trees had either been introduced from adjacent afforested land or were remnants of former forestry activities on the farm.

4.8 Wetland Type

Considering the classification provided by Dini et al. (1998), Groenvlei has the general characteristics of a palustrine wetland. It is less than 8 ha in extent (3.59 ha), is non-tidal and has vegetative coverage of more than 30% of total area (32%). However, it has a maximum depth of greater than 2 m (5.6m) which qualifies it as a lacustrine wetland.

The lake was an estuarine system at a stage in its past, but this is no longer the case. Also, Groenvlei fails the criteria of an endorheic wetland as it is not a closed drainage system, with water discharging into the groundwater system along the southern shores.

4.9 Land Use

Groenvlei is bounded by the Goukamma Nature Reserve in the south, the eastward expanding village of Sedgefield in the west, and forest plantations in the north and east. The N2 national road was built along the northern shore of the lake, with the now disused Choo Tjoe railway line positioned directly north of the national road.

Tourism is the most important economic activity of Sedgefield. A hotel, chalets and caravan parks have long been established on the western shores of the lake. Two bush camps exist on the south shore. Excluding the area to the west, the area surrounding Groenvlei is rural in character and dwelling densities are low. A number of residential developments are being planned or proposed for the area, highlighting the increasing developmental pressure being exerted on the study area. No industrial, agricultural or mining activities take place adjacent to Groenvlei.

The lake is used for sailing, canoeing and fishing, bass fishing being particularly popular. Several hiking trials have been established along the southern shores of Groenvlei.

5 DESCRIPTION OF AQUIFER

5.1 Preamble

Proposed urban development in and around Sedgefield and limited municipal water supplies resulted in a number of investigations to develop groundwater water supplies. Borehole data from these investigations were added to a database that formed the basis of this assessment (Appendix B). At present, little groundwater is used in the immediate vicinity of the lake and the aquifer system is considered to be in a pristine state. Where used, groundwater is utilised for domestic supply and garden irrigation.

5.2 Aquifer Types

Groenvlei is located on an unconfined primary aquifer system comprising medium- to fine-grained unconsolidated to semi-consolidated aeolian sand. The aquifer does not owe its water storing and transmitting capabilities to secondary processes. The Eden Primary Aquifer [1] extends from Wilderness in the west to the Knysna lagoon in the east, a distance of about 35 km. At its widest, the aquifer is some 8 km wide. Little is known about the thickness of the aquifer. Based on available data, the primary aquifer beneath Groenvlei is at least 15 m thick. However, the aquifer could be tens of metres thick.

5.3 Hydraulic Properties

The primary aquifer is transmissive and capable of producing high-yielding boreholes. In addition to high-yielding boreholes drilled directly north of the town as part of the drought alleviation measures of 2009 (Parsons, 2009d), other boreholes have been drilled for water supply or monitoring purposes. The testing of these boreholes provided a means of quantifying the hydraulic properties of the aquifer (Driscoll, 1986; Kruseman and de Ridder, 1990; Domenico and Schwartz, 1997).

[1] This name is proposed to allow for easy reference.

Single wellpoints [1] are routinely used in the lower-lying parts of Sedgefield to provide water for garden irrigation. Small diameter tubes (32 mm) with slots cut into them and covered with fine netting [2] are installed to a depth of 8 m to 10 m, usually using a water-jetting technique. Water is abstracted by means of a vacuum pump and typically yields 0.4 L/s. Parsons (2003) estimated groundwater abstraction in Sedgefield for garden irrigation amounted to 750 m^3/d. This equated to about 35% of the village's peak summer demand.

Wellpoint clusters, an array of a varying number of spikes connected together, are used to produce larger amounts of water and have been tested at yields of 3 L/s and more. Because the layout of the clusters (shown in Figure 5.1) does not conform to assumptions associated with pumping tests, it is difficult to interpret hydraulic properties from the test data. However, information gleaned from the tests supports the conclusion that the aquifer is highly transmissive in character.

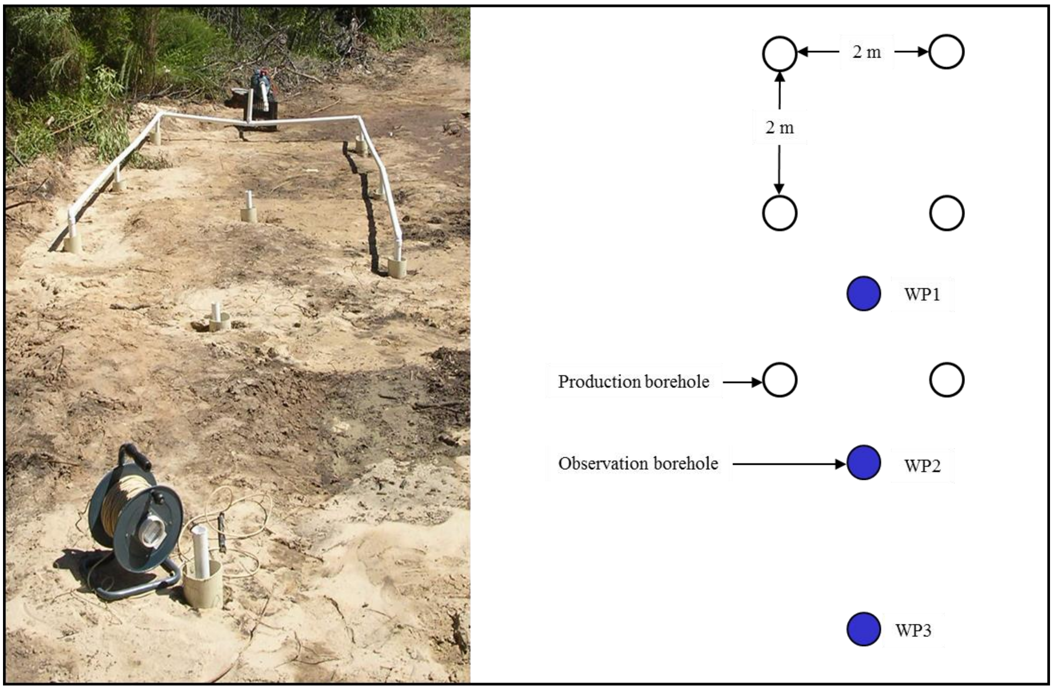

Figure 5.1: Layout of the Myoli Beach wellpoint cluster showing the joined spikes in the background and observation spikes in the foreground.

1 Locally referred to as spikes.
2 Typically pantyhose are used.

Using the Thiem distance – drawdown method described by Kruseman and de Ridder (1990) and Domenico and Schwartz (1997) [1], it was possible to approximate both T and S using drawdown data monitored while conducting constant discharge tests on LPWC and WMWC directly west of Groenvlei (Table 5.1, Figure 5.2). It was considered that the influence of the wellpoint array diminishes with distance. Distance from the pumped well and drawdown data was plotted on a semilogarithmic graph (Figure 5.3), from which Δs and r_o were determined. The following equations were then used to determine T and S:

$$\mathbf{T = \frac{2.3\ Q}{2\pi\ \Delta s}}$$

$$\mathbf{S = \frac{2.25\ Tt}{r_o^{\ 2}}}$$

where	T	=	transmissivity (m^2/d)
	Q	=	yield (m^3/d)
	Δs	=	drawdown per log cycle (m)
	S	=	storativity
	t	=	time (days)
	r_o	=	intersect of straight line with 0 m drawdown.

While the interpreted T values appear reasonable, determination of S was sensitive to small changes in r_o, as reflected in the outcome of the analyses.

Table 5.1: Hydraulic parameters of the primary aquifer approximated from distance – drawdown data from constant discharge tests conducted on two wellpoint clusters directly west of Groenvlei.

Cluster	**Duration (mins)**	**Yield (L/s)**	**T (m^2/d)**	**S**	**Source**
LPWC-east	2 880	2.9	390	0.24	Parsons (2005)
LPWC-south	2 880	2.9	430	0.02	Parsons (2005)
WMWC	4 320	0.8	160	0.55	Parsons (2006d)

1 The method uses the Cooper-Jacob equation.

Figure 5.2: Locations of boreholes and wellpoint clusters subjected to constant discharge pumping tests.

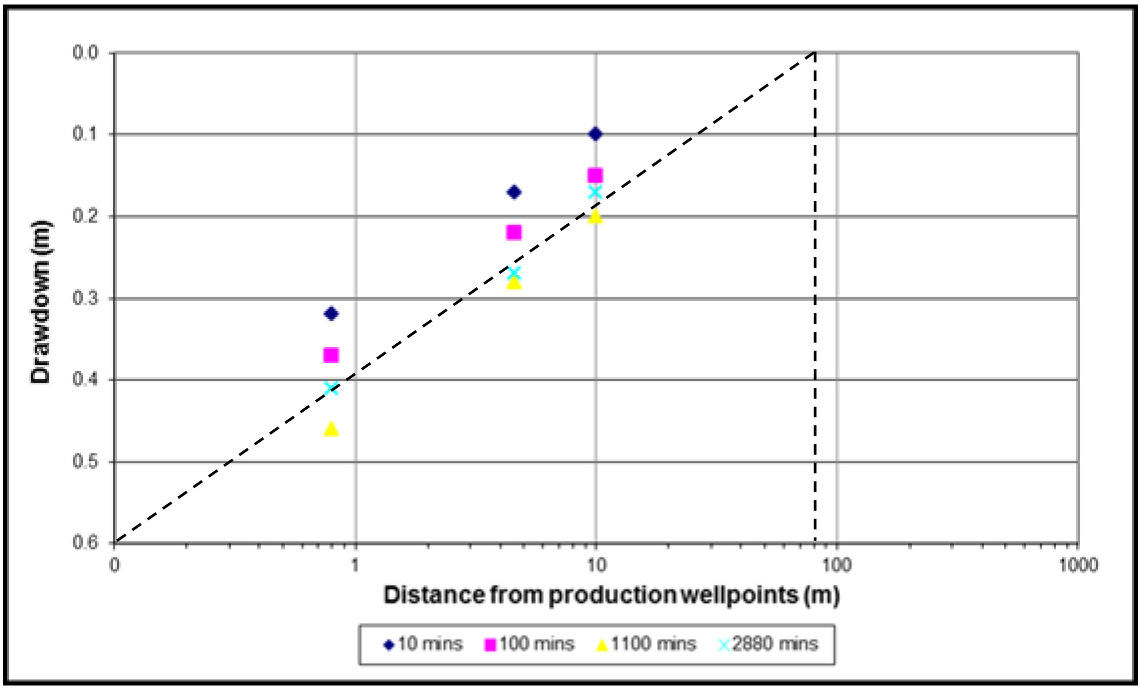

Figure 5.3: The eastward distance – drawdown curve from the testing of the Lake Pleasant wellpoint cluster (LPWC).

Most of the boreholes tested in the Sedgefield area were also subjected to step tests. Step tests are useful for determining the yield of boreholes and establishing the depth of pump installation. Often the primary purpose of step tests conducted in the study area was to develop the boreholes, with determination of probable yield being secondary. An example of the graphs generated from a step test is presented in Figure 5.4. The step tests usually produced graphs with similar characteristics:

- The drawdown induced by pumping rapidly stabilises;
- The boreholes are generally high yielding; and
- Boreholes recovered quickly on cessation of pumping, often in a matter of minutes.

Unlike step tests that provide information on the borehole, constant discharge tests are used to glean information about the aquifer. Boreholes subjected to constant discharge tests are listed in Table 5.2 and their positions shown in Figure 5.2. Generally, constant discharge tests are pumped at the highest rate practically possible, the strategy being to learn as much as possible about aquifer response to pumping. In most instances this strategy was not

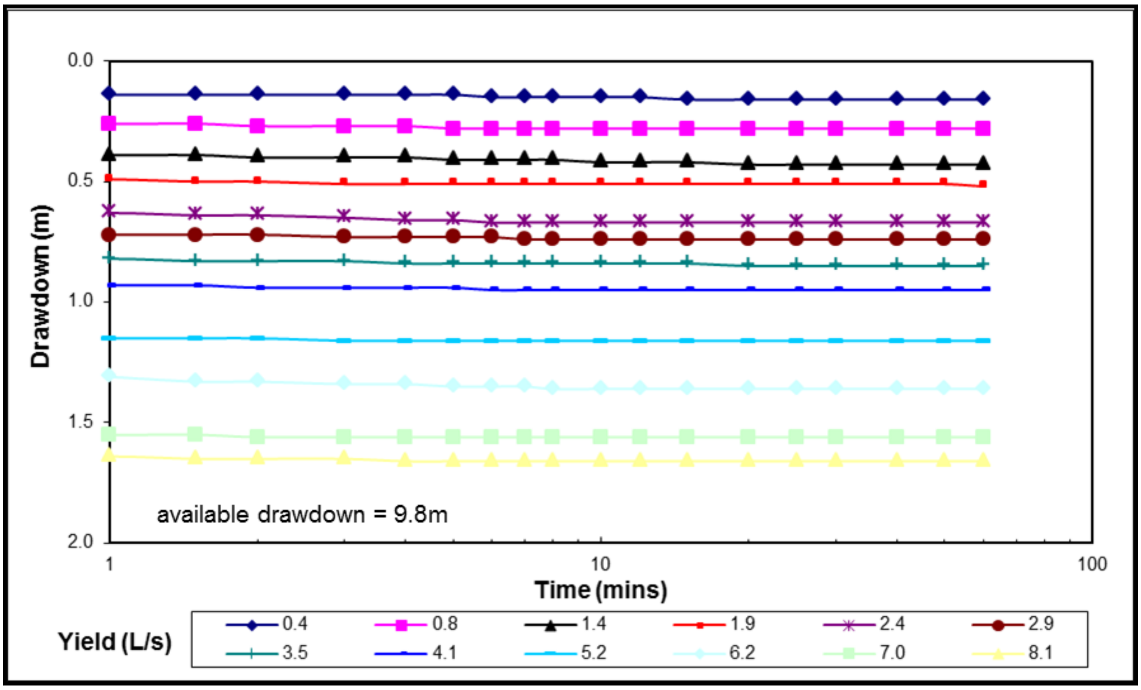

Figure 5.4: Monitored aquifer response while conducting a step test on BHE5 (After Parsons, 2009d).

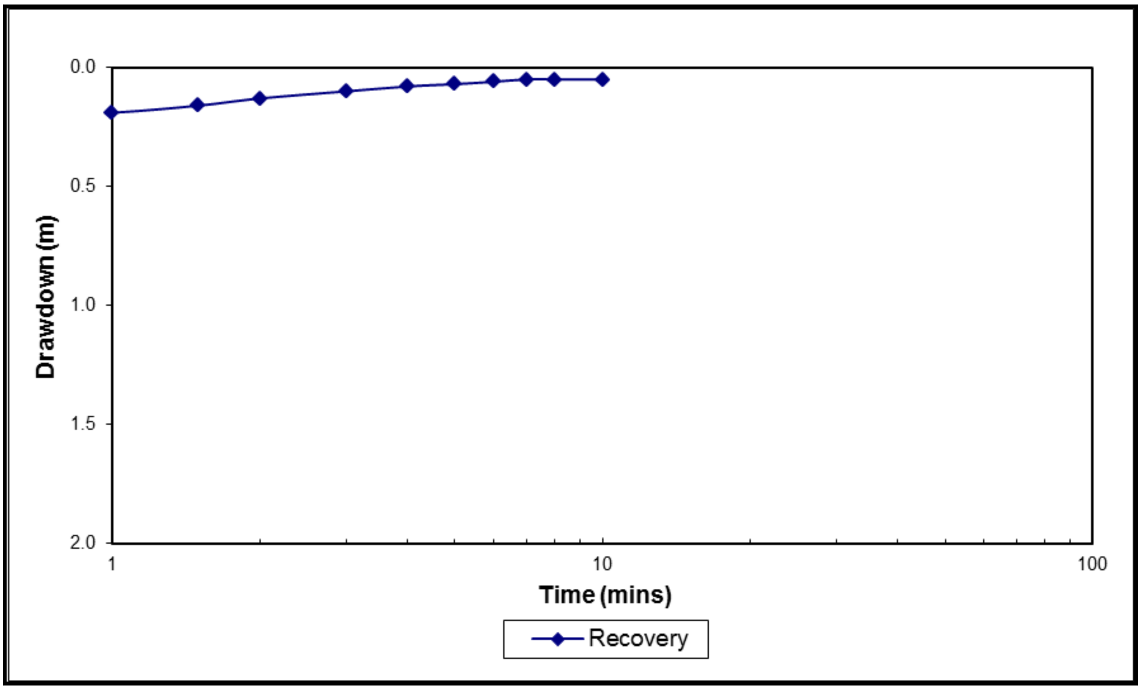

Figure 5.4 cont.: Borehole recovery after cessation of the step test conducted on BHE5.

Table 5.2: Interpreted transmissivities of boreholes near Groenvlei subjected to constant discharge pumping tests.

Bh No.	Borehole Penetration (m)	Duration (mins)	Yield (L/s)	Draw-down (m)	Recovery (mins)	T early (m^2/d)	T Late (m^2/d)	T recovery (m^2/d)	Source of Data
BH5	17.5	4 200	1.9	1.9	2 000	175	-[**]	175	Parsons (2004b)
BH6	6.2	4 320	1.2	4.2	1 500	-	-	270	Parsons (2004b)
BHE5	11.3	4 320	7.1	1.8	300	680	-	840	Parsons (2009d)
BHE9	23.2	2 880	0.8	0.4	10	270	270	-	Parsons (2009d)
BHE10	23.0	4 320	2.1	0.6	4	1660	-	-	Parsons (2009d)
DESA1 (2) [*]	58.3	*7 200*	*20.0*	*10.6*	*150*	*210*	*220*	*743*	Parsons (2013a)
DESA2 [*]	53.9	*1 440*	*14.3*	*1.6*	*1*	-	-	-	Parsons (2011)
DOD1	12.6	2 880	1.1	2.0	10	30	46	58	Parsons (2008b)
DOD1B	19.0	2 880	2.5	3.5	3	66	75	-	Parsons (2008b)
GRU1	16.0	4 320	0.42	0.1	-	-	90	-	Parsons (2006c)
GRU2	14.2	4 320	0.5	1.3	2	225	225	-	Parsons (2009c)
GRU3	13.7	4 320	0.5	0.9	2	285	285	-	Parsons (2009c)
SEN1	10.8	4 320	1.0	0.4	1	360[***]	360[***]	-	Parsons (2006b)

Notes:

[*] 100 m from the sea.

[**] water level rose by 0.15 m between 100 m and end of test

[***] very flat graph

employed because of concerns about nearby surface water bodies with non-potable water quality. Rather, boreholes were tested at the rate required to meet water demand. BHE5 was pumped at 7.1 L/s while the two boreholes used to produce feedwater for the desalination were pumped at 20.0 L/s and 14.3 L/s. The two feedwater boreholes failed the assumption that the aquifer is of infinite areal extent as the boreholes are 100 m from the sea.

All of the tested boreholes only partially penetrate the aquifer. Excluding the two desalination plant feed boreholes, the average depth that the tested boreholes penetrate the aquifer is 15.2 m. Partial penetration induces vertical flow, invalidating the assumption of pumping test theory that flow is horizontal. Vertical flow was exhibited at GRU1 by the immediate deterioration of groundwater quality, in spite of pumping the borehole at a very low rate and inducing a drawdown of only 0.1 m. Methods are available for correcting for partial penetration (Kruseman and de Ridder, 1990), but are dependent on knowing the thickness of the aquifer. In the case of Groenvlei, this is not known.

Failure to address partial penetration leads to an over-estimate of transmissivity. In their assessment of simulated single-well tests interpreted with the Cooper-Jacob straight-line method, Halford et al. (2006) reported transmissivity estimates of simulated unconfined aquifers averaged twice known values. The partial penetration problem may not be as limiting as expected if it is assumed the interpreted parameters are representative of that (shallow) part of the aquifer that sustains Groenvlei. Deeper flow may not interact with Groenvlei at all, underpassing the lake and discharging to the sea. Interpreted transmissivity values thus may not reflect the entire thickness of the aquifer, but are rather apparent transmissivities representative of that part of the aquifer influenced by the pumping tests.

Graphs resulting from the testing of BHE5 and GRU2 are presented in Figure 5.5 and Figure 5.6 as examples of the characteristic aquifer response. The challenge of interpreting the pumping test data from the unconfined primary aquifer related to the flat gradients of the drawdown curves and the related difficulty of determining Δs. All the graphs displayed similar characteristics, even those boreholes pumped at a high rate (BHE5) and those adjacent to the sea (DESA1, DESA2). Where possible, the Cooper-Jacob approximation addressed in Section 2.5.3.1 was used to interpret both drawdown and recovery data (Table 5.2). It is of interest that only BHE9 and DOD1 behaved as one would expect of an unconfined aquifer, with the graphs showing delayed-yield characteristics. Also, pumping BHE10 at 2.1 L/s for 4 320 mins did not produce a response in BHE9 located just 57 m away.

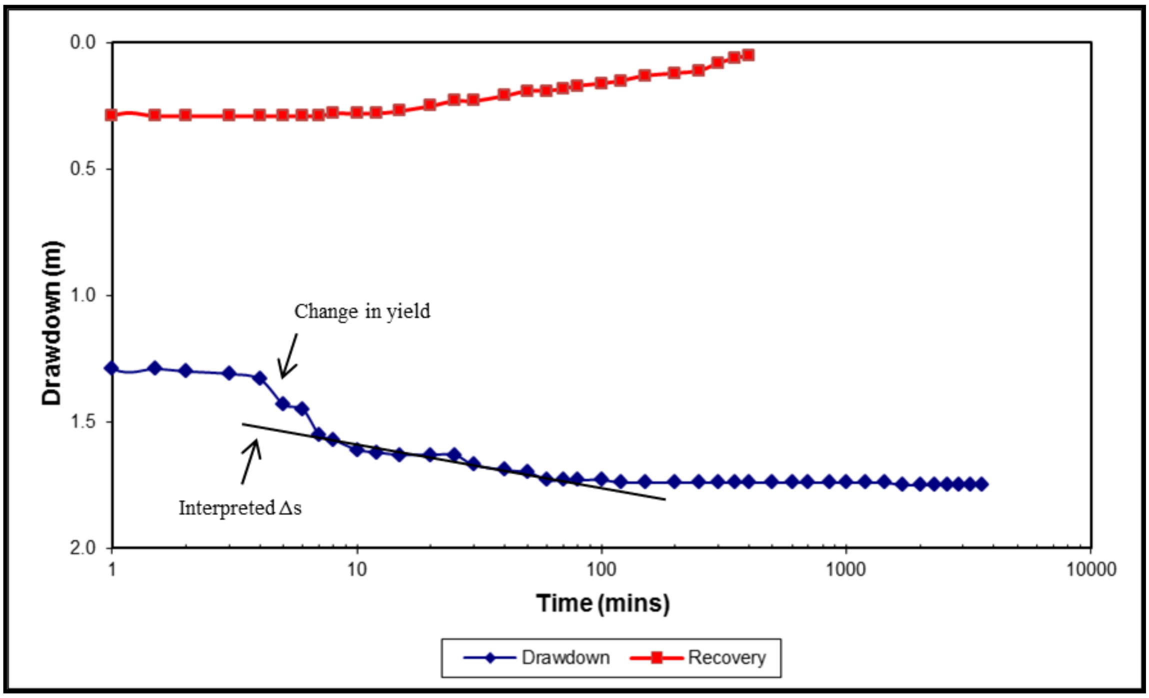

Figure 5.5: Drawdown and recovery of BHE5 when subjected to a constant discharge test at a rate of 7.1 L/s (After Parsons, 2009d).

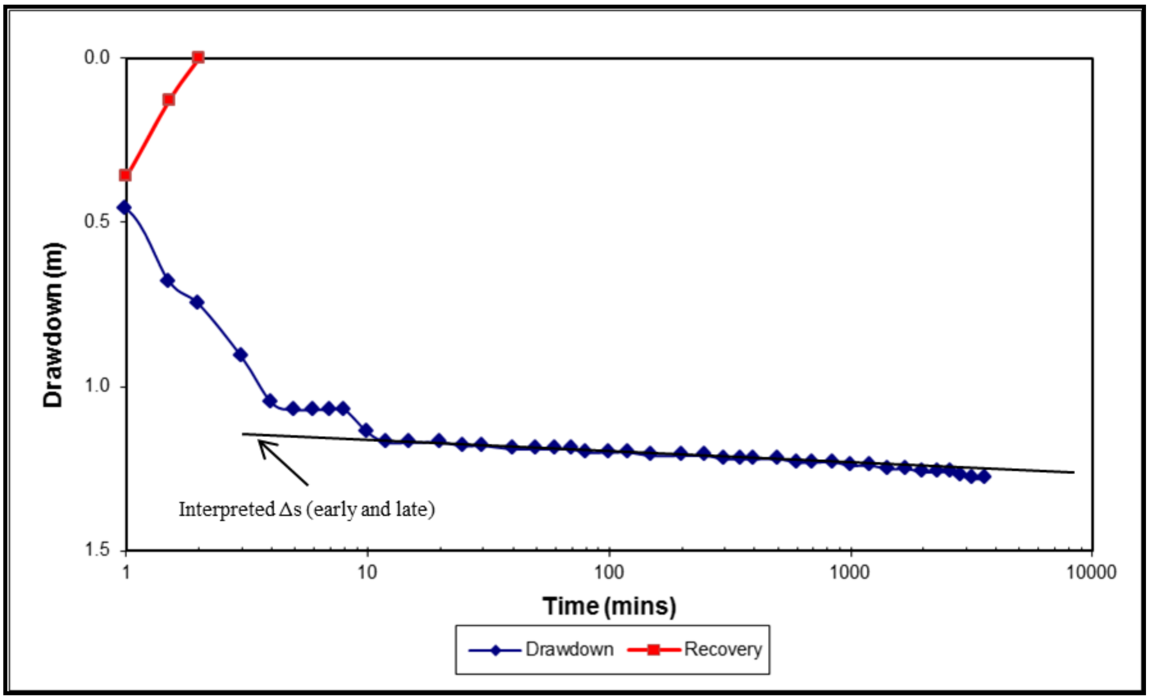

Figure 5.6: Drawdown and recovery of GRU2 when subjected to a constant discharge test at a rate of 0.53 L/s (After Parsons, 2009c).

Vandoolaeghe (1989) had a similar experience where boreholes tested in the Cape Flats displayed similar characteristics, with the groundwater level reaching its maximum depth only a few minutes after pumping started. This is characteristic of vertical leakage and dictates that approaches such as the Walton or Hantush methods be used to interpret the pumping test data. The early time data could also be interpreted using the Boulton curves. However, interpretation was made difficult by the almost immediate onset of leakage (or delayed yield) and too few data points were available for curve matching.

Inspection of the interpreted T values suggested the high value of 1 660 m^2/d obtained at BHE10 was unrepresentative of the general area. The average T value determined from the remaining data was 270 m^2/d. Relatively low T values (30 m^2/d – 90 m^2/d) were obtained at boreholes directly north and south of Groenvlei (DOD1, DOD1B and GRU1) while high values were obtained at BHE5 (680 m^2/d - 840 m^2/d).

Monitoring aquifer response to pumping the emergency borehole wellfield for a month allowed for the Excel-based Cooper-Jacob wellfield model to be used to determine hydraulic properties of the aquifer. The model was described by Baker and Dennis (2012) and Murray et al. (2012), and is based on the commonly used Cooper-Jacob approximation of the Theis groundwater flow equation. Six of the emergency boreholes drilled directly north of Sedgefield in 2009 were pumped continuously at a collective rate of 13.7 L/s for 30 days. The drawdown induced by pumping is shown in Figure 5.7. Based on observed drawdowns and the interpreted radius of influence, transmissivity and storativity were modelled to be 300 m^2/d and 0.15. Initially, Fleisher (1990) set the specific yield of the Witzand wellfield in Atlantis at 25%, but modelling results prompted Fleisher and Eskes (1992) to reduce this to 15.7%.

By way of comparison, Dennis (2010) used T and S values of 300 m^2/d and 0.2 in her model of the area directly west of Groenvlei. Vivier (2009) used a T value of 150 m^2/d is his model of the area around Groenvlei. The hydraulic properties determined during this research are also comparable to those of the Atlantis Aquifer, Cape Flats Aquifer and the Zululand Coastal Aquifer (Henzen, 1973; Fleisher and Eskes, 1992; Kelbe and Rawlins, 1992a, 1992b; Wright and Conrad, 1995; Meyer et al., 2001).

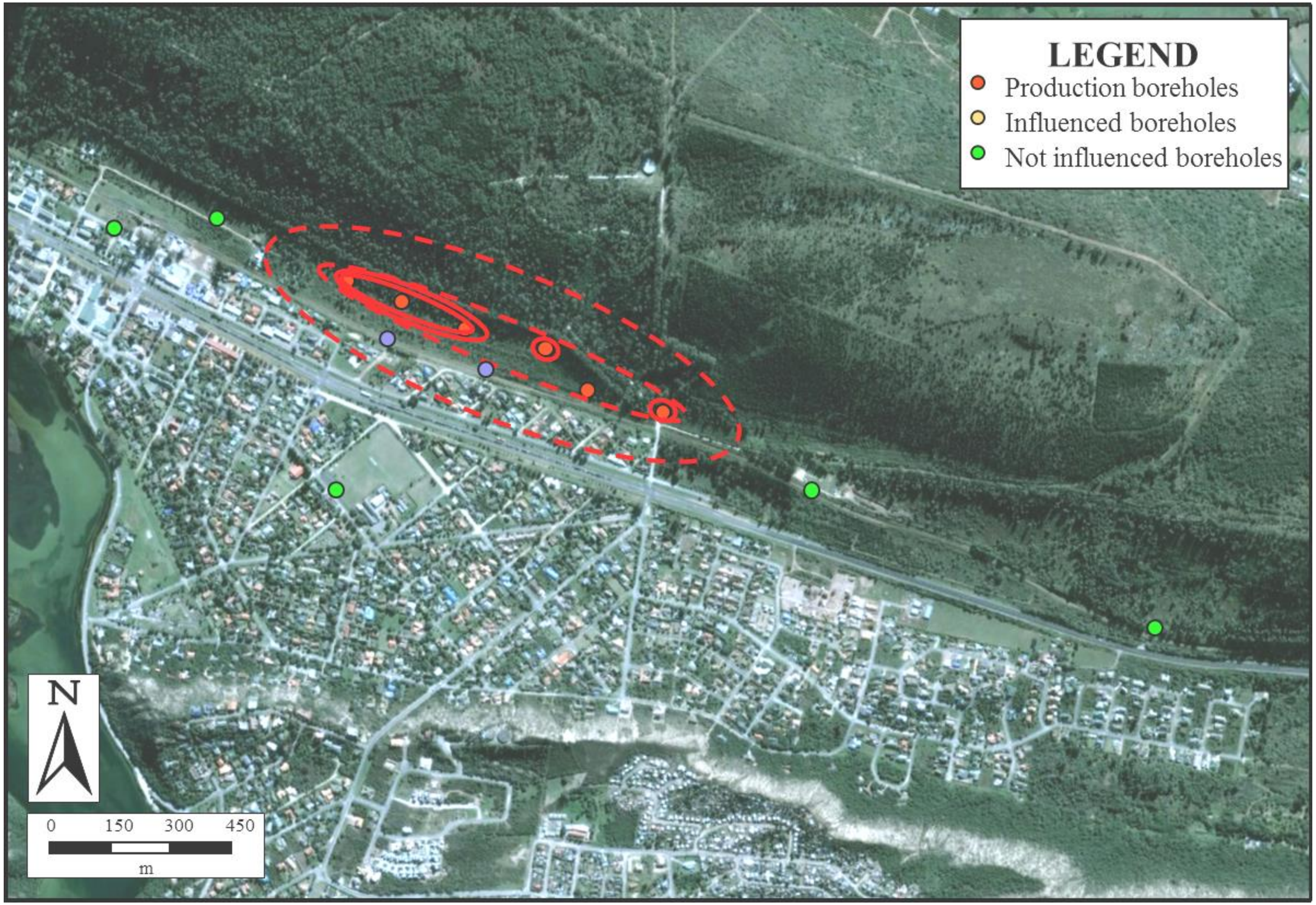

Figure 5.7: Interpreted radius of influence resulting from abstraction from the Sedgefield emergency wellfield for a period of a month. This information was used in the Cooper-Jacob wellfield model to determine aquifer parameters T and S (After Parsons, 2013b).

5.4 Groundwater Levels

Groundwater levels have been measured in an around Sedgefield, mainly on an ad hoc basis. Some monitoring has been undertaken that allowed temporal patterns to be assessed, but limited monitored groundwater level data are available for the area in close proximity to Groenvlei. Elevations of only a few of the wellpoints and boreholes have been accurately surveyed. When not surveyed, borehole elevations were interpolated from 1 : 10 000 orthophoto maps of the area on which contours are drawn at 5 m intervals.

Groundwater levels monitored at MWP4 and the water level of Groenvlei monitored between May 2005 and February 2006 displayed a close correlation (Figure 5.8). While the groundwater level remained constant between May and October, the water level of the lake gradually increased. Both levels responded to the 59 mm of rainfall that fell between 11 and 17 November 2005. It is of interest that during peak summer, the rate of decline of water level in the lake and groundwater was the same. During mid-December 2005, the rate of decline in groundwater levels reduced, but that of the vlei continued at the same rate. This could suggest the effects of evapotranspiration on groundwater decreases as the water level in the lake drops.

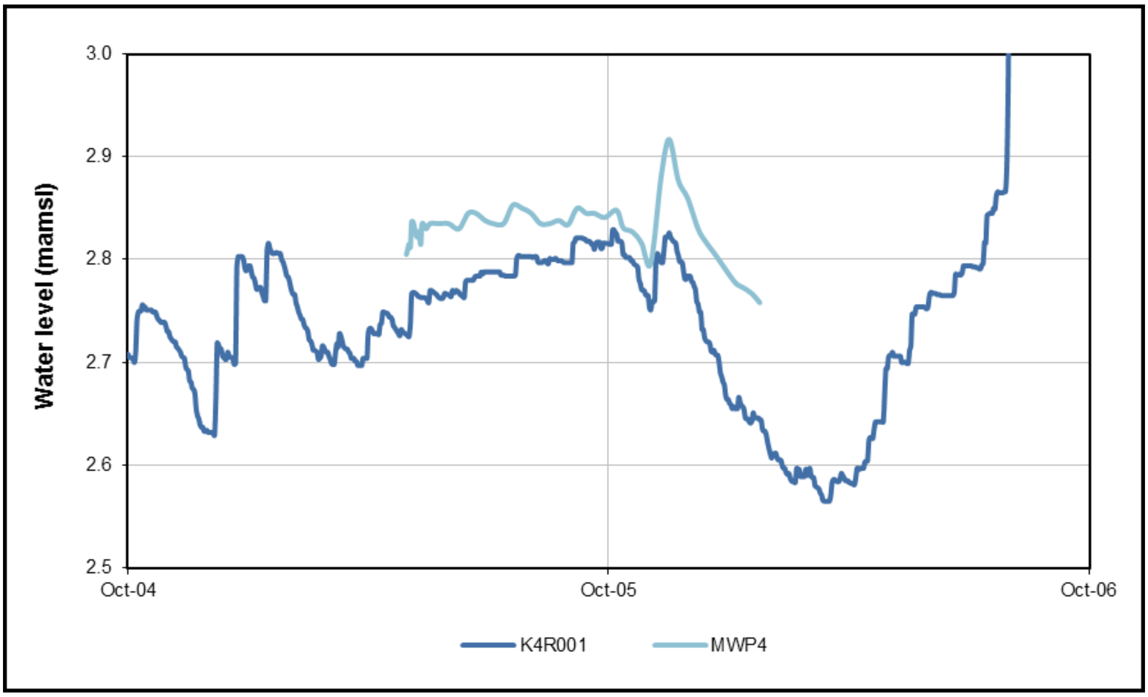

Figure 5.8: The relationship between groundwater level (WP4) and surface water level (K4R001). MWP4 was located 50 m from the edge of the lake.

Monitoring south of Groenvlei confirmed water flows from the lake southwards to the sea (Figure 5.9). Groundwater fluctuations observed in GRU1 are similar, but slightly more subdued than the water level of the lake. Also, the groundwater response lags behind that of the lake by about 2 months. The monitored groundwater level at GRU3 is significantly more subdued than that at GRU1. The observed pattern indicates a lessening of the control of the lake water level with distance to the sea.

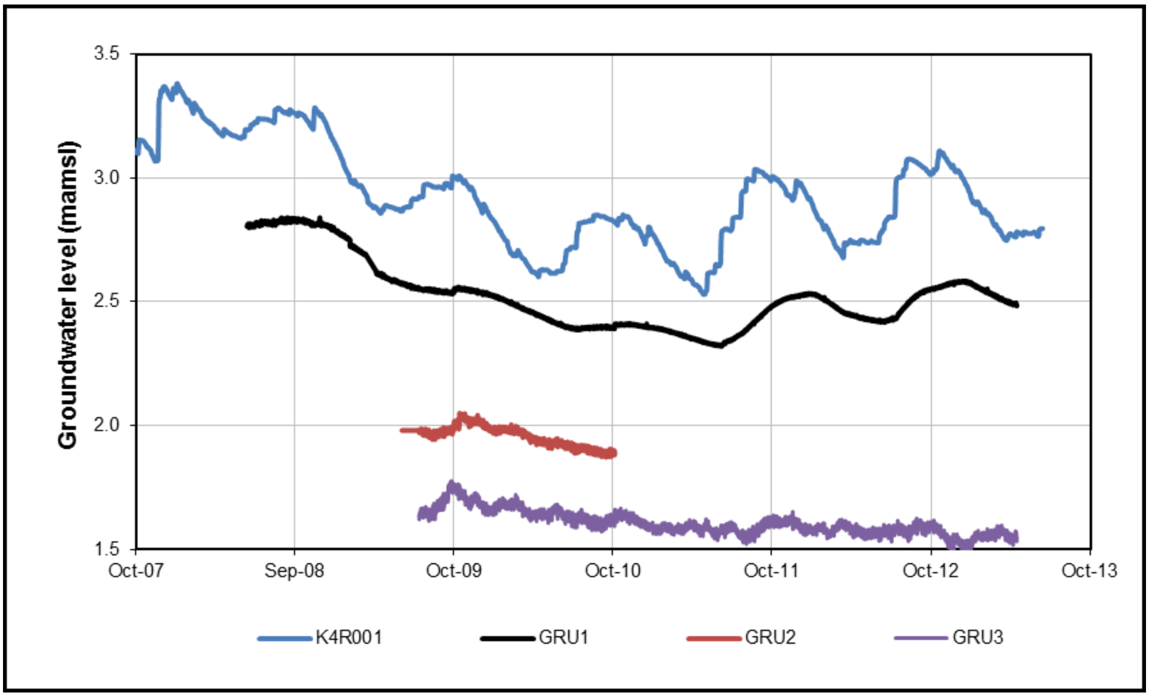

Figure 5.9: Monitored groundwater levels south of Groenvlei, confirming flow from the lake southwards to the sea.

MWP2, MWP5, MWPE1 and BHE8 display similar temporal trends, but different from those of BH9 and DOD1A (Figure 5.10). The former group have shallow groundwater levels, while the latter two boreholes have a vadose zone of more than 23 m. The exception to this is BHE8 which has a depth to groundwater of 20 m. It is thought boreholes with a shallower depth to groundwater respond quicker to recharge events, while a thicker vadose zone results in a more subdued and delayed response buffered by the vadose zone.

A review of Figure 4.7 and the groundwater level contour map of the study area (Figure 5.11) showed the vadose zone to be deceptively thick. It is only in the west in the low-lying parts of Sedgefield that the vadose zone is 10 m thick or less. To the north of Groenvlei the vadose zone is between 10 m and 50 m thick and to the south the vadose can be up to 200 m thick. Drilling deep boreholes at Sedgefield (BH1, BH2, BH3, BH4, BH5, BH6 and BH7) revealed that much of the dune comprised unconsolidated sands, but relatively thin lenses of clay were encountered in places.

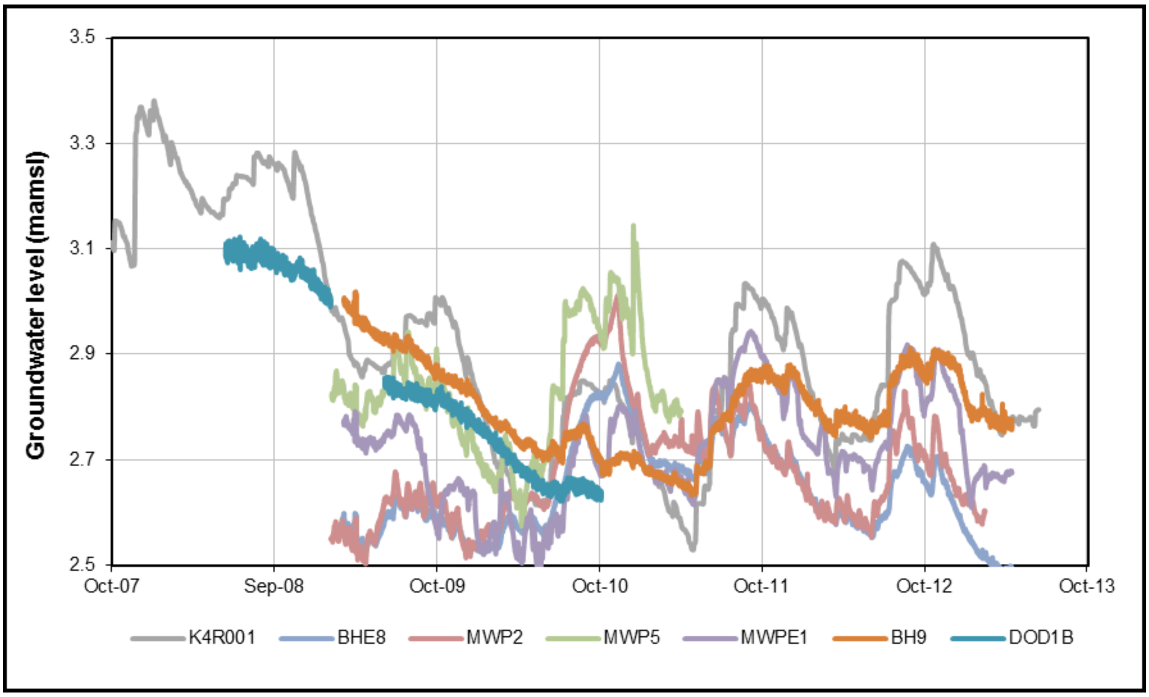

Figure 5.10: Monitored groundwater levels reflecting the difference in response between boreholes with a shallow water level and those with a thicker vadose zone.

5.5 Direction of Groundwater Flow and Hydraulic Gradients

Available groundwater level data were used to construct a groundwater level contour map of the study area (Figure 5.11), allowing for determination of the direction of groundwater flow and hydraulic gradients (Figure 5.12). Localised assessments of the direction of groundwater flow at the wastewater treatment works (Parsons, 1997), directly west of Groenvlei at Lake Pleasant and Windemere (Parsons, 2005a, 2006d), near the mouth of Swartvlei (Parsons, 2006b, 2006e, 2011) and directly north of the village (Parsons, 2009d, 2013b) informed construction of the map, as did sea level and the water level of Swartvlei and Groenvlei.

Figure 5.11: Groundwater level contour map of the study area, with location of the boreholes at which groundwater level measurements were taken.

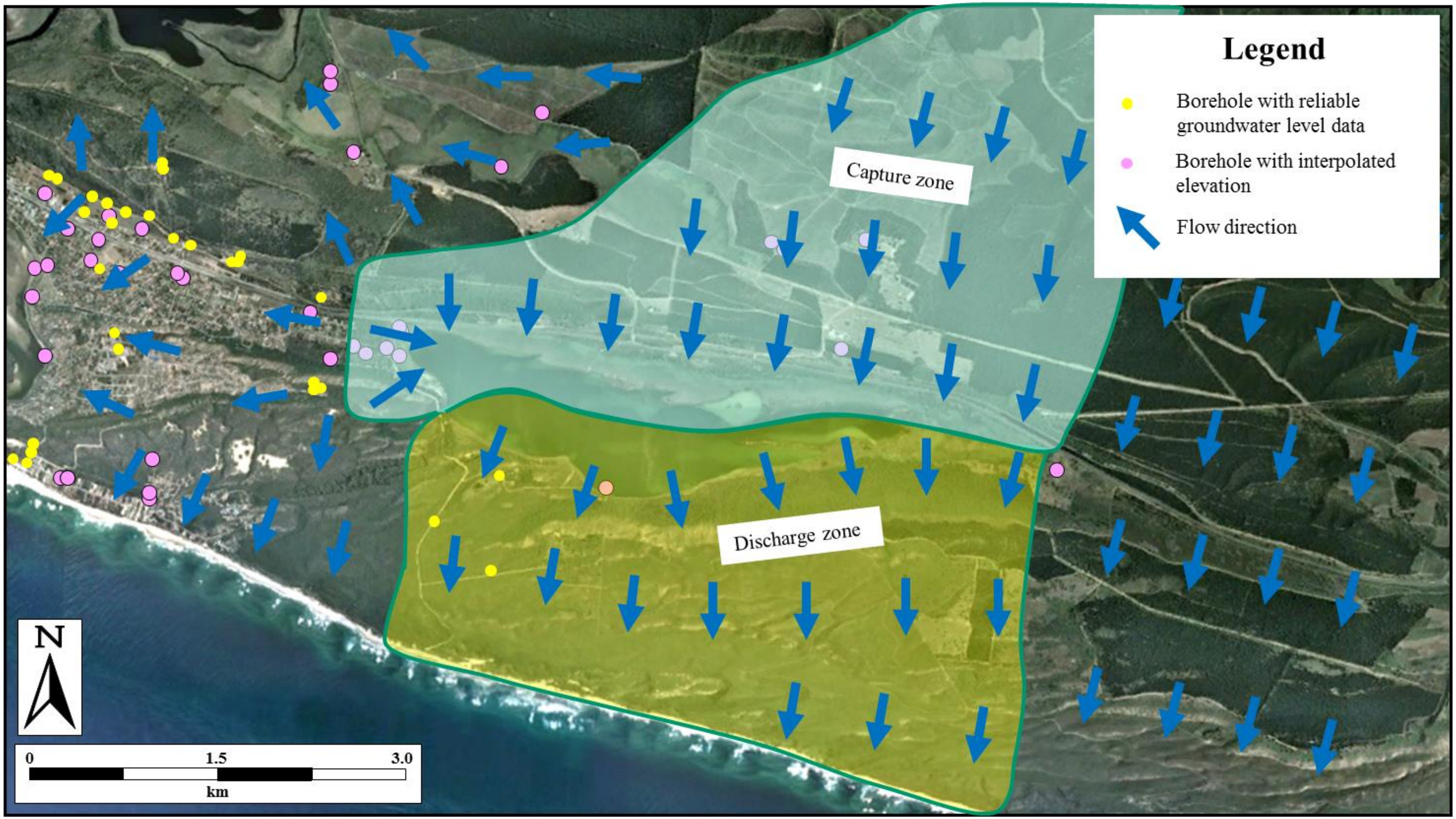

Figure 5.12: Interpreted direction of groundwater flow (blue arrows) used to delineate the capture and discharge zones of Groenvlei.

The uneven distribution of reliable groundwater level data [1] required the contour map be constructed by hand, with surface water levels playing a key part in the contouring process. Data based on borehole elevations interpolated from 1 : 10 000 orthophoto maps was also used, but the data were ignored if better data were available nearby. Most of the groundwater level data are located west of Groenvlei, with critical groundwater level data being measured in three boreholes south of the lake (GRU1, GRU2 and GRU3). Because seasonal groundwater level fluctuations were observed to be small, temporal variance of the data was not limiting. Where multiple groundwater level measurements were available for a borehole, the median groundwater level was used.

Interpretation of the groundwater level data confirmed the flow-through character of the lake and allowed for the capture and discharge zones of Groenvlei to be delineated. Insufficient data were available to accurately define the full areal extent of the capture zone. However, the interpreted contour data suggests the capture zone covers an area of some 25 km^2 (Figure 5.12). The inflow boundary was measured to be 6 040 m in length while the discharge boundary was 4 520 m long. A 100 m section in the south-western corner of the lake was deemed to be a no-flow zone.

Hydraulic gradients were found to be relatively flat. Along the western and northern inflow boundary, the hydraulic gradient was interpreted to be 0.0016. This is similar to that measured north-west of the lake where the elevation of the 10 emergency boreholes was accurately surveyed (Parsons, 2009d). The gradient flattened slightly from west to east along the southern discharge boundary, with an average of 0.0013.

The gradients presented above are similar to those determined at other primary aquifers along the South African coast (Henzen, 1973; Timmerman, 1985, Dyke, 1992, Kelbe and Rawlins, 1992a; Wright, 1991; Du Toit and Weaver, 1995; Kelbe et al., 1995; Wright and Conrad, 1995; Meyer et al., 2001; Parsons, 2009a). Similarly, the temperature profiling of Paridaens and Vandenbroucke (2007) presented in Section 2.1.4, the EC profiling of boreholes south of Groenvlei (Section 5.7.2), and stable isotope analysis of Groenvlei and groundwater (Section 5.7.4) support the interpreted flows into and out of Groenvlei.

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1 Includes only boreholes that have had their elevation determined by a land surveyor or boreholes whose elevation could be reliably determined from a 1 : 10 000 orthophoto.

5.6 Recharge

A review of estimates of recharge of other primary aquifers along the South African coast yielded an average of 20% MAP (Section 2.6.1). The availability of chloride concentration data for both rainfall and groundwater suggested the CMB method could be used to assess recharge. Similarly, the availability of monthly rainfall and groundwater level data suggested the water balance methods described in Section 2.6.3 could also be applied.

5.6.1 CMB method

Accepting a Cl concentration of about 20 mg/L in rain (Section 2.6.2) and a harmonic mean of 135 mg /L for the Cl concentration of groundwater north of Groenvlei, recharge was calculated to be 14.8% MAP. This estimate is less than expected in an area with a high infiltration rate and no surface run-off. It is also below the average determined for coastal primary aquifers.

5.6.2 Water balance methods

Application of the water balance methods proved more challenging as there were at least three parameters that needed to be considered when trying to match the modelled groundwater levels with the measured data. For example, the SFV method required specific yield, catchment size, lag and inflow and outflow to be approximated in the recharge estimation process. Also, two sets of rainfall were available for the analysis. Modelling with the EARTH model showed the outcome using the two rainfall data sets to be quite different (Figure 5.13). Assuming a specific yield of 15%, recharge could range between 13% and 25% MAP.

Poor fits were obtained with the CRD method and the SVF method, with the thickness of the vadose zone (and resulting lag time) and interpreted high specific yield possibly playing a contributing role. The EARTH model yielded reasonable results when a high S value was used (Figure 5.14 and 5.15). Notwithstanding an interpreted recharge of about 20% when using an S of 0.15, the water balance methods indicated:

- S is high, and may be in the range of 0.20 – 0.25;
- Recharge is high, with 20% possibly being a minimum estimate;
- The vadose plays an important role in storing water as it percolates downward to the saturated zone.

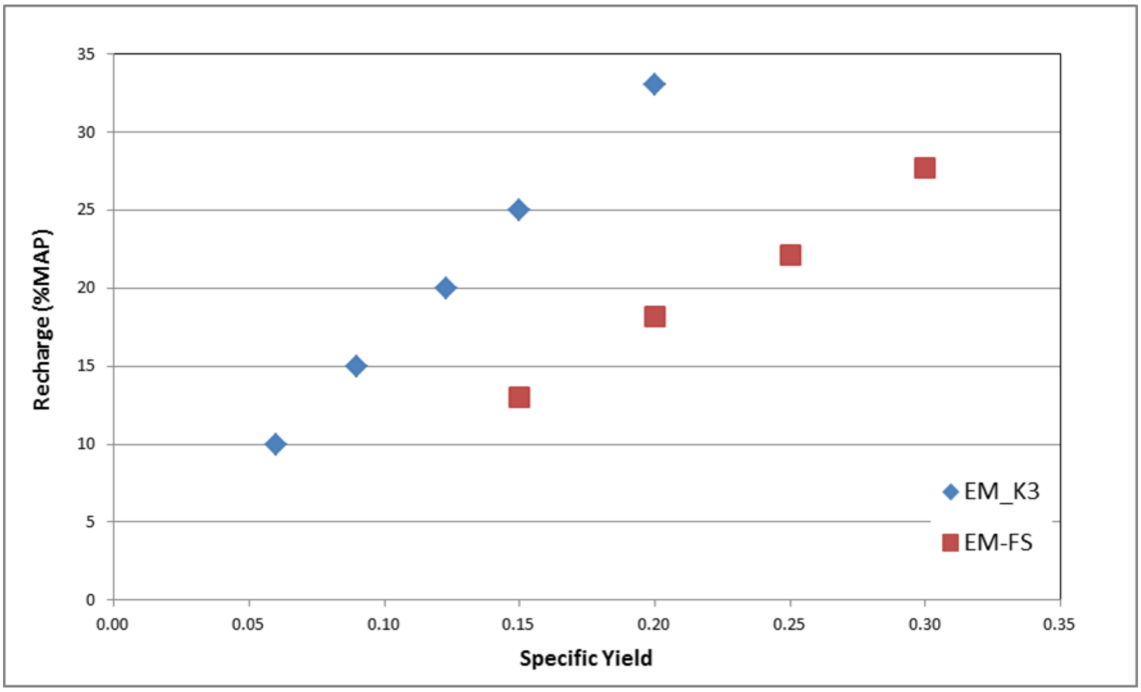

Figure 5.13: Modelled relationships between recharge and specific yield using the EARTH model, groundwater level data from BHE8 and monthly rainfall data measured at K3E003 and the Sedgefield Fire Station.

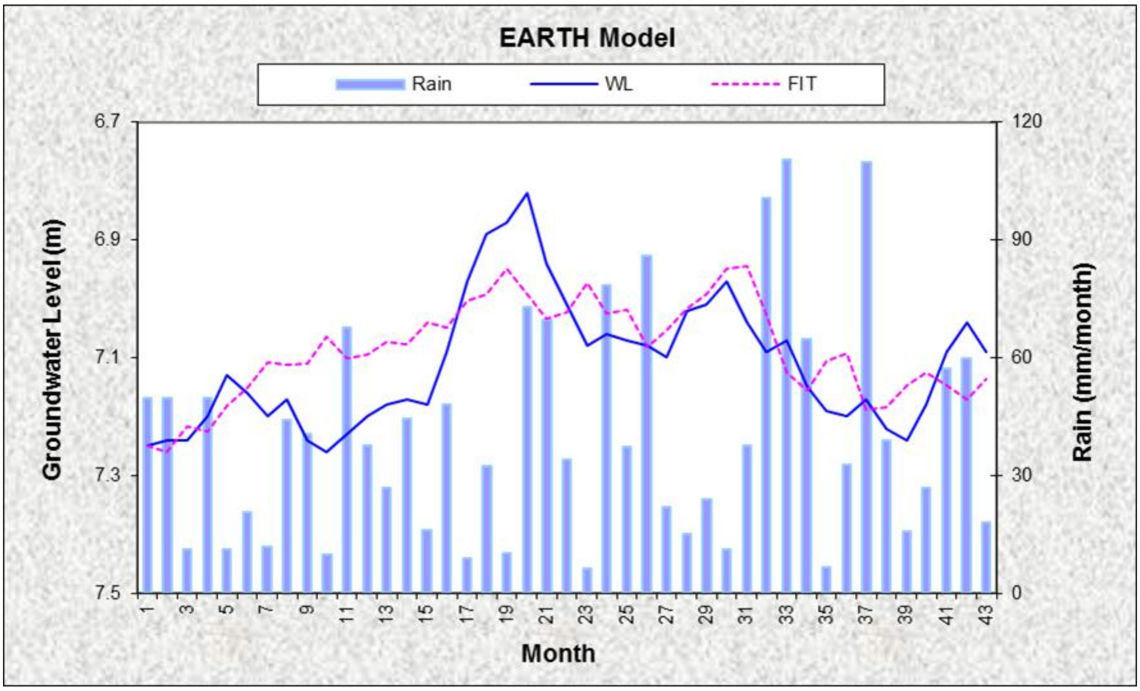

Figure 5.14: Modeling recharge at MWP2 using rainfall data measured at K3E003, a recharge of 21% MAP, a S of 0.15 and a resistance of 121.

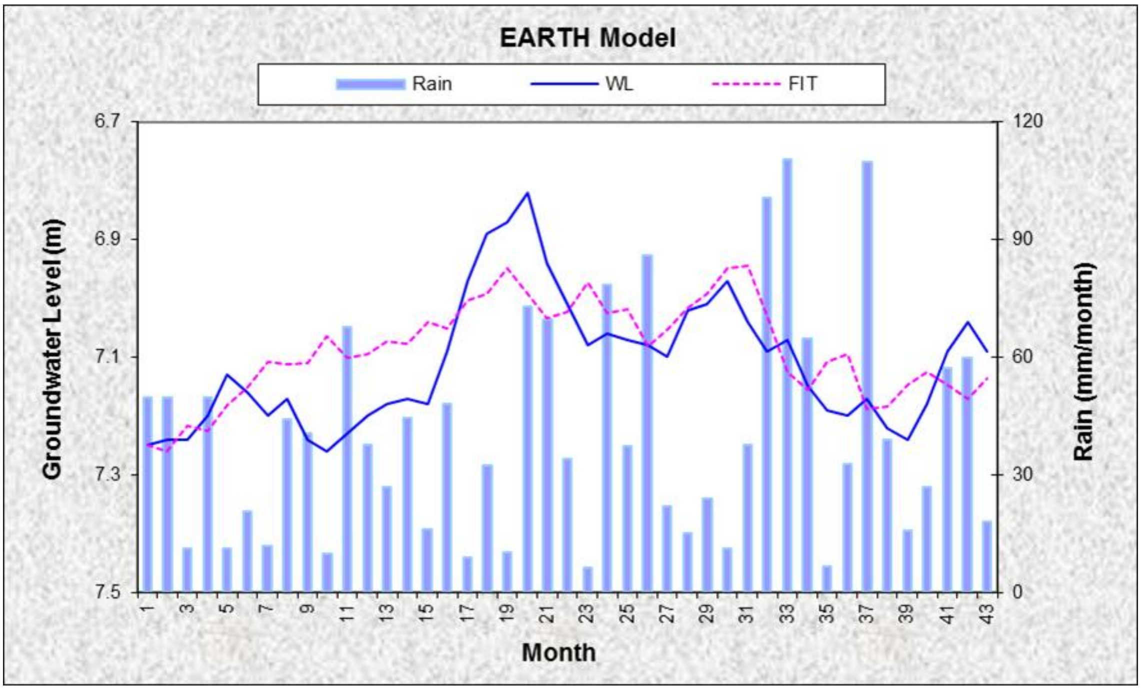

Figure 5.15: Modeling recharge at MWP2 using rainfall data measured at K3E003, a recharge of 28% MAP, a S of 0.20 and a resistance of 121.

5.7 Groundwater Quality

5.7.1. Preamble

Sampling of groundwater in and around Sedgefield indicated the water has a mixed cation and anion character (Figure 5.16). Generally, Mg and SO_4 concentrations are less dominant than Na, Ca, Cl and TAL. The character of the groundwater ranges from CaAlk to NaCl. Groundwater has a median EC of 105 mS/m and a median pH of 7.5. Boreholes directly north and east of the lake have an EC of 90 mS/m and 60 mS/m respectively (Figure 5.16). Boreholes north-west of the site show an EC typical of the general area (100 mS/m) while boreholes to the west have a slightly higher EC at 140 mS/m. Boreholes and wellpoints directly south of Groenvlei display a quality similar to the lake i.e. NaCl water with an EC of 500 mS/m. The chemical characteristics of surface and groundwater support groundwater level data, indicating water discharges from the lake along its southern boundary into the groundwater, whereafter it flows southwards to the sea.

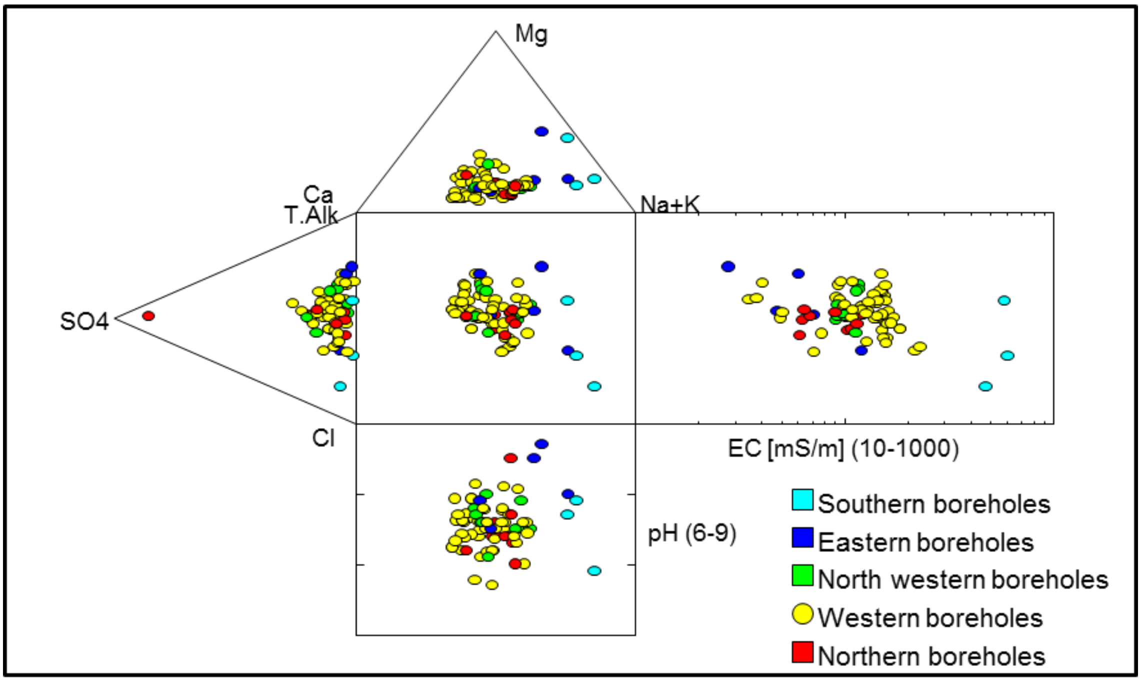

Figure 5.16: Durov diagram showing the hydrochemical character of water monitored at Groenvlei and groundwater sampled in the near vicinity of the lake

5.7.2 Groundwater quality south of Groenvlei

Multiple salinity profiling in GRU1 revealed a two-layered profile (Figure 5.17) [1]. The upper layer is 5 m thick and has an electrical conductivity of 60 mS/m. This is followed by a transition zone also 5 m thick, whereafter the electrical conductivity stabilises at about 270 mS/m. The profile is interpreted to reflect good-quality, recently recharged groundwater overlying brackish groundwater discharged from the lake. This interpretation was supported by the hydrochemical character of the different waters (Figures 4.12 and 5.16) as well as isotope data addressed in Section 5.7.4.

Profiling of GRU2 and GRU3 did not reveal the same stratification (Figure 5.18) [2]. In the upper parts the salinity was similar to that expected elsewhere in the Eden Primary Aquifer, but with a gradual increase in EC in the direction of flow toward the sea. The stratification gradient in GRU2 was very marked while that in GRU3 was more gradual, with these

[1] The first profile conducted in March 2006 was soon after the borehole was drilled, and before the drilling mud had been completely removed and the borehole properly developed.

[2] The profile presented in Figure 5.18 was undertaken using an uncalibrated EC meter. The data should not be compared with other EC measurements, but a comparison of the three profiles in the graph is valid.

differences possibly reflecting local flow conditions. The final measurements of each profile suggest an increased degree of mixing of lake water and groundwater along the flow path to the sea. It is probable groundwater discharging into the sea south of Groenvlei is completely mixed and has an EC in the order of 130 mS/m – 150 mS/m (Figure 5.19).

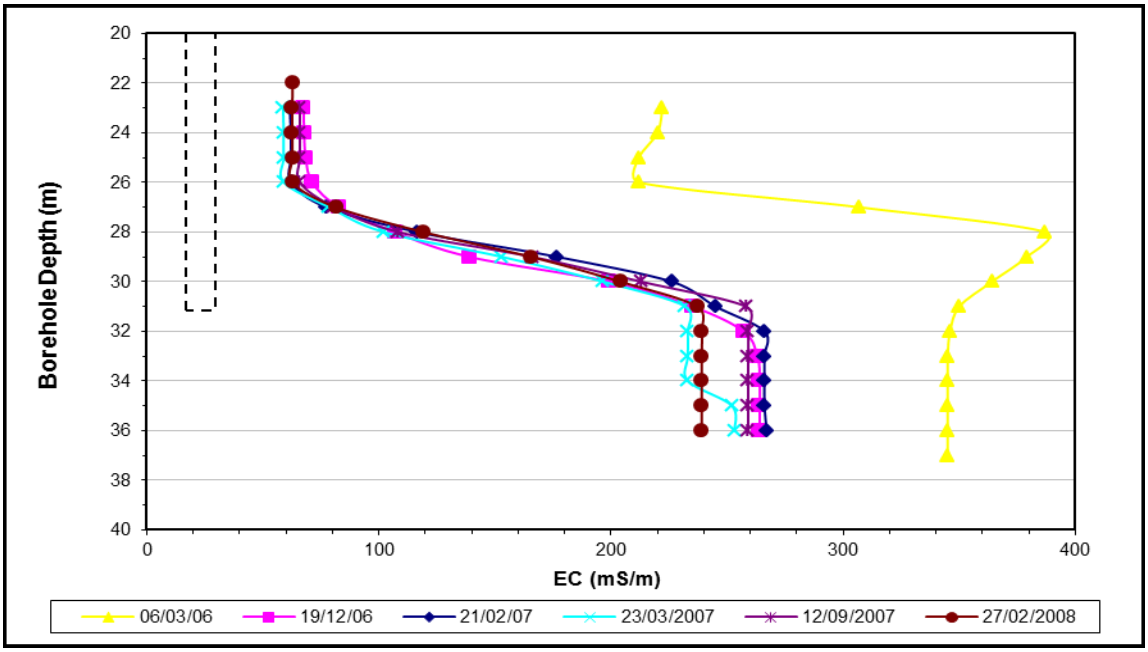

Figure 5.17: Salinity profiles of GRU1. The stippled lines demonstrate the position of the screen.

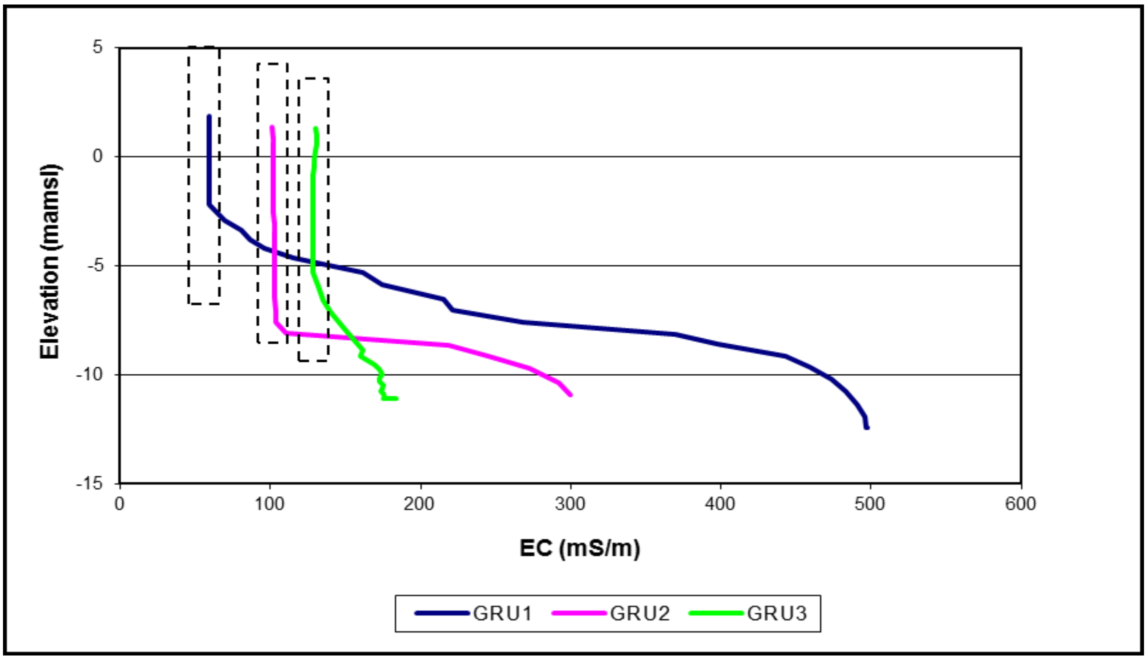

Figure 5.18: Salinity profiles of the three GRU boreholes positioned between Groenvlei and the sea. The stippled lines demonstrate the position of the screen in each borehole.

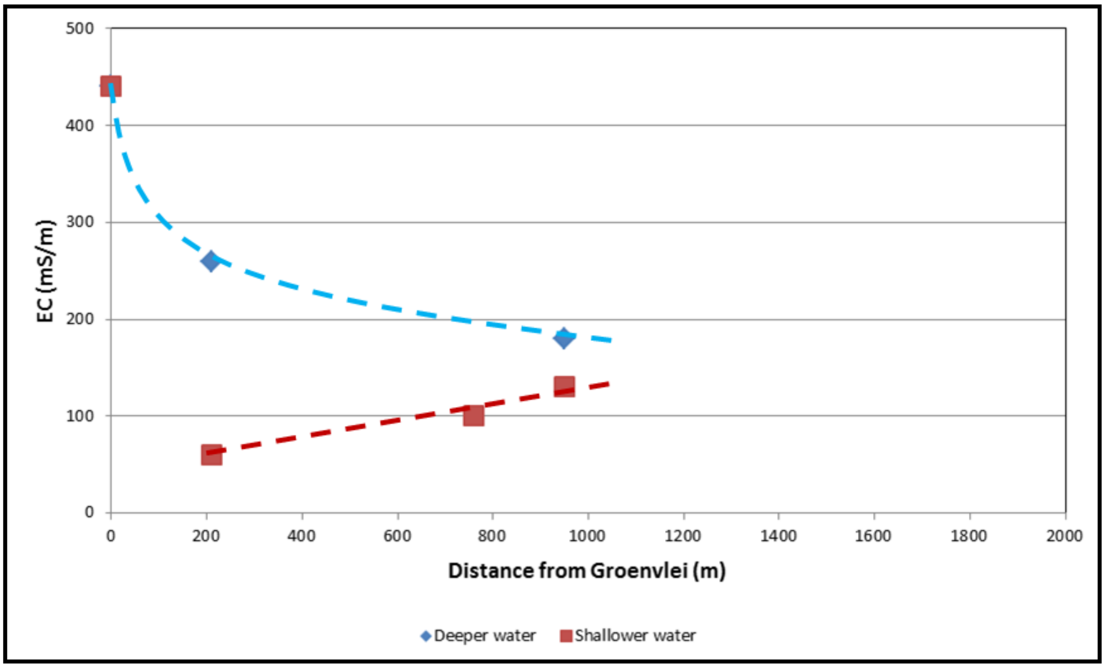

Figure 5.19: Change in EC with distance from Groenvlei in shallow and deeper groundwater. Both data sets indicate dilution of brackish lake water and mixing of shallow and deeper groundwater.

In the analysis of the EC profile data, it is acknowledged that the screens in GRU1, GRU2 and GRU3 may not reach a depth where the salinity of the deeper groundwater has stabilised. The ramification of this is that the EC in GRU1, for example, may be 440 mS/m at a depth of 40 m below ground level. Notwitstanding this limitation, the profiling clearly establishes the stratified groundwater quality south of Groenvlei.

5.7.3 Groundwater quality west of Groenvlei

Sampling of boreholes and wellpoints directly west of Groenvlei revealed a zone at least 200 m wide in which groundwater changed from its ambient salinity (100 mS/m) to that of Groenvlei (440 mS/m) (Figure 5.20). In this zone the hydraulic gradient is clearly towards the lake (i.e. groundwater discharges into the lake), but is relatively flat. WM1 is located 35 m from the edge of the lake and has a salinity similar to that of the lake. Electrical conductivity then rapidly reduces to 100 mS/m over the next 100 m, after which it remains constant with increasing distance from the vlei.

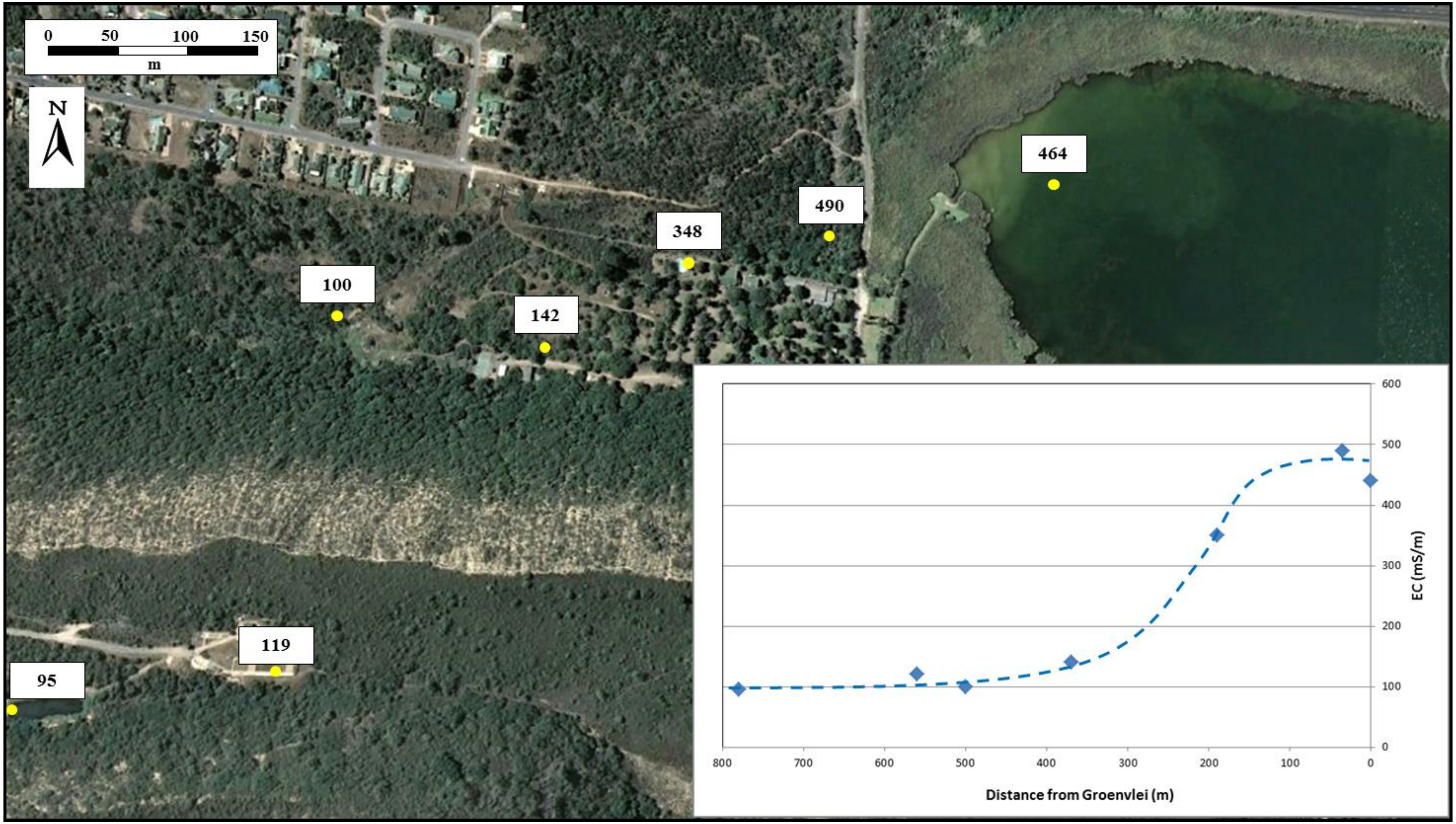

Figure 5.20: The salinity profile of the north western corner of Groenvlei, showing a relatively broad zone of transition where the salinity of water changes from that of the lake to that of groundwater.

This relatively broad zone of transition may be the result of diffusion of salts from the more saline lake to the lower salinity groundwater. However, it is possible the relative change of water level in the lake and groundwater causes a temporary and localised reversal of flow directions. In compiling Figure 5.8, it was interpolated MWP4 was at an elevation of 5 mamsl. However, if the wellpoint is at a level of just 5 cm lower (i.e. 4.95 mamsl), then a situation as depicted in Figure 5.21 could occur. In most instances, the groundwater level at MWP4 is higher than that of the lake. Toward the end of winter or after significant rain when the level of the lake rises faster than groundwater, then water could flow from the lake into the aquifer. During this time brackish water would be transported from the lake into the groundwater fringing the surface water body. Reversal of flow would persist until the level of the lake started to fall with the onset of summer and increased evapotranspiration.

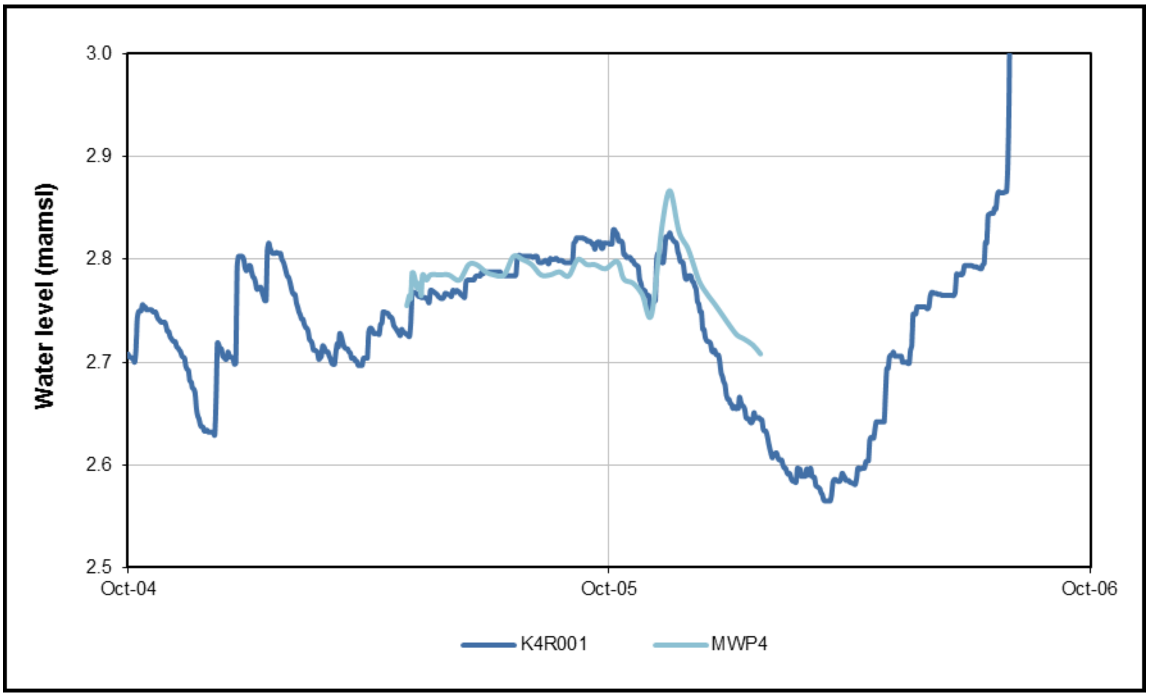

Figure 5.21: Assuming the elevation of MWP4 was 4.95 mamsl (as opposed to 5 mamsl used in Figure 5.8), flow from the lake into the aquifer can occur when the lake level rises (a) in response to rainfall or (b) as a result of reduced evapotranspiration losses in winter.

Monitoring of three wellpoints directly east of Groenvlei by Roets (2008) did not reveal the same zone of transition. However, the wellpoints are 560 m from the perimeter of the lake and thus unlikely to be influenced by the postulated flow reversals. The validity of the model that relative changes in water level result in localised changes of groundwater flow direction can only be tested by detailed and accurate water level monitoring.

5.7.4 Stable isotope data

Comparison of stable isotopes in groundwater in the vicinity of Groenvlei and in the lake provided support for interpreted directions of groundwater flow.

5.7.4.1 Plot of δD against $\delta^{18}O$

The commonly plotted stable isotope graph indicated lake water plots in a narrow field and had a distinct evaporative signal (Figure 5.22). The easternmost samples (GROE15, GROE17 and GROE18) had the least enriched signals of the surface water samples, reasons for which are not clear.

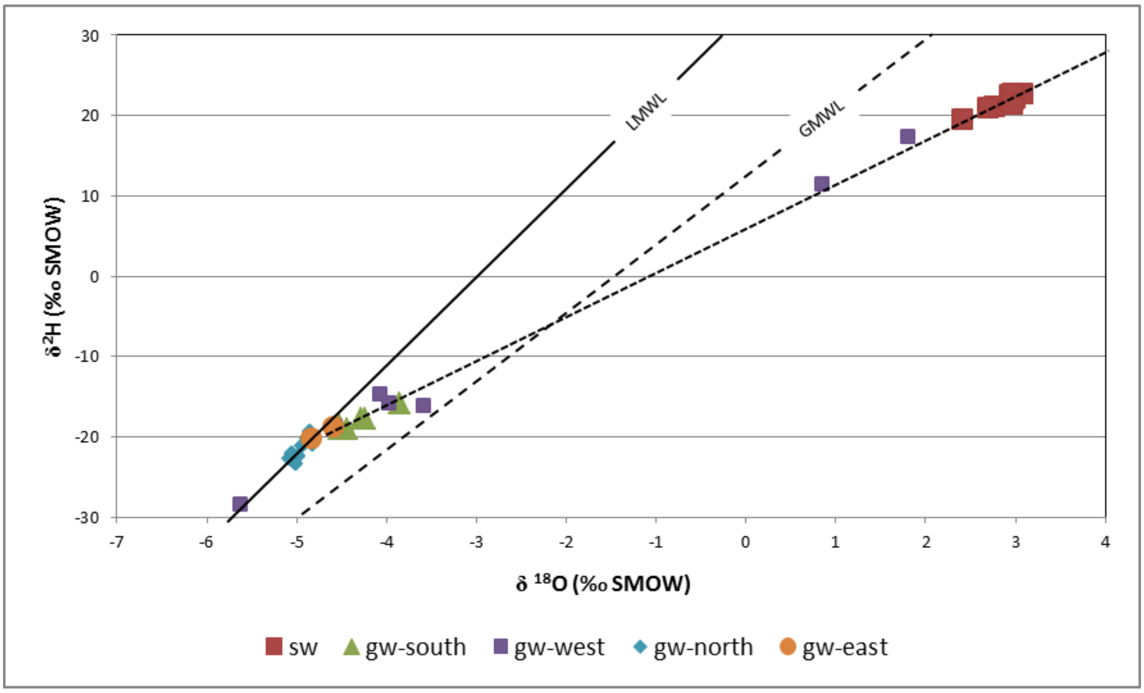

Figure 5.22: A plot of δD against δ18O, highlighting the evaporative signature of surface water. The global meteoric water line (GMWL) and interpreted local meteoric water line (LMWL) are also shown.

Groundwater samples taken north of Groenvlei probably best represented the isotopic signature of groundwater in the area. However, the signature was slightly more enriched with lighter isotopes when compared to the Cape Flats Aquifer (Harris et al., 1999), with the difference probably the result of isotopic differences in rain that recharged the two systems and the topographic effected caused by the 200 m high dunes between Groenvlei and the sea.

Samples taken from wellpoints at GR/W/UA plotted near that of surface water. These wellpoints were in fact in the lake, and like WM1 nearby, had elevated salinities. This may reflect a build-up of salts in shallow groundwater on the fringe of the lake, an issue discussed in greater detail in Section 6.3.

Both LP6 and WM1 – earlier interpreted to have been influenced by surface water – showed signs of δD and ^{18}O enrichment, but were not distinctly different from other groundwater samples. It was only with close examination that the evaporative characteristics were apparent (Figure 5.23). Similarly, GRU1 also showed signs of enrichment of heavier isotopes, with the deeper samples being slightly more enriched than the shallower ones because of evaporation from the lake.

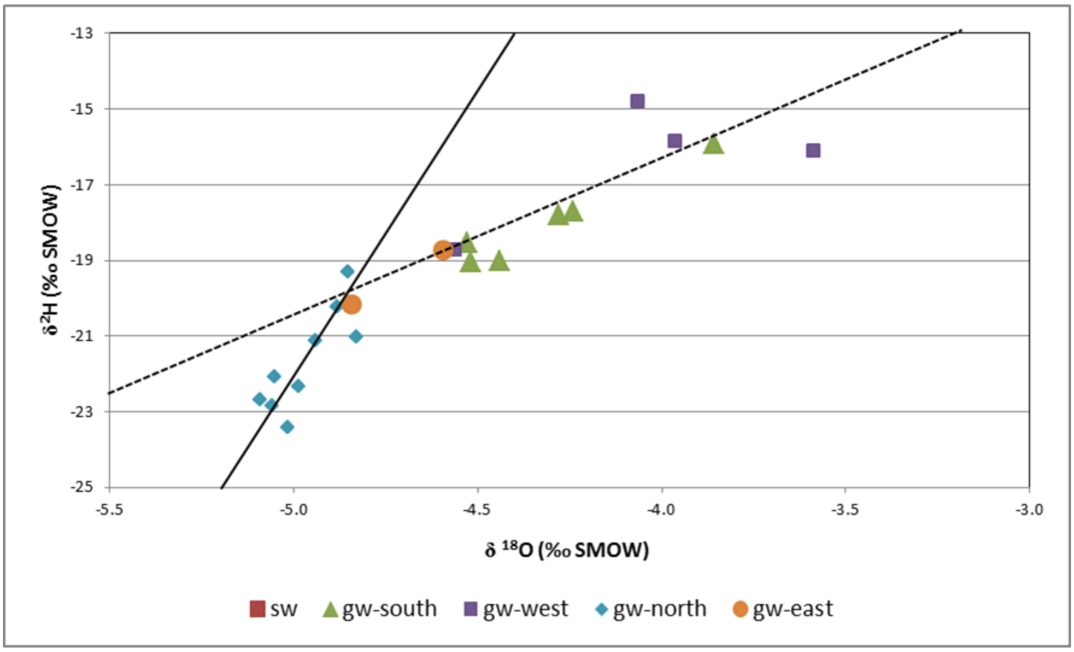

Figure 5.23: An expanded view of Figure 5.22 showing the isotopic signature of groundwater north of Groenvlei and the fractionation caused by evaporation to groundwater west and south of the lake.

5.7.4.2 Plot of δD vs Cl

The brackish character of surface water allowed it to be readily identified using EC and most of the major ions. A plot of the concentration of δD against that of Cl also highlighted the difference between the two waters (Figure 5.24), with groundwater being depleted with respect to δD and surface water being enriched. The graph displays similar characteristics to that of Figure 5.22 with the influence of surface water on groundwater west and south of Groenvlei being apparent.

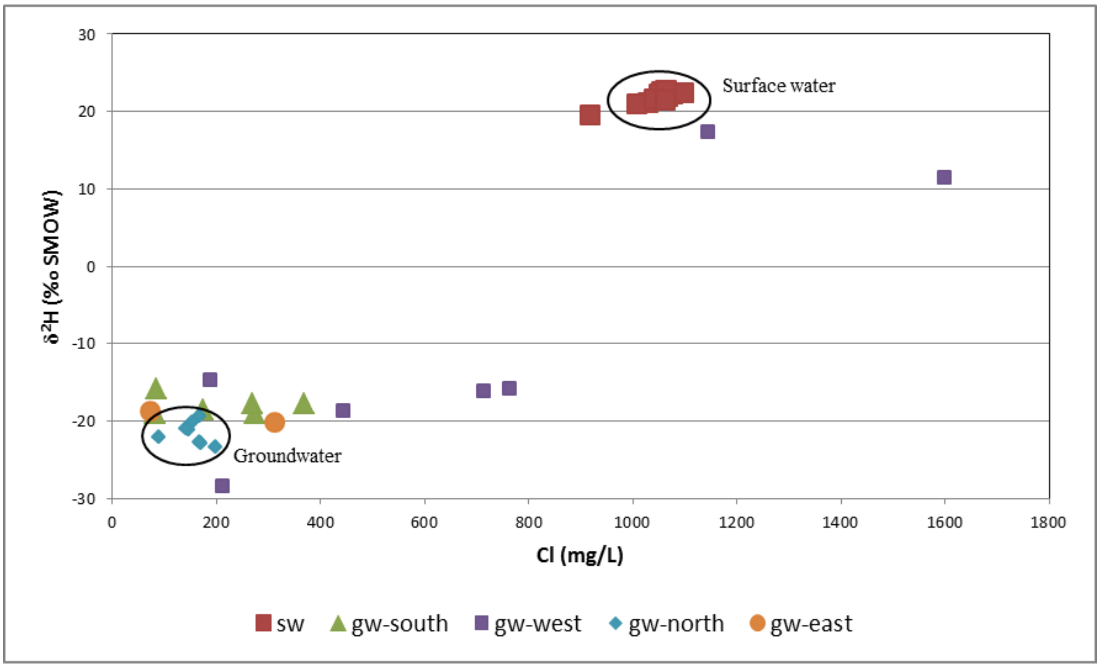

Figure 5.24: A plot of δD against Cl, highlighting the different characteristics between surface water and groundwater and the influence of surface water on groundwater west and south of the lake.

5.7.5 Radon data

Distinct differences in ^{222}Rn concentrations were observed in the surface water and groundwater (Figure 5.25). Water sampled from the lake was mostly devoid of ^{222}Rn, but concentrations of less than 0.5 Bq/L were found in places. It was not possible to determine an average concentration for groundwater as once samples collected by bailing, and those known to be impacted by surface water, were removed from the record, too few samples remained to

Figure 5.25: Concentration of 222Rn found in surface and groundwater in and around Groenvlei.

provide meaningful statistics. However, it would appear that the average concentration of ^{222}Rn in groundwater is between 8 Bq/L and 12 Bq/L. The use of ^{222}Rn as a tracer at Groenvlei did not yield any new information. However, application of this technique requires further consideration in terms of sampling protocol and positioning of the sampling points.

5.8 Groundwater Use

Groundwater use in the immediate vicinity of Groenvlei is limited to three boreholes used for domestic supply and two wellpoints used for garden irrigation. Total groundwater use was estimated at 4 000 m^3/a [1].

Most houses in and around Sedgefield located at an elevation of 10 mamsl or less make use of spikes for garden irrigation. These are not shown on Figure 5.26. Parsons (2003) estimated this groundwater use could amount to 275 000 m^3/a - or 35% of the municipally supplied water. In recent years, deeper boreholes have been drilled in elevated areas such as Cola Beach, and the private use of groundwater may have increased to, say, 350 000 m^3/a. All of this groundwater use falls outside Groenvlei's capture zone, and it is improbable it could impact the hydrology of the surface water body. Only about 40 houses are located within Groenvlei's capture zone, but few would use groundwater as the depth to groundwater is at the limit of a centrifugal pump's lifting ability.

Knysna Municipality established 8 production boreholes in the village for municipal water supply purposes (Parsons, 2013b). These boreholes are used to boost water supplies during the peak holiday periods of April, December and January. Assuming a pumping regime of 75 days a year, total municipal groundwater abstraction amounts to 100 000 m^3/a. The municipal production boreholes are distant from Groenvlei (2 km), outside the lake's capture zone, and have a limited radius of influence (Figure 5.7). Consequently, the municipal production boreholes have no influence on the hydrology of Groenvlei.

Two high-yielding boreholes have been established in the Myoli Beach car park to provide feedwater to the desalination plant (Figure 5.26). These boreholes are 3 km from Groenvlei and also do not influence the hydrology of the lake.

[1] This equates to 0.003 mm/d of the inflow into Groenvlei.

Figure 5.26: Groundwater use in the immediate vicinity of Groenvlei.

6 QUANTIFYING GROUNDWATER'S CONTRIBUTION TO GROENVLEI

6.1 Daily Water Balance

6.1.1 Adaption of water balance for Groenvlei

No surface water drainage features drain into Groenvlei. Inspection of rainfall and water level records showed that, in most instances of high rainfall, the rise in water level in the lake was equivalent to depth of rainfall. For example, an exceptional rainfall of 257 mm over two days from 22 November 2007 resulted in a rise in water level of 240 mm (Figure 6.1). If run-off were to contribute to Groenvlei, it would be expected the rise in water level would be significantly greater than the depth of rainfall, and this was not the case. A plot of rainfall events and rises in water level of more than 40 mm indicated a relationship where the rise was roughly equivalent to rainfall (Figure 6.2). No bias towards rise in water level being greater than rainfall was observed. The two exceptions to the 107 events investigated probably relate to errors in recording rainfall.

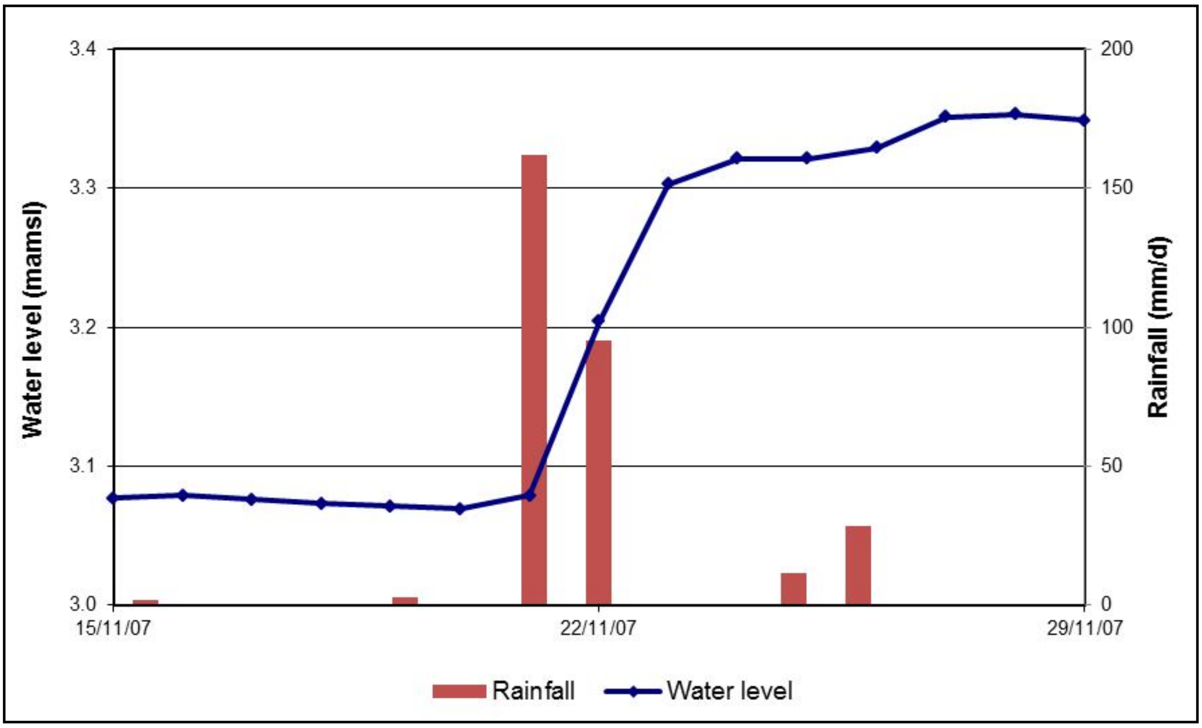

Figure 6.1: Rainfall of 257 mm over 2 days in late November 2007 caused a rise in lake water level of 240 mm.

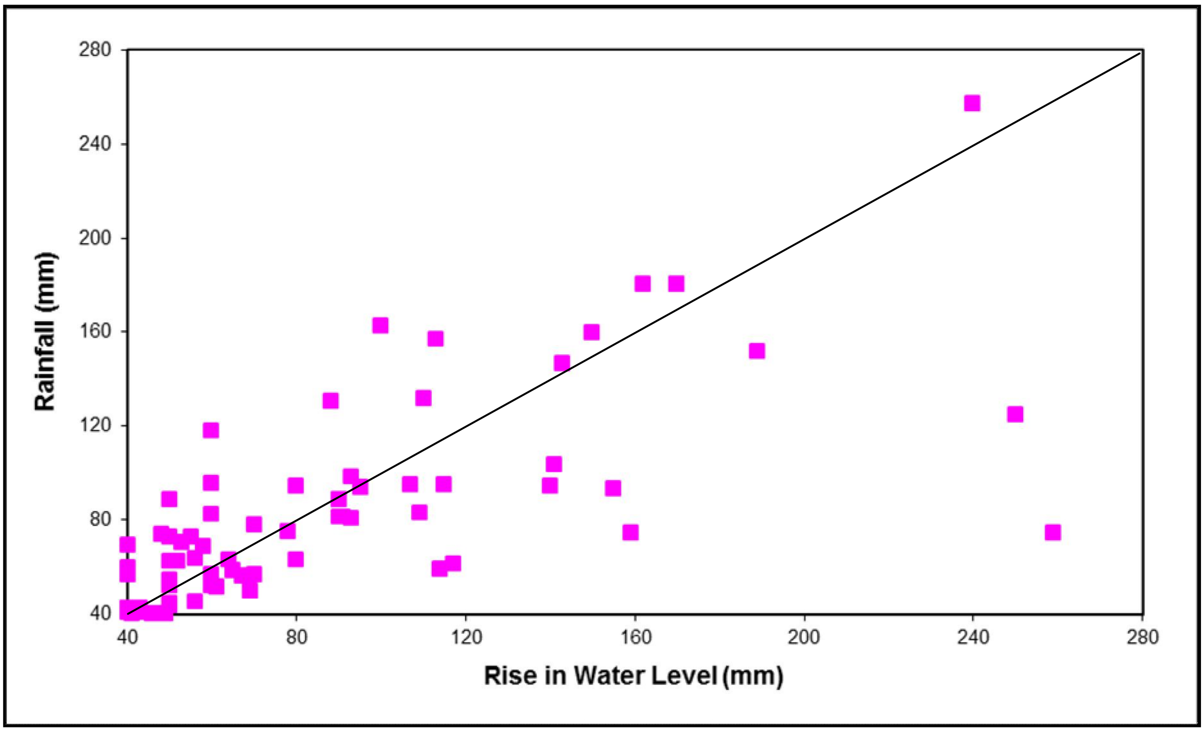

Figure 6.2: Comparison of rainfall and rise of water level observed between 1981 and 2012.

In light of these observations, the run-off component was removed from the water balance equation presented in Section 2.5.2 and simplified as follows:

$$(\mathbf{P} + \mathbf{G_i}) - (\mathbf{E} + \mathbf{G_o} + \mathbf{A}) = \boldsymbol{\Delta}\mathbf{s}$$

Water abstraction from Groenvlei has not been quantified, but is expected to be both limited and insignificant. The quality of water is such that it is not fit for domestic consumption and most water quality variables exceed the DWAF (1998) and SANS (2006) standards for drinking water (see Section 4.6). The number of potential users of water from Groenvlei is very small. To all intents and purposes, the abstraction term can also be removed from the water balance equation:

$$(\mathbf{P} + \mathbf{G_i}) - (\mathbf{E} + \mathbf{G_o}) = \boldsymbol{\Delta}\mathbf{s}$$

In this water balance, two unmeasured components remain, namely groundwater inflow and groundwater outflow. By rearranging the water balance and introducing a nett groundwater contribution component (i.e. groundwater inflow less groundwater outflow), a single unknown remains. As the other three components are measured on a daily basis, it is possible to solve the equation:

$$(G_i - G_o) = \Delta s - P + E$$

The conceptual understanding of the hydrology of Groenvlei is captured in this equation, with water level being a function of precipitation and groundwater inflow, and evaporation and groundwater outflow.

6.1.2 Application of water balance model

6.1.2.1 Excel spread sheet

A daily water balance model was constructed using an Excel spreadsheet. It was assumed the nett groundwater contribution remained constant over time, allowing a facility to be built into the spreadsheet that enabled the modelled water level to be matched to the monitored lake level data by simply changing the nett groundwater contribution. Such curve-matching techniques are widely applied in geohydrological studies, particularly in the interpretation of pumping test data (Driscol, 1986; Kruseman and de Ridder, 1990 and van Tonder et al. 2004). Previously, Parsons (2008a) used the same approach to estimate the nett groundwater contribution to Groenvlei to be 2.01 mm/d (Figure 6.3). He used A pan evaporation data to approximate lake evaporation. However, other approaches are documented in the literature (see Section 2.5.2.5) and it was decided to apply two other commonly-used methods to estimate lake evaporation:

- Multiplying S pan data by lake factors provided by Midgley et al. (1994); and
- Multiplying A pan data by 0.7.

For ease of reference, application of the Midgley et al. (1994) approach is referred to as Model 1, multiplying A pan data by a factor of 0.7 is Model 2, and use of A pan data to approximate lake evaporation is referred to as Model 3.

The impact of the lake evaporation model used on the water balance is best illustrated by considering the range of annual lake evaporation estimates obtained from the different models. Evaporation losses estimated from S pan and A pan data (Model 3) amounted to 1 028 mm/a and 1 435 mm/a respectively. Lake evaporation approximated using Model 2 amounted to 1 005 mm/a, while losses estimated using Model 1 amounted to 866 mm/a. This represents a range of 569 mm/a or 55% of the S pan measurement. Consequently, quantification of evaporation losses emerged as a key aspect of this research.

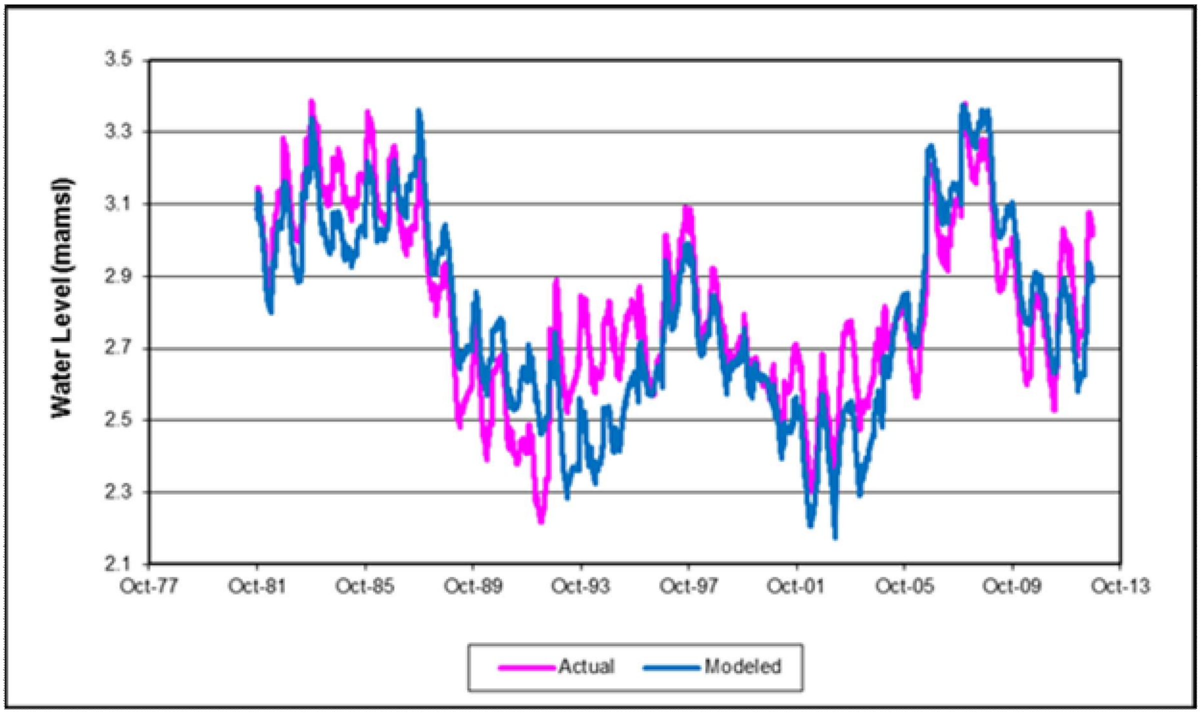

Figure 6.3: Measured water level compared to that determined using a daily water balance model. It was assumed lake evaporation was equivalent to A pan evaporation and nett groundwater contribution was set at 2.01 mm/d.

Using Model 1, the nett groundwater contribution between 1981 and 2012 was determined to be 0.52 mm/d (Figure 6.4). Model 2 yielded a nett groundwater contribution of 0.89 mm/d (Figure 6.5). The three different approaches – including that using A pan data without

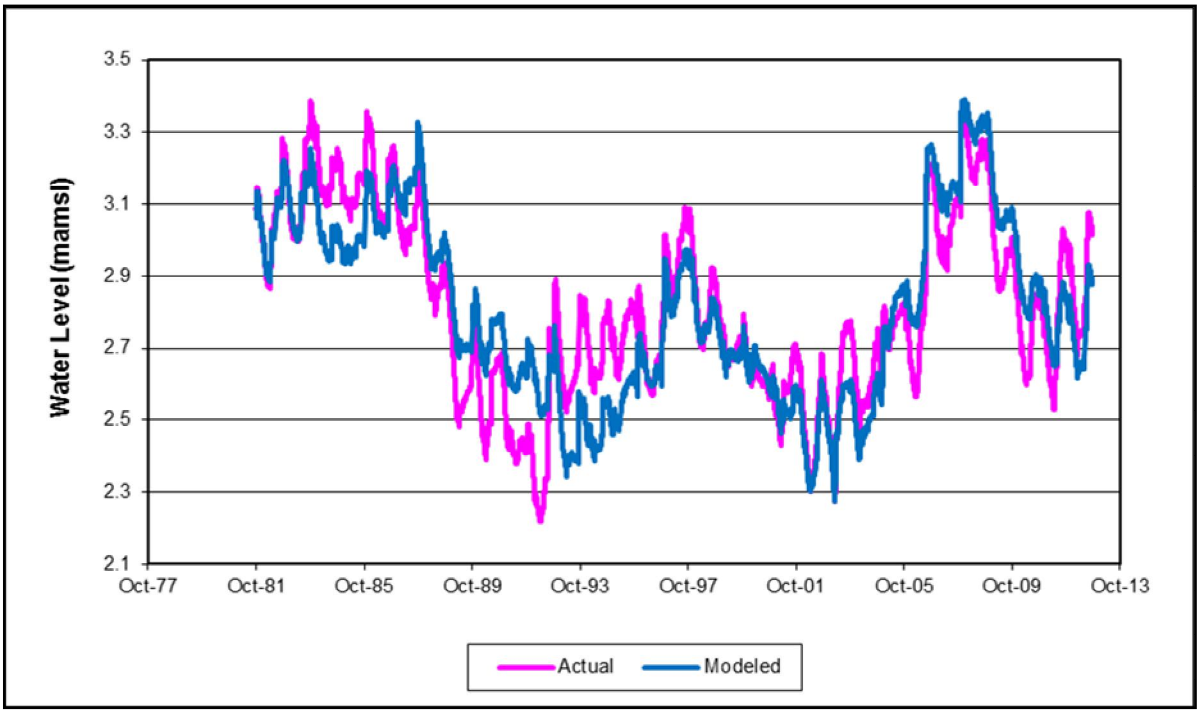

Figure 6.4: Measured water level compared to that determined using a daily water balance model. Lake evaporation was determined using Model 1 and the nett groundwater contribution was set at 0.52 mm/d.

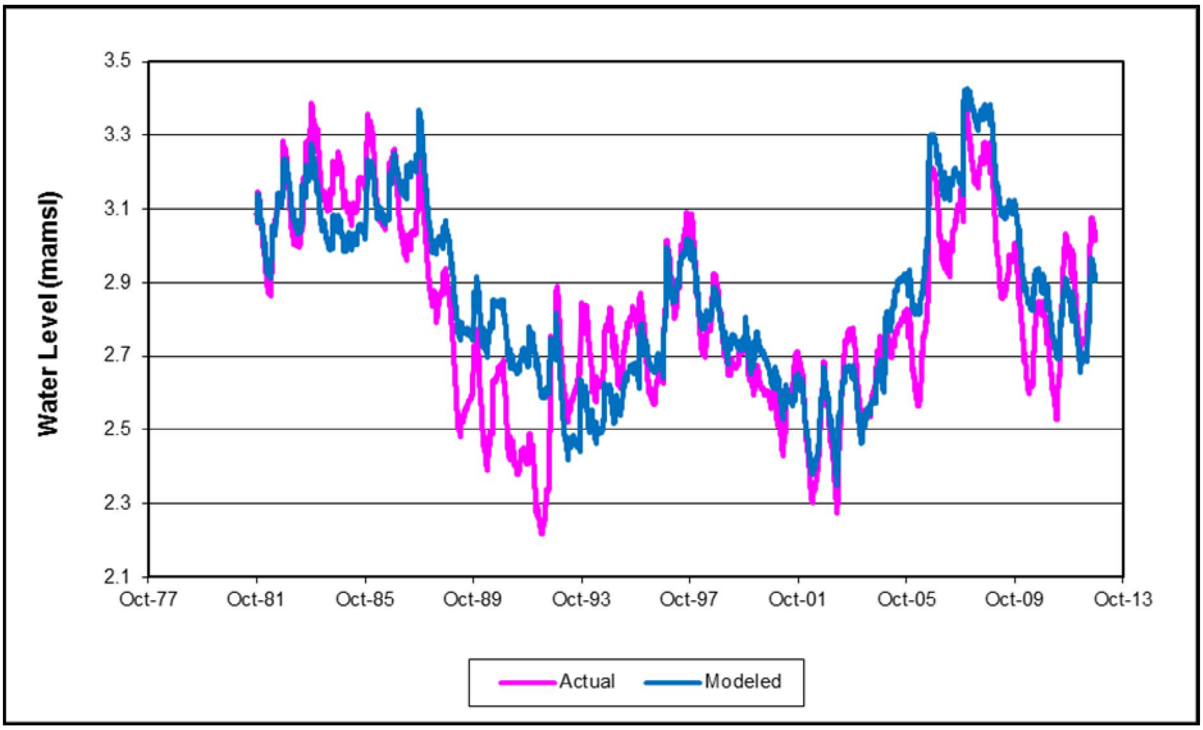

Figure 6.5: Measured water level compared to that determined using a daily water balance model. Lake evaporation was determined using Model 2 and the nett groundwater contribution was set at 0.89 mm/d.

adjustment (Model 3) – all yielded reasonable results in modelling lake levels over a 31 year period. While applying the daily water balance model, three challenges immediately became apparent:

- Missing and incorrect data in the rainfall and evaporation record;
- Understanding differences in summer and winter; and
- Quantifying lake evaporation using measured pan data.

6.1.2.2 Constant nett groundwater contribution

Before addressing the above challenges, it is prudent to consider whether the assumption that the nett groundwater contribution remained constant through time was reasonable, as it could be a source of error and / or account for differences in summer and winter groundwater contributions. It was calculated the overall hydraulic gradient between Groenvlei and the sea varies by 34% when the lake is full (3.39 mamsl) and at its lowest level (2.22 mamsl). Seasonally, lake levels vary by 0.2 m between the summer low and winter high. This amounts to a variance of 7% in the hydraulic gradient. The maximum observed variance in groundwater levels in the three boreholes between Groenvlei and the sea (GRU1, GRU2, GRU3) during the period October 2009 and April 2013 ranged between 0.12 m and 0.18 m. The equivalent maximum observed variance in lake level during the same period was 0.46 m. It was thus interpreted any change of hydraulic gradient occurs immediately downgradient of Groenvlei, and not over the entire distance to the sea.

The maximum groundwater level variation upgradient of Groenvlei (DOD1a) in the period October 2009 and April 2013 was 0.13 m i.e. similar to that observed downgradient of the lake. It was concluded even though groundwater levels may go up and down by about 0.15 m, the hydraulic gradient essentially remains the same. Because of the steeper hydraulic gradient immediately downgradient of Groenvlei when full (say July to December), the nett groundwater contribution could be up to 10% higher when compared to that of January to June. Given the relatively small difference between these two periods, the assumption that nett groundwater contribution remains constant appears reasonable.

6.1.2.3 Missing and incorrect data

The procedure followed by DWA when monitoring water levels at K4R001 (Section 3.1.2) suggests few data errors are expected in the water level record of Groenvlei. As a result, the water level record was one of the tools used to check the rainfall record i.e. if a rise in water level was not accompanied by a similar order of magnitude rainfall event (and *vice versa*), then steps were taken to verify the recorded rainfall. This was done by comparing rainfall at K3E003 to that of K4E001 (01 October 1981 – 30 June 1987) or that of the Fire Station (01 January 2009 – 30 September 2012). If there was good evidence that the rainfall recorded at K3E001 was wrong, then rainfall was added to the record (Table 6.1). The added rainfall increased the average rainfall of 653 mm/a presented in Section 4.2 by 18 mm/a or 2.7%. It is recognised the rainfall record almost certainly still contains errors difficult to identify.

Table 6.1: Instances of rainfall added to the rainfall record

Date	Rainfall (mm/d)	Justification
20/07/84	70	Significant rise in water level, period of rain
11/10/85	90	Rainfall compares with that from other rainfall stations, but added 90 mm to improve matching of graphs
15/10/92	130	No rain recorded at K3E003, but almost 100 mm at K4E001, also big rise in water level
12/04/93	60	No rain recorded at K3E003, but almost 60 mm at K4E001, also big rise in water level
19/09/93	130	No rain recorded at K3E003, but big rise in water level, 37 mm recorded at K4E001
29/11/93	60	No rain recorded at K3E003, but big rise in water level, 27 mm recorded at Mossel Bay
19/04/94	60	No rain recorded at K3E003, but big rise in water level, 21 mm recorded at K4E001
05/01/98	50	No rain recorded at K3E003, but almost 50 mm at K4E001, also big rise in water level
01/01/07	-90	Removed 97 mm from record based on absence of rise of water level

The procedure applied when patching the S pan evaporation data is described in Section 3.2.2. Using average monthly data, it was possible to make a synthetic record of evaporation that matched the measured data. This proved useful in the patching process as it allowed easy identification of missing data and highlighted differences in the data before and after October 2006 (Figure 6.6). This date coincides with that when data was sourced from DWA. While the cause of the differences has not been ascertained nor any deviations from patterns evident in the pre-2006 record identified, the difference is noted.

A key issue with the missing or incorrect data was the introduction of uncertainty in the modelling and how errors would impact the modelling from that point on. It was hence decided to model for periods of a hydrological year, and thereby prevent errors in one particular year being transferred into subsequent years. Modelling for the entire 31 year period resulted in errors being included in the record and masked through averaging. The models were hence reset at the start of each hydrological year by setting the modelled water level equal to that measured on 1 October.

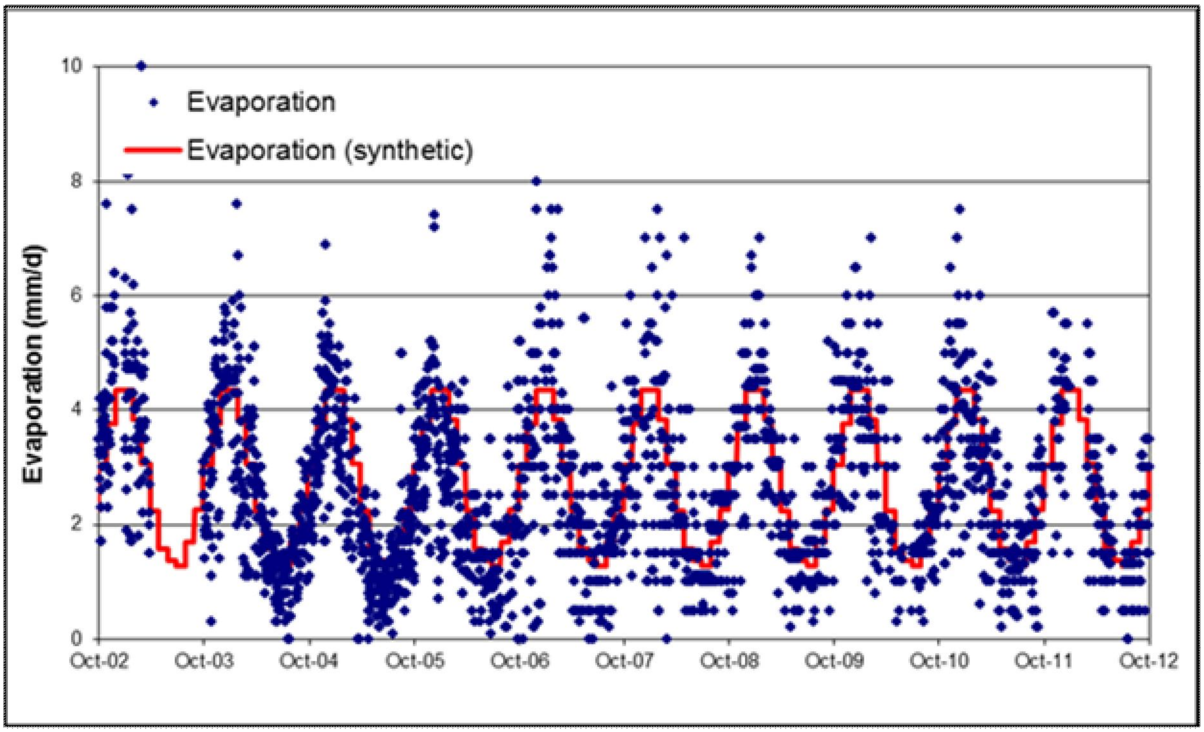

Figure 6.6: Evaporation data sourced from DWA after 2006 displays patterns different to that data sourced from DWA in 2006. Also, the synthetic record allowed for the easy identification of missing data, as was the case in early 2003.

6.1.2.4 Differences in summer and winter

When researching different methods of quantifying lake evaporation from pan data, it was consistently experienced that the modelled data could be matched to the observed data in summer, but not in winter. This is well illustrated in Figure 6.7. Different coefficients had to be applied to the evaporation data in summer and winter to get the observed and modelled data to match using the same nett groundwater contribution. Instead of using 0.88 in the winter as prescribed by Midgley et al. (1994) (Table 2.5), a winter pan coefficient of 0.50 was required to get a reasonable result when using Model 1. The summer coefficient remained at the prescribed 0.81. Similarly, applying a coefficient of 0.7 to A pan data also resulted in the winter match being poor, with the coefficient used in winter being too high. Application of a coefficient of 0.54 to the winter data yielded a better match of observed and modelled lake water levels.

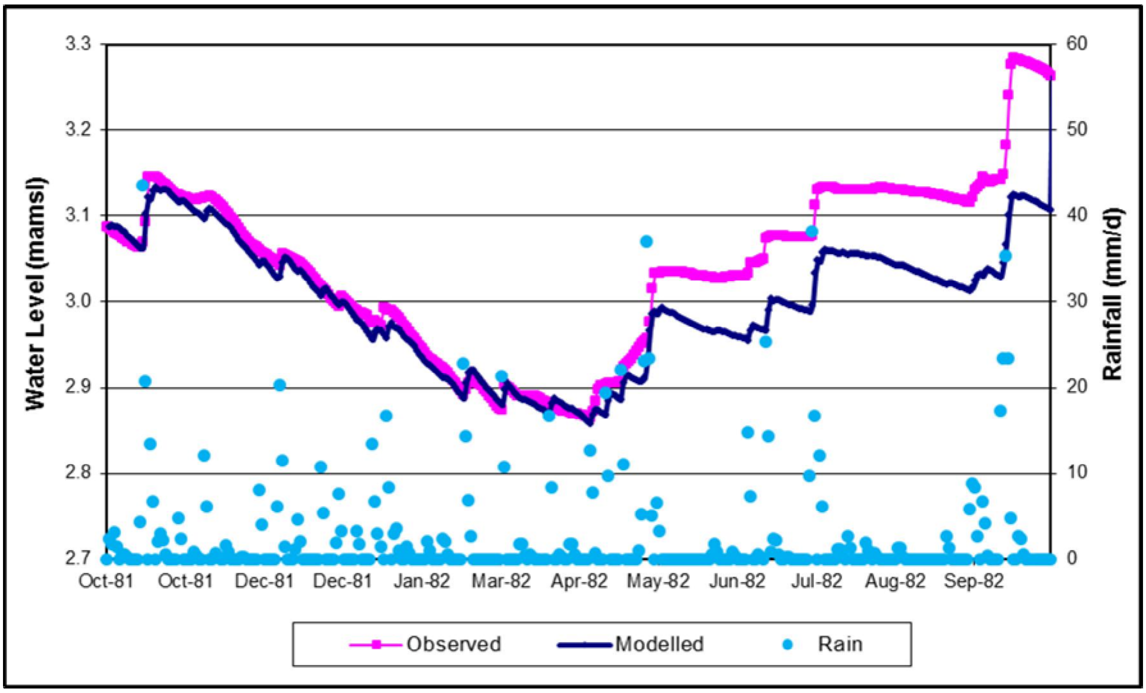

Figure 6.7: Demonstration of the inability to model the water level during the latter part of the year. In this instance, lake evaporation was determined using Model 2 and applying a nett groundwater contribution of 0.6 mm/d.

Research was undertaken to understand the differences between summer and winter evaporation patterns. Seasonal time scales were used to determine summer, early winter and late winter evaporation coefficients for the three methods of quantifying lake evaporation (Table 6.2). The results, displayed in Figures 6.8, 6.9 and 6.10, illustrate clear differences between summer and winter. The biggest differences between summer and winter were obtained with Model 1 and Model 2, while Model 3 yielded bigger nett groundwater contributions. No obvious relationship was found with the water level of the lake. While possible to get good matches between observed and modelled water levels using seasonal coefficients, the reason for the seasonal differences did not emerge from the analyses. Also, similar observations were not reported in the literature.

Table 6.2: Results of modelling using a daily water balance model of Groenvlei to determine the nett groundwater contribution.

Year Ending	Model 1						Model 2						Model 3					
	summer (mm/d)	match rating	A - J (mm/d)	match rating	J - S (mm/d)	Match Rating	Summer (mm/d)	match rating	A – J (mm/d)	match rating	J – S (mm/d)	match rating	summer (mm/d)	Match Rating	A – J (mm/d)	match rating	J - S (mm/d)	match rating
01/10/1982	0.40	5	0.85	4	0.85	5	0.60	5	1.30	4	1.30	5	2.45	5	2.30	4	2.20	4
01/10/1983	-0.40	3	1	4	1	4	0.20	4	1.50	3	1.50	3	1.40	4	2.00	3	2.40	3
01/10/1984	0.60	4	0.95	5	0.95	5	0.75	5	1.20	4	1.20	4	2.40	5	2.40	5	2.40	5
01/10/1985	0.00	4	0.55	4	0.55	4	0.35	4	0.90	5	0.90	5	1.70	3	1.70	3	1.70	3
01/10/1986	0.00	4	0.4	4	0.5	4	0.15	4	1.00	5	1.00	5	1.80	5	1.80	5	1.80	5
01/10/1987	-0.50	4	0.7	4	0.7	4	-0.25	4	1.10	4	1.20	4	1.10	5	1.50	3	2.20	2
01/10/1988	0.40	5	1	3	0.6	5	0.65	5	1.60	5	1.60	5	2.25	5	2.25	4	2.25	5
01/10/1989	-0.15	5	1.1	5	1.1	5	0.20	5	1.55	5	1.55	5	1.75	4	2.30	5	2.30	5
01/10/1990	-0.40	5	0.9	5	0.9	5	-0.20	5	1.30	5	1.30	5	1.35	5	2.05	5	2.05	5
01/10/1991	0.00	3	0.4	3	0.4	3	0.25	4	0.60	4	0.65	4	2.00	4	1.35	4	1.35	4
01/10/1992	-0.10	3	1.45	4	1.45	4	0.35	3	1.70	3	1.70	3	1.00	3	2.50	3	2.50	3
01/10/1993	0.10	3	0.95	4	0.95	4	0.40	2	1.40	3	1.50	3	2.30	3	2.30	3	2.30	3
01/10/1994	0.60	3	0.6	3	0.6	3	0.50	3	1.25	5	1.25	5	2.05	3	2.05	5	2.05	5
01/10/1995	-0.20	4	0.85	5	0.85	5	0.15	4	1.30	5	1.30	5	2.00	3	2.00	5	2.00	5
01/10/1996	-0.40	4	0.35	5	0.7	5	-0.20	3	0.90	4	1.20	4	1.60	4	1.55	5	2.00	4
01/10/1997	0.50	4	0.75	4	0.75	4	0.90	3	1.20	4	1.30	4	2.40	4	2.40	4	2.40	4
01/10/1998	-0.05	5	0.8	5	0.8	4	0.25	5	1.20	4	1.30	3	1.85	4	1.85	4	2.10	3
01/10/1999	0.00	5	0.8	4	0.8	4	0.45	4	1.15	3	1.25	3	1.90	4	1.95	4	1.95	4
01/10/2000	0.30	4	0.4	4	0.5	4	0.55	5	0.85	5	1.00	5	1.75	3	1.80	4	1.80	4
01/10/2001	0.50	3	1	4	1	4	0.70	4	1.25	4	1.25	4	2.30	4	2.30	4	2.30	4
01/10/2002	-0.10	5	0.65	4	0.65	4	0.20	5	1.00	5	1.20	5	1.85	5	1.85	5	1.85	5
01/10/2003	0.00	5	1.1	4	1.1	4	0.30	5	1.50	3	1.50	3	1.90	5	2.40	3	2.40	4
01/10/2004	-0.10	5	0.6	5	0.6	5	0.30	5	0.95	4	1.20	5	1.60	5	1.90	5	1.90	5
01/10/2005	-0.10	4	0.4	5	0.4	5	0.15	4	0.20	4	1.15	5	1.25	3	1.60	5	1.60	5
01/10/2006	-0.45	4	0.3	3	0.3	3	0.28	3	1.40	3	1.60	3	1.30	4	1.30	3	1.60	4

01/10/2007	0.10	4	1	4	1.2	4	0.40	4	1.30	5	1.30	5	1.60	4	2.10	4	2.30	4
01/10/2008	-0.20	3	0.8	5	0.8	5	0.10	4	1.15	5	1.15	5	1.40	3	2.00	5	2.10	5
01/10/2009	-0.05	4	0.85	4	0.85	4	0.20	4	1.10	5	1.20	5	1.80	4	1.80	4	2.00	5
01/10/2010	-0.05	4	1.2	4	1.2	4	0.40	5	1.50	4	1.65	4	1.90	4	2.40	4	2.40	4
01/10/2011	0.00	4	0.6	4	0.6	4	0.45	5	1.40	3	1.40	3	1.80	5	2.30	3	2.30	3
01/10/2012	-0.40	4	0.5	4	0.5	4	0.15	4	1.10	4	1.10	4	1.50	4	1.75	4	1.75	4
Average	0.00		0.77		0.78		0.31		1.19		1.28		1.78		1.99		2.07	
Std dev.	0.31		0.28		0.27		0.26		0.31		0.23		0.38		0.33		0.29	
Min.	-0.50		0.30		0.30		-0.25		0.20		0.65		1.00		1.30		1.35	
Max.	0.60		1.45		1.45		0.90		1.70		1.70		2.45		2.50		2.50	

Note:

A rating was given to each curve matching using scores of 1 (no match) to 5 (very good match).

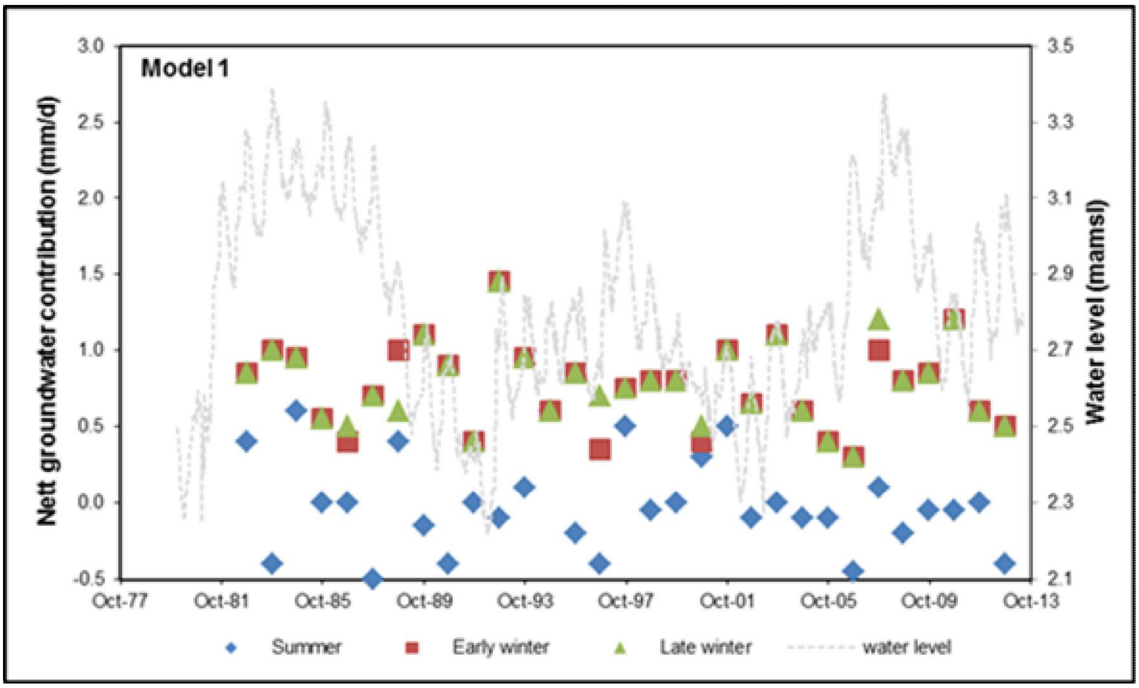

Figure 6.8: Summer and winter nett groundwater contributions to Groenvlei determined using Model 1 to calculate lake evaporation.

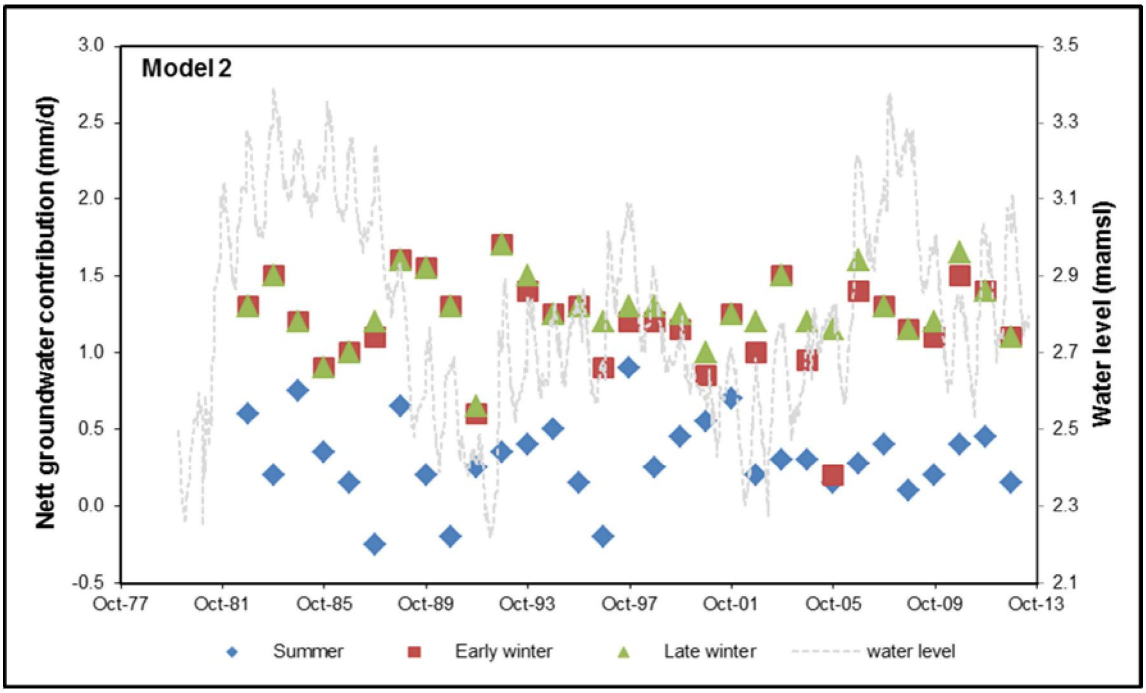

Figure 6.9: Summer and winter nett groundwater contributions to Groenvlei determined using Model 2 to calculate lake evaporation.

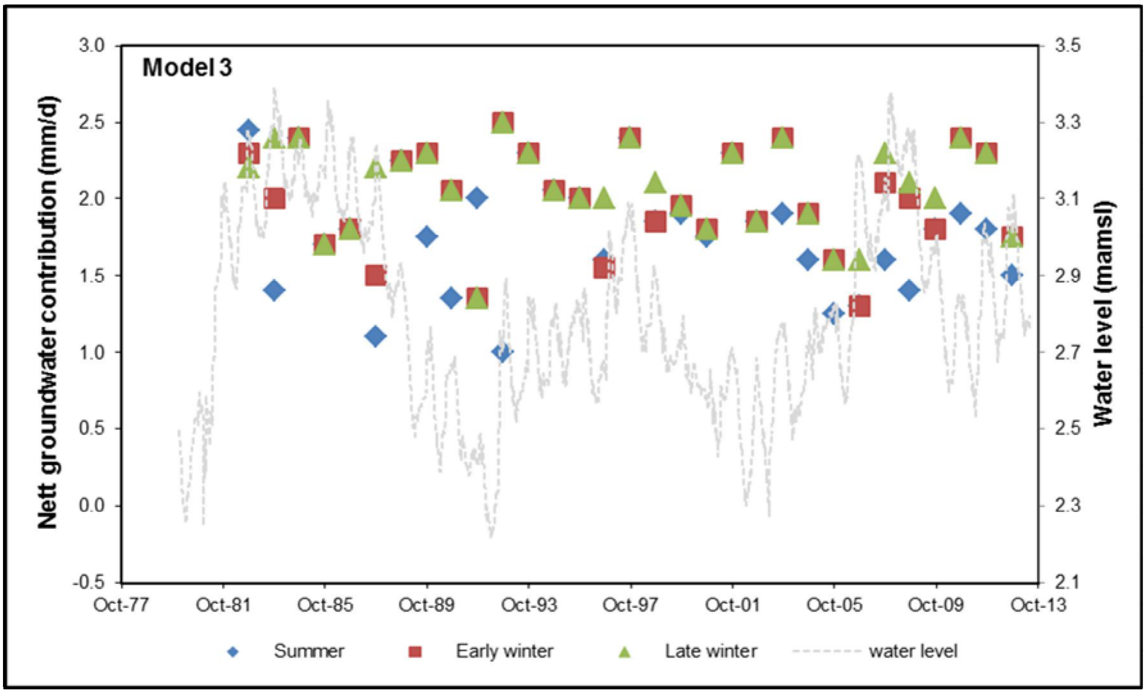

Figure 6.10: Summer and winter nett groundwater contributions to Groenvlei determined using Model 3 to calculate lake evaporation.

During this process of trying to understand the differences between summer and winter losses, the assumption and conceptualisation pertaining to losses from the reed collar being the same as that of open water was challenged (Section 4.4). In light of information about transpiration patterns of vegetation in the reed collar (Section 2.5.2.6) it was decided to split the evaporation function into open water evaporation and transpiration from the reed collar. During summer months, transpiration from the reed collar was set at 10% greater than that from open water, but it was assumed during winter the vegetation was dormant and no (or very little) transpiration occurred. Open water evaporation was assumed to be the same as that of the lake evaporation calculated using the three models.

Much better results were obtained when separately considering open water evaporation and transpiration from the reed collar than when assuming evaporation losses from the two components were the same. When applying the latter approach, it was found:

- Observed and modelled water levels could be matched in summer using a single coefficient, but not in winter;

- To be able to match the water levels for a hydrological year, a summer and winter coefficient had to be applied to the pan data;
- Alternatively, different nett groundwater contributions had to be applied the water balance for summer and winter;
- Negative nett groundwater contributions in the summer months resulted when applying Model 1 and Model 2.

The need for (a) different summer and winter coefficients and / or nett groundwater contributions and (b) calculated negative groundwater contributions were situations conceptually difficult to explain, and were considered improbable.

In contrast, the dormancy of the reed collar in winter is well supported in the literature while evidence exists that transpiration losses could be greater than evaporation from open water (Section 2.5.2.6). Analyses of the annual nett groundwater contributions determined using the three models (Table 6.3) showed a negative nett groundwater contribution was only modelled in three of 90 runs (3.3%) as opposed to the 20 negative results obtained when open water evaporation and transpiration were assumed to be the same.

The ability to model the water level of Groenvlei using a single and defensible approach with respect to evaporation losses is illustrated in Figure 6.11. In this instance, open water losses were assumed to be equivalent to lake losses determined using Model 2. Transpiration losses from the reed collar during summer were set at 110% of lake evaporation and the reeds were considered to be dormant in winter. The standard approach of setting summer transpiration losses from the reed collar at 10% more than the calculated lake evaporation and accepting the vegetation to be dormant in winter lead to good results in 63%, 41% and 50% of the time when respectively applying Model 1, Model 2 and Model 3. In most of the other instances, summer transpiration losses had to be increased to generate a good match between the observed and modelled water levels. Except for a few instances when using Model 3, the dormancy assumption remained in place.

Table 6.3: Results of modelling using the water balance with evaporation and transpiration considered independently of each other.

Year Ending	Model 1					Model 2					Model 3				
	NGC Et=110% (mm/d)	Match Rating	NGC Best (mm/d)	Et summer (% El)	Et Winter (% EL)	NGC Et=110% (mm/d)	Match Rating	NGC best (mm/d)	Et summer (% El)	Et winter (% EL)	NGC Et=110% (mm/d)	Match Rating	NGC best (mm/d)	Et summer (% El)	Et winter (% EL)
01/10/1982	0.50	5	0.50	110	0	0.75	5	0.75	110	0	2.40	3	2.40	100	100
01/10/1983	0.20	3	0.70	170	0	0.20	3	1.20	200	0	1.80	5	1.80	110	0
01/10/1984	0.55	5	0.55	110	0	0.85	5	0.85	110	0	2.20	3	2.00	90	20
01/10/1985	0.20	5	0.20	110	0	0.50	5	0.50	110	0	1.60	4	1.60	100	0
01/10/1986	0.00	4	0.10	120	0	0.25	5	0.25	110	0	1.70	4	1.70	110	100
01/10/1987	-0.20	4	0.00	150	0	-0.10	4	0.35	160	0	1.10	5	1.10	110	0
01/10/1988	0.35	4	0.35	110	0	0.65	5	0.65	110	0	2.30	5	1.60	70	0
01/10/1989	0.00	3	0.40	140	0	0.30	3	0.80	150	0	1.65	5	1.55	100	0
01/10/1990	0.00	3	0.00	110	0	0.10	3	0.70	175	0	1.50	5	1.50	110	0
01/10/1991	0.15	4	0.15	110	0	0.45	5	0.45	110	0	1.40	4	1.40	120	0
01/10/1992															
01/10/1993	0.45	4	0.45	110	0	0.80	4	0.80	110	0	1.80	4	1.80	110	0
01/10/1994	0.15	5	0.15	110	0	0.45	4	0.45	110	0	1.30	4	1.30	120	0
01/10/1995	0.25	5	0.25	110	0	0.45	3	0.80	150	0	1.50	5	1.50	110	0
01/10/1996	-0.40	3	0.00	150	0	0.00	3	0.50	175	0	1.10	5	1.10	110	0
01/10/1997	0.55	5	0.55	110	0	0.90	4	0.90	110	0	1.60	5	1.60	110	0
01/10/1998	0.20	5	0.45	150	0	0.30	3	0.90	165	0	1.80	4	1.60	100	0
01/10/1999	0.20	5	0.20	110	0	0.40	4	0.60	130	0	1.40	3	1.30	100	0
01/10/2000	0.20	5	0.20	110	0	0.50	5	0.50	110	0	1.80	3	1.20	80	0
01/10/2001	0.60	5	0.60	110	0	0.80	4	0.80	110	0	2.30	4	2.00	100	0
01/10/2002	0.15	5	0.30	130	0	0.35	4	0.75	150	0	1.80	4	1.60	100	0
01/10/2003	0.20	3	0.90	160	0	0.40	3	1.25	175	0	2.00	5	2.00	110	0
01/10/2004	0.20	5	0.20	110	0	0.45	4	0.65	140	0	1.40	5	1.40	110	0

01/10/2005	0.00	4	0.00	130	0	0.20	5	0.20	110	0	1.30	4	1.00	100	0
01/10/2006	0.20	4	0.70	200	0	0.15	3	0.90	175	0	1.60	5	1.60	110	0
01/10/2007	0.00	3	0.80	200	0	0.30	3				1.60	5	1.60	110	0
01/10/2008	0.15	4	0.15	110	0	0.40	4	0.45	140	0	1.30	5	1.30	110	0
01/10/2009	0.20	5	0.25	120	0	0.40	4	0.60	130	0	1.80	4	1.50	100	0
01/10/2010	0.40	4	0.90	200	0	0.40	3	1.00	175	0	1.80		1.80	110	0
01/10/2011	0.20	3	1.50	250	0	0.40	3			0	1.80	4	2.40	150	0
01/10/2012	0.25	5	0.25	110	0	0.50	4	0.70	130	0	1.40	5	1.40	110	0
Average	0.20		0.39	134	0	0.42		0.69	137	0	1.67		1.59	106	7
Std. dev.	0.22		0.34	37	0	0.24		0.25	28	0	0.34		0.34	13	25
Min.	-0.40		0.00	110	0	-0.10		0.20	110	0	1.10		1.00	70	0
Max.	0.60		1.50	250	0	0.90		1.25	200	0	2.40		2.40	150	100

Notes:

NGC - nett groundwater contribution

Et = 110% - summer transpiration set at 10% more than lake evaporation determined by that model, winter transpiration set at 0% of lake evaporation (i.e. basic assumption)

NGC best – best estimate of NGC using Et summer and Et winter

Et summer – Transpiration in relation to lake evaporation used to determine NGC best

Et winter – Transpiration is relation to lake evaporation used to determine NGC best

A rating was given to each curve matching using a score of 1 (no match) to 5 (very good match)

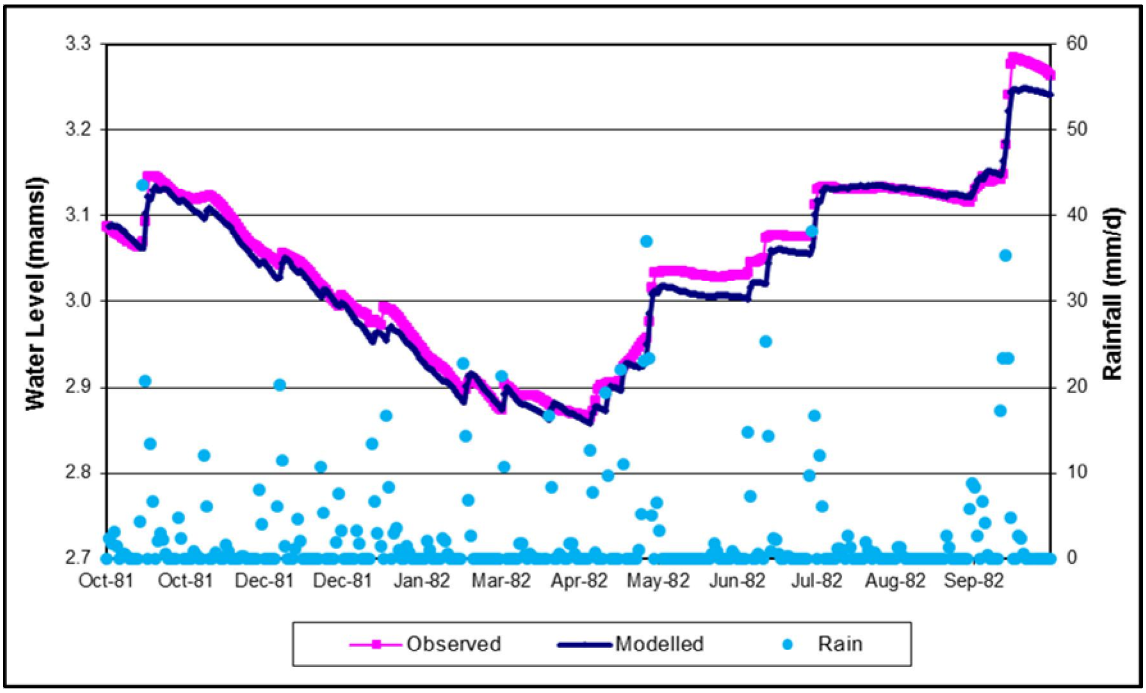

Figure 6.11: In comparison to Figure 6.7, good results were obtained when modelling with the reed collar being dormant in winter and summer transpiration being 10% more than open water losses. Nett groundwater contribution was computed to be 0.7 mm/d.

At times Model 3 predicted higher water levels than observed in the winter months. To get a good match, the summer transpirations had to be set less than lake evaporation, and in three instances the winter dormancy assumption had to be overruled (Table 6.3). This was considered the exception to the rule and not in keeping with the conceptual understanding of the hydrology of Groenvlei.

6.1.2.5 Summary of results from water balance model

Data analysis to this point had not provided an indication of which of the three models provided better estimates of lake evaporation (Table 6.4). The approach prescribed by Midgley et al. (1994) (Model 1) yielded the smallest nett groundwater contribution and the results appeared reasonable. While 63% of the record produced good results when assuming dormancy in winter and summer transpiration to be 10% more than lake evaporation, analysis of the data yielding the best fit for each hydrological year indicated transpiration during summer be 30% more than open water evaporation. The results strongly support the winter dormancy assumption.

Table 6.4: Estimates of nett groundwater contribution using three methods to quantify lake evaporation from pan data.

Parameter	Model 1	Model 2	Model 3
Nett groundwater contribution (Et summer 110%, ET winter 0%)	0.20	0.42	1.67
Nett groundwater contribution (mm/d) (best)	0.37	0.68	1.58
Summer transpiration (% lake evaporation)	130	138	106
Winter transpiration (% lake evaporation)	0	0	4

Using A pan data multiplied by 0.7 (Model 2) yielded slightly larger nett groundwater contributions and the results also appeared reasonable. The summer transpiration was also higher than that determined when using Model 1, but within ranges presented in the literature. The dormancy assumption is supported.

Model 3 – which assumes lake evaporation to be equivalent to A pan evaporation – yielded significantly higher nett groundwater contributions. Also, this approached required summer transpiration to be similar to lake evaporation. The winter dormancy assumption was not support in two instances of the 30 year record.

The implication of the three approaches is that Model 3 requires significantly greater groundwater flow into Groenvlei than Model 1 and Model 2. Calculation of groundwater inflow and outflow provided an indication of which model yielded the best or most reasonable results. Similar hydraulic gradients up- and down-gradient of the lake suggested smaller nett groundwater contributions might be more representative of prevailing conditions. This aspect is addressed further in Section 6.2.

6.2 Application of Darcy's Law

In the previous section, the nett groundwater contribution was presented as the unknown of the water balance. However, this is not true as the groundwater components can be calculated using Darcy's Law, as described in Section 2.5.3. Information required to apply Darcy's Law includes aquifer transmissivity, hydraulic gradient and the width through which the groundwater must flow. These groundwater characteristics are described in Chapter 5 and presented in Table 6.5.

Table 6.5: Calculation of groundwater flow into and out of Groenvlei.

Component	T (m2/d)	i	W (m)	Q (Mm3/a)	Q (mm/d)
Inflow Northern boundary Western boundary	270	0.0016	6040	0.952	0.73
Outflow Southern boundary	270	0.0013	4520	0.579	0.44
Nett groundwater contribution				0.373	0.29

The greatest uncertainty in the above calculation relates to transmissivity. The average value interpreted from pumping tests conducted in and around Groenvlei (Section 5.3) was used to determine nett groundwater contribution. A nett groundwater contribution of 0.29 mm/d was similar to that computed using Model 1 in Section 6.1. If a T value of 300 m^2/d were used (as per Dennis, 2010), a nett groundwater contribution of 0.34 mm/d would result. This again supports the outcome using Model 1. If the harmonic mean was used as a measure of central tendency, then a T value of 150 m^2/d would be used to compute a nett groundwater contribution of 0.16 mm/d. Such an estimate would also be in line with that obtained using Model 1 to set lake evaporation. The groundwater flow calculations point to Model 1 being the best approach to use, but do not discount Model 2 being a reasonable approach. However, the groundwater flow calculations do not support Model 3 as a well-founded approach.

As a check on the Darcy flow calculations, the boundaries determined in Section 5.5 and the average flux presented in Section 2.5.3.3 were used to determine groundwater flow into and out of Groenvlei (Table 6.6). A nett groundwater contribution of 0.17 mm/d to Groenvlei was then calculated by dividing the sum of flows by the area of the lake. This estimate is the same as that determined using a T of 150 m^2/d, but slightly lower than that determined using the water balance or average T value in the flow calculations. It is possible the Eden Primary Aquifer is more permeable than the other coastal primary aquifers, thereby resulting in the average flux method under-estimating flow into and out of the lake.

Table 6.6: Calculation of groundwater inflow and outflow using an average flux determined from the literature.

Boundary	Width (m)	flux (m3/d/m)	flow (Mm3/a)	flow (mm/d)
Groundwater inflow	6040	1.11	2.45	0.68
Groundwater outflow	4520	-1.11	-1.83	-0.51
Nett groundwater contribution			0.62	0.17

6.3 Chemical Mass Balance

Cl is typically used in chemical mass balance models because it is conservative in nature. Also, in this study, it was the only element for which concentration in rainfall was available. The harmonic mean of EC and Cl concentrations of different waters in and around Groenvlei is presented in Table 6.7.

Table 6.7: Harmonic mean of EC and Cl concentration of water in and around Groenvlei

Water	EC (mS/m)	Cl (mg/l)	Samples
Rainfall	-	20	1
Groundwater (in)	87	135	109
Evaporation	0	0	0
Groenvlei	438	1 215	429

In terms of the mass balance, it was assumed the Cl concentration of Groenvlei is representative of the quality of water leaving the lake and discharging into the aquifer. The Cl concentration of the deeper, more saline water sampled at GRU1 – located directly south of the lake – is 400 mg/L (6 samples), suggesting significant change over a distance of only 240 m. The ratio of Cl concentration to EC is similar for groundwater flowing into Groenvlei and the upper water samples at GRU1 (Table 6.8). The deeper water at GRU1 has a ratio intermediate of that of Groenvlei and the other groundwaters. The reason for this difference in ratio is the lake water is strongly NaCl in character, while the groundwater has a mixed cation and anion character (NaCa – ClAlk).

Table 6.8: Harmonic mean of EC and Cl concentration of groundwater in and around Groenvlei and the Cl / EC ratio of the different groundwaters.

Water	EC (mS/m)	Cl (mg/l)	Ratio Cl / EC	Samples
Groundwater (in)	87	135	1.6	109
Groenvlei	438	1 215	2.8	429
GRU1 (shallow)	65	90	2.0	6
GRU1 (deep)	196	400	1.4	6

The Cl concentration of rain discussed in Section 2.6.2 was used in the chemical mass balance model, while that of evaporation was accepted as 0 mg/L (Section 2.5.2.5). Small amounts of chemicals may be removed by evaporation and / or wind entrainment, but such amounts are likely to be so small that they will be masked by data errors and uncertainties.

Results of initial chemical mass balance modelling are presented in Table 6.9. A poor result was obtained when applying the chloride concentrations described above with the results of the daily water balance and groundwater flow modelling. A balance of –526 tons/a (or – 298%) indicates a nett release of chloride, with too much Cl leaving the system. Analysis of the results suggested the following potential sources of error:

- The flow data determined using the daily water balance and Darcian calculations were wrong;
- The chemical data used in the chemical mass balance were wrong;
- Failure to specifically consider dry deposition;
- Chemical processes take place within the lake that were not considered; and / or
- A combination of the above.

Table 6.9: Initial chemical mass balance of Groenvlei

Component	Flow Mm3/a	Cl conc (mg/L)	Cl load (tons/a)
Rainfall	2.41	20	48
Groundwater in	0.95	135	129
Inflow	3.36		177
Evaporation	2.12	0	0
Transpiration	0.74	0	0
Groundwater out	0.58	1 215	703
Outflow	3.44		703
Balance (%)	*-2.3*		*-298*

The first two potential sources of error seem unlikely. Reasonably good agreement was obtained between the daily water balance and Darcian flow calculations, with the former being based on climatic data and lake level monitoring and the latter on hydraulic gradients and aquifer properties. It was possible to get the chemical mass balance to balance by retaining the nett groundwater contribution of 0.3 mm/d, but decreasing groundwater flow into and out of the lake (Table 6.10).

Table 6.10: Revised chemical mass balance of Groenvlei retaining the nett groundwater contributions, but reducing inflows and outflows.

Component	Original Flow mm/d	Modified Flow mm/d	Cl conc (mg/L)	Cl load (tons/a)
Rainfall	1.84	1.84	20	48
Groundwater in	0.73	0.36	135	64
Inflow	2.56	2.20		112
Evaporation	1.62	1.62	0	0
Transpiration	0.56	0.56	0	0
Groundwater out	0.44	0.07	1 215	111
Outflow	2.62	2.25		111
Balance (%)	*-2.3*	*-2.3*		*0.5*

This required reducing groundwater inflow by 50% and groundwater outflow by 84%. Motivating high confidence in the estimates of hydraulic gradient and a moderate to high confidence in the length of the outflow boundary, a T value of 33 m^2/d was needed to model the outflow required to balance the chemical mass balance model i.e. almost an order of magnitude less than the average T of the Eden Primary Aquifer. Pumping test data from three boreholes south of the lake (GRU1, GRU2, GRU3) indicated transmissivities to be in line with the average values (Table 5.2) and not in the order required for a better balance.

Similarly, the Cl concentrations of the component parts used in the mass balance are also of high confidence. The concentration of groundwater flowing into Groenvlei and that of the lake itself are based on data sets of 109 and 429 samples respectively. Notwithstanding inter-seasonal fluctuations driven by rainfall patterns, the standard deviation of the Cl concentration of Groenvlei is relatively small at 150 mg/L. PGB1, located directly north of Groenvlei and sampled five times, has a Cl concentration of 100 mg/L, while two wellpoints at the north western edge of the inflow boundary had Cl concentrations of 190 mg/L. This also points to confidence in the Cl concentration used. A good balance could be obtained if the Cl concentration of rainfall was set at 240 mg/L or a Cl concentration of 690 mg/L was used for inflowing groundwater. However, no evidence is available to support such levels.

The dry deposition of chloride in the Groenvlei catchment was not specifically considered in the chemical mass balance. Chloride deposited on the groundwater by dry deposition would be transported into the groundwater system by infiltrating and percolating rainfall, and thus included in the groundwater flowing into the lake. Also, the measurement of chloride concentration by Weaver and Talma (2005) specifically included dry deposition. Consequently, exclusion of a dry deposition composition component in the water balance is unlikely to be a source of error.

The above discussion illustrates that it was simply not possible to balance the chemical mass balance in that form when maintaining a nett groundwater contribution of about 0.3 mm/d and using data considered of high confidence. A remaining option is Cl (and other chemicals) being retained or consumed in the lake, with the assumption that leakage from the lake having the same chemistry as the lake itself, being wrong. To achieve a good balance and conform to observations, the Cl concentration of the outflowing water would have to be in the order of 400 mg/L [1] or 33% of lake Cl concentration. Also, it is acknowledged the Cl load not accounted for in the water balance is small (a) in terms of the Cl load of Groenvlei (11 047 tons) and, when converted to a concentration, equates to 52 mg/L Cl or 4.3% of the Cl concentration of Groenvlei.

[1] This equates to a load of 230 tons/a i.e. 24% more than required to balance the incoming salts. Given that the steady-state model does not account for fluctuations in salinity of the reed collar or salt losses through wind entrainment, water losses to the subsurface at a concentration of 400 mg/L are feasible.

There is clear evidence the chemistry of the lake evolves as the input waters merge to form the NaCl dominated lake water with an EC of 440 mS/m. The southern Cape's rainfall composition in the lowaltitude regions contains elevated levels of Na and Cl because of high levels of windblown maritime aerosols in the coastal regions, with other constituents being relatively depleted (Van Wyk et al., 2011). The CaAlk character of the incoming groundwater dissipates as it blends with lake and rain water. Also, the incoming groundwater has a negative calcium carbonate precipitation potential (CCPP) (-23) [2], indicating the salt is taken into solution rather than precipitates out of solution. Conversely, Groenvlei water has a CCPP of 26 and calcium carbonate will precipitate out of solution and account for the lake-bottom calcareous muds described by Martin (1960a). This process will also enhance the NaCl character of the lake water.

It is well documented that the interaction of groundwater and vegetation has the ability to influence water quality. The salinisation problem of eastern Australia is possibly the most extreme example, but the work of McCarthy et al. (1993) and Humphries et al. (2011a, 2011b) is more relevant to the Groenvlei situation. These researches argue evapotranspiration by vegetation concentrates salts in groundwater, which in turn influences vegetation distribution. A change in groundwater quality can also result in the precipitation of less soluble minerals such as $CaCO_3$ and SiO_2.

Similarly, water chemistry changes as water leaves the lake either through evaporation or leakage. Evaporation removes pure (or almost pure) water, leaving the salts behind. If Groenvlei were endorheic, the salts would accumulate and the lake would tend towards a salt pan. However, removal of salts in solution keeps the lake in its current balance. Roets (2008) observed three sets of wellpoints west and south of the lake and in the reed collar all yielded salinities greater than that of the lake (Figures 6.12 and 6.13). This supports the contention that physical, chemical and biological processes in the reed collar and hyporheic zone retain salts, and prevent them leaving the lake. Through daily or weekly wave action caused by wind, the more saline water at the edge of the lake is mixed back into the lake water. This process of re-entrainment of salts is illustrated in Figure 6.14.

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[2] Calculated using the Stasoft4 package (Morrison and Loewenthal, 2000)

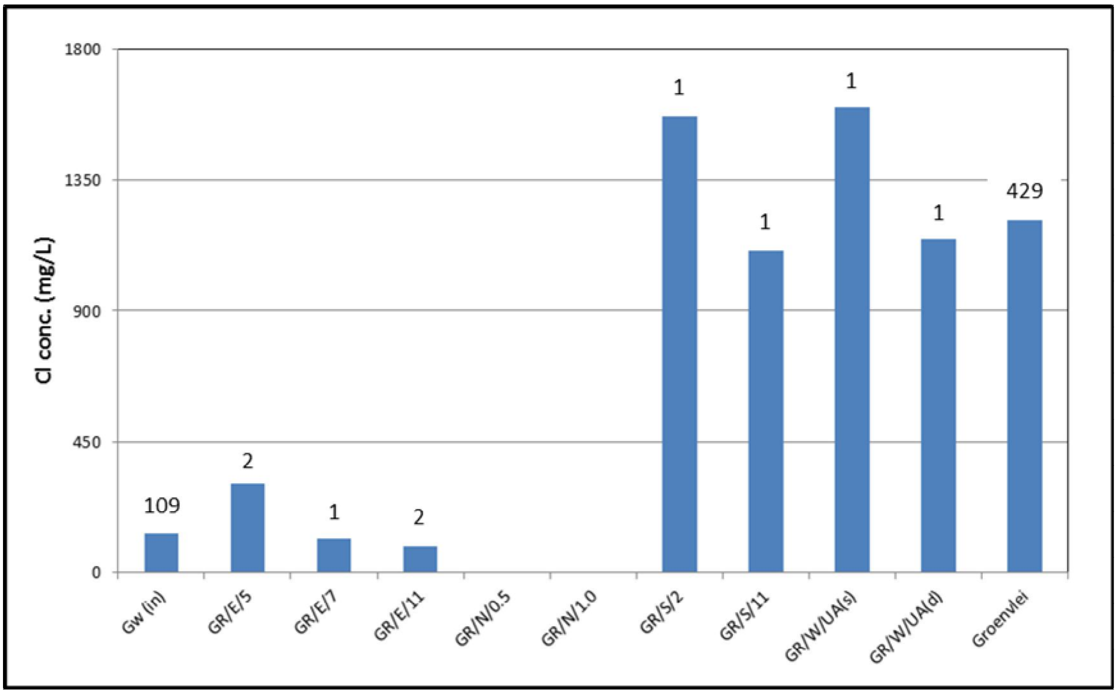

Figure 6.12: Cl concentration measured by Roets (2008) in shallow wellpoints sunk at the edge of Groenvlei. The harmonic mean of groundwater and Groenvlei determined during this study are included for comparison, while the number of samples that were available for analysis is indicated above each bar.

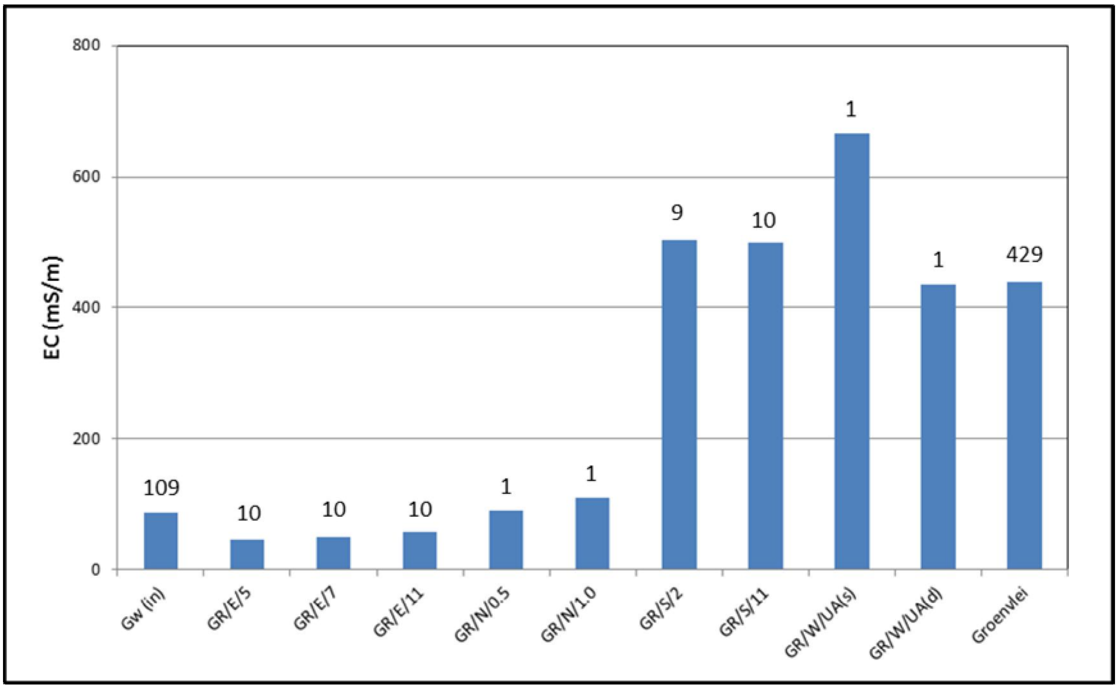

Figure 6.13: EC measured by Roets (2008) in shallow wellpoints sunk at the edge of Groenvlei. The harmonic mean of groundwater and Groenvlei determined during this study are included for comparison, while the number of samples that were available for analysis is indicated above each bar.

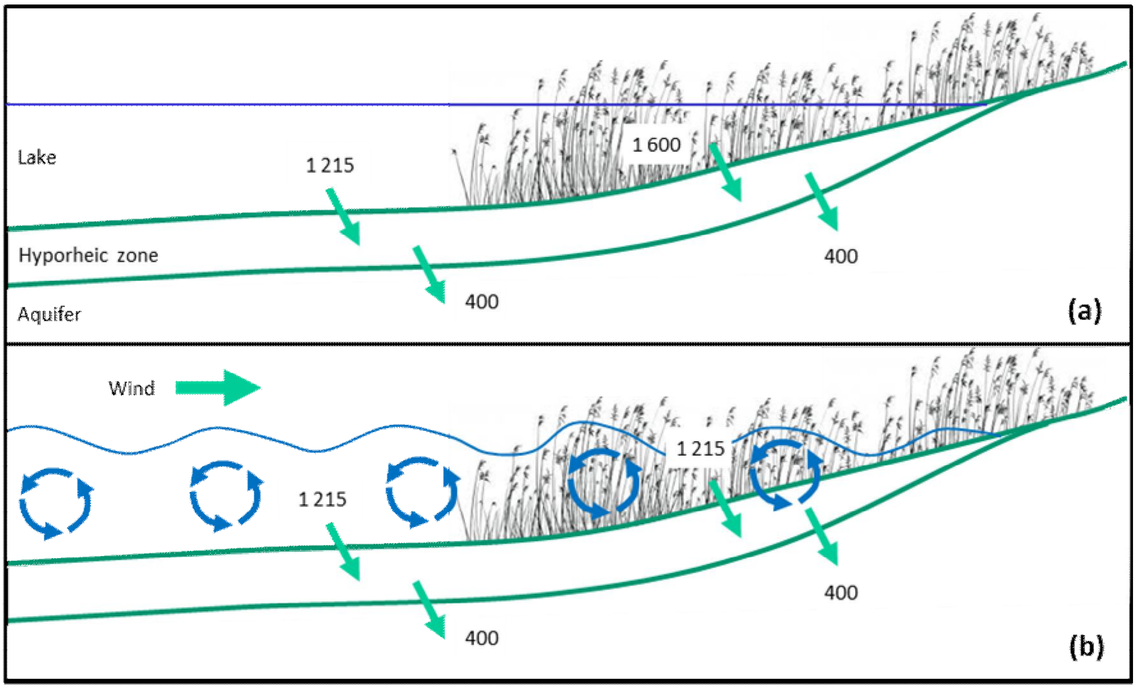

Figure 6.14: (a) Conceptualisation of the build-up of salts in the reed collar and hyporheic zone, and (b) the re-entrainment of salts by wind and mixing of the lake. The concentrations depicted are in mg/L Cl.

Water quality monitoring at Groenvlei revealed quality to be variable, ranging by 50 mS/m to 80 mS/m over a period of between weeks and months (Figure 4.11). This variability is hard to explain given the large salt load of Groenvlei in relation to the small inputs from rain and groundwater. However, if it is considered that the sampling station K4R001 is positioned at the edge of the reed collar (Plate 3.1), then the observed fluctuating quality is readily explained by the re-entrainment process described above.

Re-modelling the chemical mass balance of Groenvlei including the re-entrainment component described above allowed for a good chemical mass balance to be obtained (Table 6.11). It was not possible to measure the re-entrainment component; consequently it was assumed the salt load re-entrained to be equivalent to the nett loss predicted in Table 6.9. Based on this chemical mass balance, the concentration of Cl brought back into the lake is 58 mg/L (i.e. 4.8% of the lake water) while that discharged to the aquifer could be in the order of 890 mg/L.

Table 6.11: Revised chemical mass balance of Groenvlei including the re-entrainment component of salts in the vegetative fringe.

Component	Flow Mm3/a	Cl conc (mg/L)	Cl load (tons/a)
Rainfall	2.41	20	48
Groundwater in	0.95	135	129
Inflow	3.36		177
Evaporation	2.12	0	0
Transpiration	0.74	0	0
Groundwater out	0.58	1 215	703
Re-entrainment			-526
Outflow	3.44		177
Balance (%)	*-2.3*		*0.0*

6.4 Integration of Results

The groundwater contribution to Groenvlei has been determined using three different approaches, two of which are independent of each other. The chemical mass balance approach used flow data and salt concentrations to balance the incoming and outgoing salt load. Notwithstanding the need to review the mechanisms of salt movement in the vegetative fringe, the results from the different approaches corresponded well with each other and appear reasonable in terms of the conceptualisation of the hydrology of the lake. Both the daily water balance method and Darcian flow calculations indicated nett groundwater contribution to be about 0.30 mm/a. This is substantially less than the 2.03 mm/d put forward by Parsons (2008a), with the difference being attributed mainly to the modelled evaporation losses. Parsons (2008b) assumed lake evaporation to be equivalent to A pan evaporation. This research explored the role of open water evaporation and transpiration losses from the reed collar and found the combined evapotranspiration losses to be less than originally assumed.

Construction of a groundwater level contour map and additional pumping test data allowed flow boundaries, hydraulic gradients and aquifer properties to be better defined. In turn, this allowed better estimates of groundwater flow into and out of Groenvlei to be made. A degree of confidence is taken that the Darcian flow calculations independently yielded a similar nett groundwater contribution to that resulting from the daily water balance model. It is thus proposed groundwater flowing into Groenvlei amounts to 0.95 Mm^3/a or 0.73 mm/d. Groundwater is of good quality, having an EC in the order of 90 mS/m and a Cl concentration of 135 mg/L. Groundwater flow out of Groenvlei – or leakage – amounts to 0.58 Mm3/a or 0.44 mm/d. This water is more saline, having a salinity of about 270 mS/m and a Cl concentration of about 400 mg/L.

Direct rainfall is the dominant source of water, accounting for 71.6% of the input. Groundwater provides the balance of 28.4%. Evaporation (61.7%) and transpiration (21.4%) account for the greatest losses, collectively being responsible for 83.1% of flow out of Groenvlei. Discharge into the aquifer accounts for the remaining 16.9% of outflow.

7 DISCUSSION OF RESULTS

7.1 Water Balance Modelling

The water balance is a fundamental building block upon which conceptual models of hydrologic systems are constructed (Healy, 2010). Using a water balance model to quantify groundwater contributions to Groenvlei highlighted the need to (a) conceptualise a problem and (b) continually test for reasonableness of approach and results obtained. This is a key tenet of science and application of the scientific method. An integrated approach using independent methods to address an issue provided a way of improving and verifying results. Application of Darcy's Law to Groenvlei provided a means of calculating groundwater inflow and outflow – an important aspect of this research. Coupling this to a water balance, a chemical mass balance and other techniques such as isotope sampling provided a test of the reasonableness of the outcomes and refinement of the parameters used.

As simple as the water balance may seem, and as easy as it is to get a result, good results are subject to the quality of data available, an appreciation of the uncertainties associated with the data and how they impact the water balance, and a sound conceptual understanding of the hydrological system. Conceptualisation of an endorheic system, for example, would drive a computational bias that water does not leave the lake. Clearly the availability of new data (such as that from three boreholes south of Groenvlei) improves the ability to correctly conceptualise situations, but the availability or absence of data is a critical aspect of all science. The availability of new information allows for the falsification of existing knowledge, thereby allowing knowledge and understanding to grow and improve.

This research was based on the water balance method, but supported by Darcian flow calculations and chemical mass balance modelling. The independence of the three approaches provided a degree of confidence in the outcomes of the research, but a degree of variability in the water balance results was observed that require further deliberation. Nonetheless, observations and the results of calculations match the conceptual understanding of the hydrology of Groenvlei and point to a convergence of evidence.

7.2 Conceptualisation and Understanding of Groenvlei

This research, together with that undertaken previously, has contributed to a better understanding of the hydrology of Groenvlei. This is important if the lake is to be conserved and properly managed so it remains in a natural and healthy state. The current conceptual understanding of the hydrology of Groenvlei is presented in Figure 7.1 and discussed below. Included in the discussion are some previous claims and conclusions that have been shown to be wrong and that need to be addressed if our knowledge of the system is to continue to grow.

7.2.1 Groenvlei is not endorheic

The immediate slopes around Groenvlei slope inwards towards the water body, providing the impression that the wetland is inward draining. However, it has clearly been established through analysis of water levels, chemical profiling, temperature measurement and isotope sampling that water drains out of the lake along the southern boundary. It has also been established that the removal of salts through seepage along the southern boundary is critical to the chemistry of Groenvlei. If the lake was terminal in character (i.e. no outflows other than evaporation), a build-up of salts would occur and the water quality would become saline or even hypersaline. The hypothesis that Groenvlei is a flow-through system, with groundwater discharging into the lake from the north and water discharging from the lake into the subsurface in the south is thus accepted.

Subsurface outflow invalidates claims that flow (*rheîn*) remains within (*éndon*) the system (Jones and Day, 2003; Duncan, 2006, Roets et al., 2008b, Vromans et al., 2010 and others). The hydrological world has moved to a holistic and integrated understanding of hydrological systems and the hydrological cycle, as has South African legislation. If the integration of surface water and groundwater is ever to become a reality in this country, the study and administration of water resources require hydrologists and ecologists appreciate the hydrological system as a whole.

7.2.2 Permeability south of Groenvlei is not low

Fijen (1995) postulated low permeability of cemented dune rock and aeolianite south of Groenvlei resulted in small seepage losses from the lake. To the contrary, results of pumping

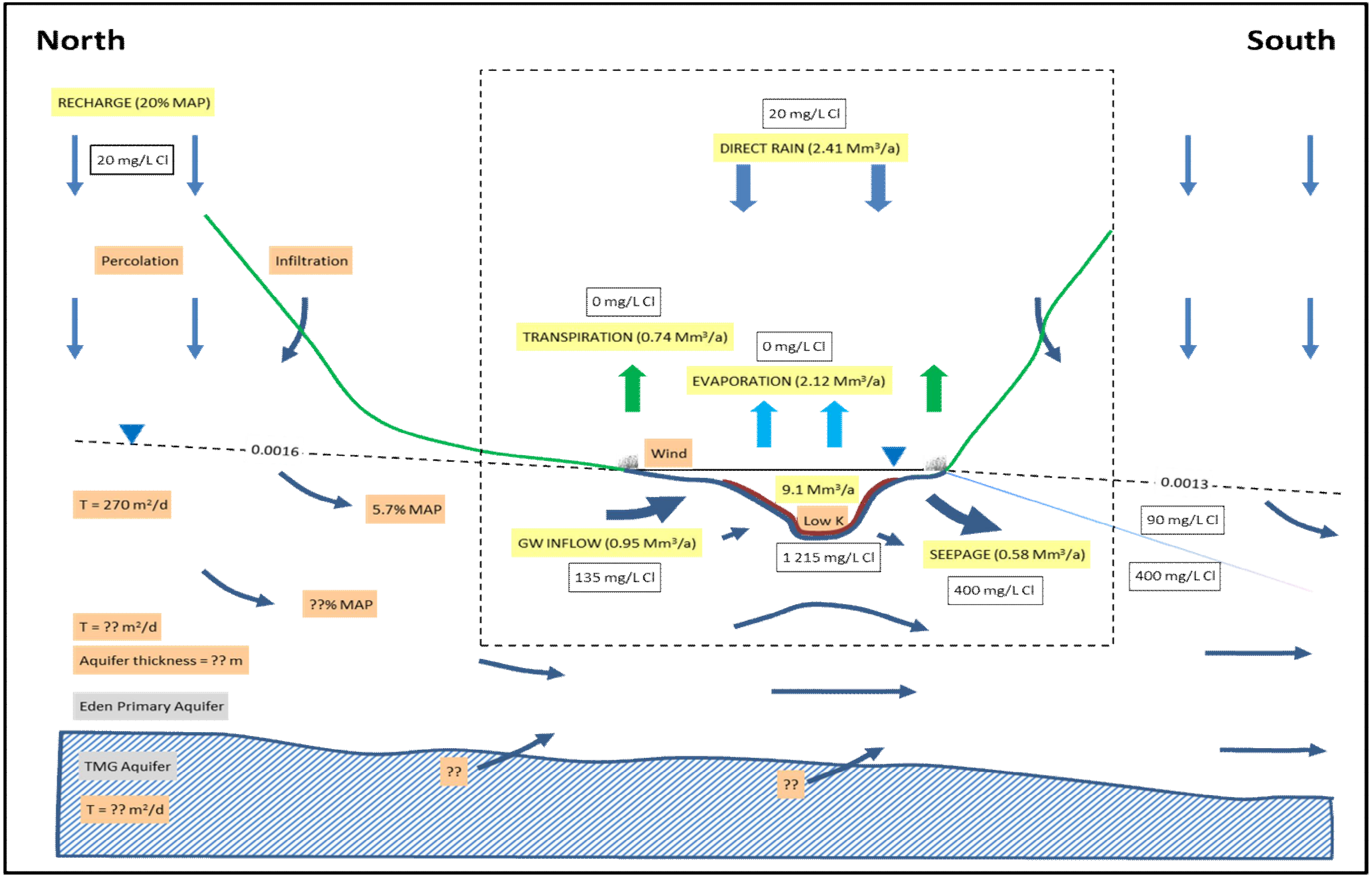

Figure 7.1: Illustration of the conceptual understanding of the hydrology of Groenvlei by means of a north – south profile.

tests conducted south of Groenvlei and elsewhere in the immediate vicinity confirmed the aquifer to be permeable (Section 5.3). If Fijen (1995) had had this information at his disposal, his water balance may have been substantially different to that presented.

Vivier (2009) motivated his estimate of leakage from the lake by arguing colmation caused a reduction in hydraulic conductivity of lake sediments. He proposed a minimum K of 0.2 m/d. Notwithstanding limitations of seepage meters discussed in Section 2.5.3.1, Paridaens and Vandenbroucke (2007) measured K at four points in the lake [1] using seepage meters, providing average values of 4.3 m/d, 1.5 m/d, 7.2 m/d and 18 m/d for each point. These measurements are between one and two orders of magnitude greater than that put forward by Vivier (2009). Also, it is submitted colmation is unlikely to be an important process controlling the hydraulic properties of lake edge sediments because of regular wind-induced turbulence and wave action leads to decolmation [2]. Field measurement is required to investigate the contention that lower K values of lake sediments hinder seepage along the southern boundary of the lake.

7.2.3 Groenvlei is not fed by TMG Aquifers

There is no scientific evidence that Groenvlei is in any way connected to TMG Aquifers. A considered review of the PhD thesis (Roets, 2008) and papers (Roets et al., 2008a, 2008b) that claim such a linkage found water level data available at that time was not considered, the geohydrological conceptualisation (particularly that relating to hydraulic gradients and water chemistry) was flawed, and no scientifically credible data or information was presented in support of that hypothesis. No new information has come to light since the assessment by Parsons (2009b) that requires the assessment be reviewed. Information pertaining to the thickness of the primary aquifer and the nature of the underlying bedrock would strengthen the argument that the underlying fractured aquifer does not sustain Groenvlei. Following the findings of Weaver (1998) relating to the isotope signature of TMG-derived groundwaters, examination of ^{87}Sr / ^{86}Sr ratios in waters in and around Groenvlei could also prove useful. In light of available information, the hypothesis that there is no hydraulic link between Vankervelsvlei and Groenvlei is accepted.

1 It was not possible to determine the positions of the measurements, but at least one location was along the southern edge of the lake.

2 Birch et al. (1978) found the shallow fringes of Swartvlei dominated by sand. It was only in the central deeper (> 10 m) parts that mud and clay dominated.

7.2.4 Groenvlei is a rain-driven system

Parsons (2003) was the first to appreciate the role of groundwater in sustaining Groenvlei, estimating about 15% of the input into the lake was via the subsurface. In light of current knowledge, claims that the lake was a groundwater-driven system were overstated. Groenvlei is principally a rain-driven system, with direct rainfall onto the wetland accounting for 71.6% of input. Groundwater plays a secondary role, supplying the remaining 28.4%.

Parsons (2009a) found that even in a system as large and diverse as the St Lucia wetland, direct rainfall accounted for 59% of input, with inflow from rivers accounting for 31% of input. Direct rainfall onto wetlands should always be considered as a potential key source of water of wetlands. The implication of this for Groenvlei – and many other wetlands – is that the main hydrological input cannot be managed, except possibly by the influence of man on climate change. In its short history, Groenvlei has been subjected to a range of climatic conditions to which it has adapted.

Though groundwater is the secondary source of water, it would be wrong to assume it to be of less importance. As shown in Section 7.5.1, the contribution of groundwater still influences lake levels and that overuse of groundwater could negatively impact the state of Groenvlei.

The hypothesis that Groenvlei is a groundwater-fed system is true only in that Groenvlei is fed by groundwater and not by surface water. As direct rainfall onto the wetland provides the greatest input into the lake, and groundwater plays a secondary role, it would be correct to refer to Groenvlei as a rain-driven system. Similarly, the hypothesis that Groenvlei is not fed by surface run-off is accepted. No rivers drain into Groenvlei and comparison of rainfall data to lake level data showed the rise of the level of the lake is equivalent to rainfall.

7.2.5 Flushing rate of Groenvlei

In Section 4.5 it was estimated the flushing rate of Groenvlei was 1.4 years. Using the revised flows resulting from this research (Table 6.11), it was re-calculated water in Groenvlei is replaced every 2.7 years.

7.2.6 Groenvlei's catchment covers an area of about 25 km^2

Previous estimates of the catchment area of Groenvlei ranged between 9.5 km^2 and 13.8 km^2. These estimates were presumably based on surface water considerations and are deemed inappropriate in the absence of surface runoff. As observed by Winter et al. (2003), water table divides (boundaries of local ground water flow) commonly do not underlie local surface watersheds in areas underlain by a relatively permeable unconfined aquifer. Consideration of groundwater levels and groundwater flow directions indicated the catchment – or capture zone – to be 25 km^2 (Figure 5.12). If this is the case, it can be calculated 5.7% of rainfall in the catchment reports to Groenvlei. This estimate is low when it is considered recharge equates to at least 20% MAP and that there is no surface drainage.

7.2.7 Shifting of catchment boundaries

A nett groundwater contribution to Groenvlei of 0.3 mm/d has been determined from this research. It was also assumed this contribution would remain (relatively) constant over time. However, a degree of inter-annular variability is observed in the modelled data presented in Table 6.3 that challenges the constancy assumption motivated in Section 6.1.2.2. Winter et al. (2003) found that water table divides of small catchments comprising permeable aquifers did not correspond with their surface counterparts, and that the subsurface boundaries were not static. Few studies have researched the impact of moving groundwater divides on water fluxes to surface water bodies. Insufficient spatial and temporal groundwater level data were available during this study to consider the possible effect of changing hydraulic gradients, and no relationship was found between mean annual rainfall and nett groundwater contribution. It is possible temporal changes in hydraulic gradient could contribute to fluctuations in nett groundwater contribution, but in the absence of any evidence in this regard the assumption of constancy is considered reasonable.

7.2.8 Role of the reed collar

The role of the reed collar in the functioning of Groenvlei is a key aspect to emerge from this research, both in terms of transpiration losses and lake water chemistry. The reed collar comprises almost 32% of the wetland. Results of daily water balance modelling showed vegetation is dormant in winter and no (or very little water) is lost through transpiration at this time. Evaporation from the open water – comprising 68% of the wetland – is equivalent to that predicted using the approach proposed by Midgley et al. (1994) to quantify lake evaporation using S pan evaporation data. During summer, open water losses are also quantified using the Midgley et al. (1994) method, but transpiration from the reed collar is between 10% and 30% more than that lost from open water. Failure to appreciate this dynamic resulted in an inability to model summer and winter water levels using the same pan coefficients or nett groundwater contribution.

The fringe vegetation also plays an important role in controlling the water quality of Groenvlei, and specifically that leaving the lake along its southern shore. Chemical mass balance modelling suggests as much as 526 tons of chloride is assimilated by the vegetation each year and retained in the hyporheic zone, until entrained into the main body of water through mixing by wind. This re-entrainment occurs on a daily to weekly basis, and accounts for erratic fluctuations in salinity observed in the monitored record. This process also explains the elevated salinities observed by Roets (2008) in wellpoints at the water's edge.

7.2.9 Reversibility of local hydraulic gradients

The band of poor groundwater quality on the western perimeter of Groenvlei was observed by Parsons (2005a), but not adequately explained. Possible reasons for it included saline intrusion and / or diffusion. In addition to the temporary build-up of salts in the reed collar, it is proposed the relative rate of rise in lake level and groundwater levels during significant rain events (say 40 mm) cause localised and temporary reversal of hydraulic gradients. During this time, brackish water from the lake discharges into the groundwater system. It is also possible the entrainment of salts into the coast-parallel winds deposit salts at the western and eastern edges. Because the brackish character of groundwater in the western parts appears stable and ranges between 240 and 350 mS/m (5 samples), the process causing the condition must occur sufficiently frequently to prevent the salts being flushed out of the aquifer by groundwater flowing toward the surface water body.

The extent of the band of poor quality elsewhere at Groenvlei is not known. It is only at the western edge that sufficient boreholes and wellpoints exist to compile a transect such as that presented in Figure 5.20.

7.3 Quantification of Evaporation

Probably the most vexing problem of applying a water balance is quantifying evaporation losses, even if measured pan evaporation data are available. Evaporation losses played a key role in the water balance of Groenvlei, and as with many other studies reported in the literature, evaporation was the largest component of the water balance. It accounted for 81% of losses. Errors in quantifying evaporation are directly transposed into the residual of the water balance. As the nett groundwater contribution was the smallest component of the water balance, errors in the evaporation term have a significant impact on the outcome of modelling.

Three methods were used to quantify lake evaporation from pan data, with the calculated annual lake evaporation ranging from 866 mm/a to 1 435 mm/a. Lake evaporation potentially ranges between 2.4 mm/d and 3.9 mm/d, and it is evident this uncertainty will be significant when the nett groundwater contribution is only in the order of 0.2 mm/d and 0.5 mm/d.

The assumption that losses from the vegetative fringe are similar to those of open water has been found to be invalid. The dormancy of the vegetation during the winter months had a significant impact on the water balance, in spite of lake evaporation being small in the winter months (1 mm/d – 2 mm/d) and the reed collar only making up 32% of the wetland. Application of the water balance at Groenvlei, with evaporation and transpiration being treated independently, indicated transpiration from the reed collar could be 30% more than evaporative losses from open water.

After addressing issues related to open water evaporation and transpiration losses from the reed collar, analysis of the results of the daily water balance modelling indicated the method used by Midgley et al. (1994) to quantify lake evaporation provided the most credible results. Results obtained by multiplying A pan data by a coefficient of 0.7 appeared to be slightly too high, but use of a coefficient of 0.6 or 0.65 may be acceptable in this instance as the quantum of the coefficient is known to vary geographically. Use of A pan data without correction to quantified lake evaporation yielded a nett groundwater contribution not supported by Darcian flow calculations, indicating this approach to be inappropriate. The hypothesis that A pan data can be used to approximate lake evaporation is rejected.

The term “lake evaporation” needs to be treated with caution as it is not specifically defined to include or exclude the reed collar. Consequently a distinction was made during this research between evaporation from open water and transpiration from the reed collar. It was interpreted lake evaporation addressed by Midgley et al. (1994) refers to open water evaporation, requiring separate consideration be given to losses from the vegetative fringe.

7.4 Data Errors and Uncertainties

The GIGO [3] principle may have been popularised in the field of computing, but it applies equally to any modelling process. In terms of this study, sources of error and uncertainty can be related either to conceptualisation and / or measurement, and dictate that outcomes be treated with a degree of latitude when trying to interpret them.

The areal extent of the wetland should be a parameter easy to measure, but this is apparently not the case. Martin (1956) (2.48 km^2), Watling (1979) (2.46 km^2), Coetzee (1980) (2.50 km^2), and Fijen (1995) and Roets (2008) (both 2.5 km^2) defined the extent of the wetland by the area of open water. Parsons (2008a) recognised the wetland comprised both open water and the reed collar, and set the area of the wetland at 3.86 km^2. Paridaens and Vandenbroucke (2007) adopted the same approach, but set the area at 3.63 km^2. Vivier (2009) used an area of 3.24 km^2 while this study was based on the wetland comprising an area of 3.59 km^2. Different conceptualisations of the wetland resulted in the estimated extent ranging by almost 56%; while measurement of the wetland defined by open water and the reed collar varied by almost 20%. The implication of this variance is quite significant. The difference in the volume of water from direct rainfall calculated using the area of Martin (1956) (2.48 km^2) and the rounded off value of Fijen (1995) and Roets (2008) (2.5 km^2) equates to a flow rate of 4.25 L/s – enough to change the water level of the lake by 0.18 m.

Errors and uncertainty introduced by measurement of rainfall and pan evaporation were discussed in Sections 3.2. About 17% of the pan evaporation data from K3E003 measured between 1981 and 2012 was missing. Problems with the evaporation data are readily identified and patched, but the same is not true of rainfall data. Comparison of data measured

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[3] GIGO is an acronym for “garbage in, garbage out”

at K3E003 and the Sedgefield fire station highlight this. Missing data or poor measurement and record keeping was material when applying the daily water balance model, but less important at a monthly or annual time scale because of the length of the data record (31 years) and averaging.

Uncertainty related to quantifying transmissivity was probably just as important as that introduced by estimating lake evaporation from measured S pan data, as Darcian flow calculations were used to both support the outcome of the daily water balance model and disaggregate the nett groundwater contribution. Methods used to quantify hydraulic parameters (e.g. infiltrometer test, seepage meter, pumping tests), accuracy of data measurement and interpretation methods used to analyse the data all contributed to the range of values obtained. Further uncertainty was introduced by aquifer heterogeneity and anisotropy. In this study, the average T value derived from pumping tests (270 m^2/d) was used while recognising the uncertainty introduced through tested boreholes only partially penetrating the aquifer.

Transmissivity could vary between 150 m^2/d and more than 400 m^2/d. Using the rain contribution (2.41 Mm^3/a) and evaporation losses (2.86 Mm^3/a) presented in Table 6.11, and the groundwater gradients and flow boundaries used in Section 6.2, it was possible to calculate transmissivity to be in the order of 320 m^2/d. A convergence of evidence points to the aquifer being transmissive, with the upper part of the aquifer having a T value of the order of 300 m^2/d.

The hydraulic gradient of 0.0013 between Groenvlei and the sea is considered to be of high confidence because the level of the lake and the sea act as endpoints, while the elevation of groundwater at three intermediary points has been accurately measured. The hydraulic gradient west of Groenvlei was similarly based on accurately measured data. The gradient north of the lake was determined from groundwater elevation contours based on little measured data and inferences from topography. The same is true of delineating lake inflow and outflow boundaries. Confidence of the inflow hydraulic gradient and the flow boundaries can hence be no higher than medium.

No site-specific data and little regional data are available for quantifying the Cl concentration of rain, used to quantify recharge and in the chemical mass balance model. The estimate of

20 mg/L is generally higher than found in the literature, but local data are available to suggest the estimate could be 25 mg/L. Quantification of the Cl concentration of groundwater flowing into the lake is of high confidence, as is that of the Cl concentration leaving the lake. However, the chemical mass balance suggests that the load of salt leaving the lake is not directly controlled by the concentration of the lake water.

Applying an approach employed by Virdi et al. (2013), the possible error for the nett groundwater contribution determined using the water balance was calculated as 10%. Uncertainty related to the climatic data and the sensitivity of calculations (as illustrated by the rounding off of the lake area) suggests a high error term is warranted. Including the uncertainty of other variables used to quantify the groundwater contribution is more complicated. Qualitatively, it is argued the outcomes of the research are of medium to high confidence because the independent approaches have led to a convergence of evidence, and all the information presented has supported the presented conceptualisation of the hydrology of Groenvlei. The possible error in the relative contributions is summarised in Figure 7.2.

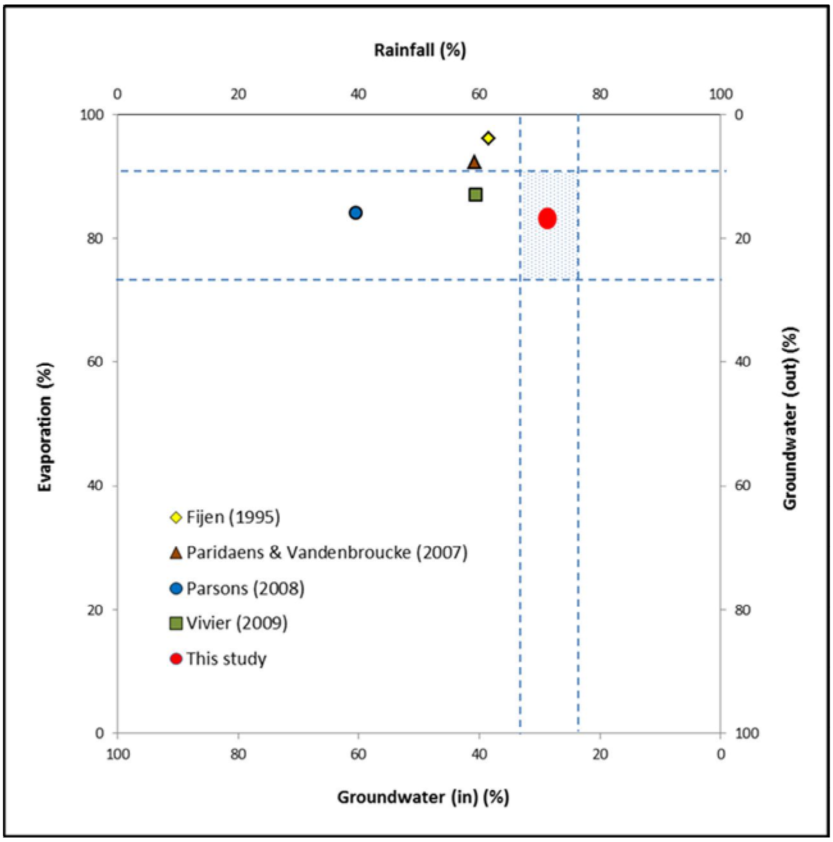

Figure 7.2: Diagram showing the relative contribution of the different hydrological components (red dot) and the interpreted maximum errors of inflow and outflow (stippled line).

7.5 Previous Studies

The results of previous estimates of the water balance of Groenvlei were compared to those of this study (Table 7.1). The range of solutions illustrated the application of a water balance approach is not as straightforward as the simple equation suggests. Fijen (1995) approached the problem from a surface water perspective, while Vivier (2009) applied groundwater models. Paridaens and Vandenbroucke (2007) and this study adopted a more empirical approach. Key differences in the studies was the conceptual understanding of the hydrology of the lake, the existence of surface water inflows, quantification of evaporation losses and leakage from the lake. Also, none of the other studies explored the difference in losses from open water and the reed fringe.

Table 7.1: Comparison of water balance calculations of Groenvlei.

Component	**Fijen (1995)**	**Paridaens & Vandenbroucke (2007)**	**Parsons (2008a)**	**Vivier (2009)**	**This study (2014)**
Lake area considered (km^2)	2.5	3.63	3.86	3.24	3.59
Rainfall (mm/d)	1.79	1.76	1.79	1.78	1.84
Surface run-off (mm/d)	0.86	-	0.00	0.40	-
Groundwater inflow (mm/d)	0.26	1.21	2.74	0.82	0.73
Input (mm/d)	**2.91**	**2.97**	**4.53**	**3.00**	**2.57**
Evaporation (mm/d)	2.95	2.60	3.75	2.88	1.62
Transpiration (mm/d)	-	-	-	-	0.56
Seepage (mm/d)	0.12	0.22	0.71	0.28	0.44
Groundwater abstraction (mm/d)	-	-	0.00	0.15	-
Output (mm/d)	**3.07**	**2.82**	**4.46**	**3.15**	**2.62**
Balance error (%)	5.3	5.2	1.6	5.2	2.3

Both Fijen (1995) and Vivier (2009) interpreted a surface run-off component. This is in spite of no evidence being available to support this, and visual and rainfall – stage data suggesting surface water does not contribute to Groenvlei. An interesting observation by Vivier (2009) implied it did not matter whether there was a surface water run-off component to Groenvlei, as his calculated surface and groundwater components both contributed to the input component of the balance. While numerically correct, it suggests:

- Modelling was done without a good appreciation of the hydrology of the system; and
- Any solutions are provided without cognisance to whether the reasons for the solutions are right or wrong.

A critical difference in the previous water balances is the method used to quantify lake evaporation. Fijen (1995) considered only the open water component of the wetland, while the reed collar was recognised as being part of the wetland by the remaining researchers. Evaporation losses ranged between 2.18 mm/d and 3.75 mm/d, a variance of 72%. Application of three different methods to quantify lake evaporation, and consideration of transpiration losses in summer and winter, led to the conclusion that using lake evaporation coefficients provided by Midgley et al. (1994) produced best results. It is offered evaporation losses used in previous studies were too high.

Fijen (1995) argued seepage losses from the lake were minimal, but the remaining geohydrologically-based studies recognised Groenvlei loses water along its southern boundary. Water discharges into the subsurface and then flows southward to the sea. Parsons (2008a) used a transmissivity of 400 m^2/d in his estimate of seepage losses, but subsequent pumping test data indicated this to be too high. With better delineation of the outflow boundary and a high confidence in the hydraulic gradient, it is probable the outflow from the lake is higher than indicated by Paridaens and Vandenbroucke (2007) and Vivier (2009).

In this study, quantifying the role of groundwater in sustaining Groenvlei was founded on different approaches and ensuring conceptualisation of the hydrology of the lake accommodated all information. The daily water balance and the computed nett groundwater contribution remained central in the quantification process. When trying to balance inflows and outflows using either Darcian flow calculations or the chemical mass balance model, any adjustments had to ensure the nett groundwater contribution remained within the calculated value of about 0.3 mm/d. All models suffer from limitations associated with conceptualisation and the availability of data, but the following considerations point to the current water balance being an improvement on previous efforts:

- The availability of additional data helped improve the quantification of hydraulic properties, hydraulic gradients, flow boundaries and groundwater quality characteristics;
- Evaporation losses took account of differences between evaporation from open water and the reed collar; and
- Chemical processes in the reed collar and hyporheic zone explained observed elevated salinities in wellpoints on the extremities of the wetland.

The ability to quantify groundwater's contribution to the lake with a high degree of certainty is still restricted by the absence of data pertaining to the thickness of the primary aquifer, groundwater level and chemistry data in the eastern parts of the study area, and monitored groundwater data. The understanding of the hydrology of Groenvlei needs to be continually reviewed as new data and information become available.

7.6 Development of Management Tools

It is not helpful to base management decisions on conceptual speculation, as is sometimes the case. Rather, tools can be developed to assist in the management and conservation of Groenvlei based on current knowledge. Using the results of this research, two spreadsheet-based models were developed to model the impact of groundwater abstraction on lake levels and consider the salinity of Groenvlei.

7.6.1 Impact of groundwater abstraction

Parsons (2008a) estimated planned abstraction of 180 m^3/d (0.0066 Mm^3/a) directly west of Groenvlei could lower the lake level by 17 mm. This was based on a groundwater inflow of 3.83 Mm^3/a i.e. substantially more than that estimated during this research. Using flow volumes determined during this research and adjusting both upgradient and downgradient hydraulic gradients as the lake level changes, it was modelled that an abstraction of 180 m^3/d from the Groenvlei catchment would reduce the lake level by 86 mm (Figure 7.3). The lake level would stabilise over a period of 15 years. This calculation assumed steady-state conditions, and ignored irrigation and wastewater returns to the aquifer. It also did not take account of potentially induced recharge and reduction of evaporation losses. As a geohydrologist, it is not possible to contextualise the impact of this on the ecology of the lake and specialist advice has to be sought from an ecologist. In terms of the monitored water level fluctuations, however, the impact of abstraction on lake levels appears small.

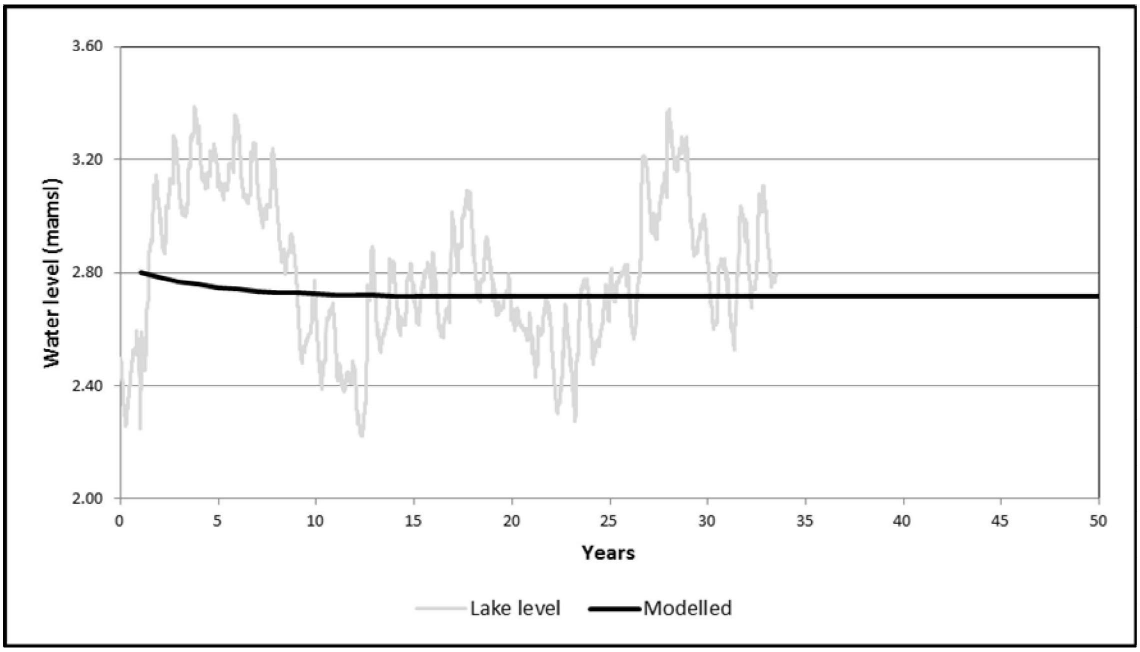

Figure 7.3: Modelled impact on lake levels when groundwater is abstracted from the Groenvlei catchment area at a rate of 180 m^3/d.

The same model could be used to produce a series of type curves for predicting the impact on lake levels when groundwater is abstracted from the Groenvlei catchment (Figure 7.4). While the steady state modelling is simplistic and based on a number of simplifying assumptions listed above, it provides an indication of the scale of impact expected. The type curves could be useful when considering groundwater use license applications, and in devising lake protection and conservation measures.

7.6.2 Lake salinity

Outcomes of the chemical mass balance were used to model salinity changes within the lake. Ignoring climate change and variation during the past 5 000 years, and assuming steady-state conditions, it is shown salinity of Groenvlei evolved rapidly from a saline water body connected to the sea to a system driven by rain and groundwater (Figure 7.5). Assuming an initial Cl concentration of 19 000 mg/L when connected to the sea, it took about 120 years for salinity levels to settle at the current concentration of 1 200 mg/L. This rapid change is conceptually supported by the high flushing rate of 2.7 years presented in Section 7.2.5.

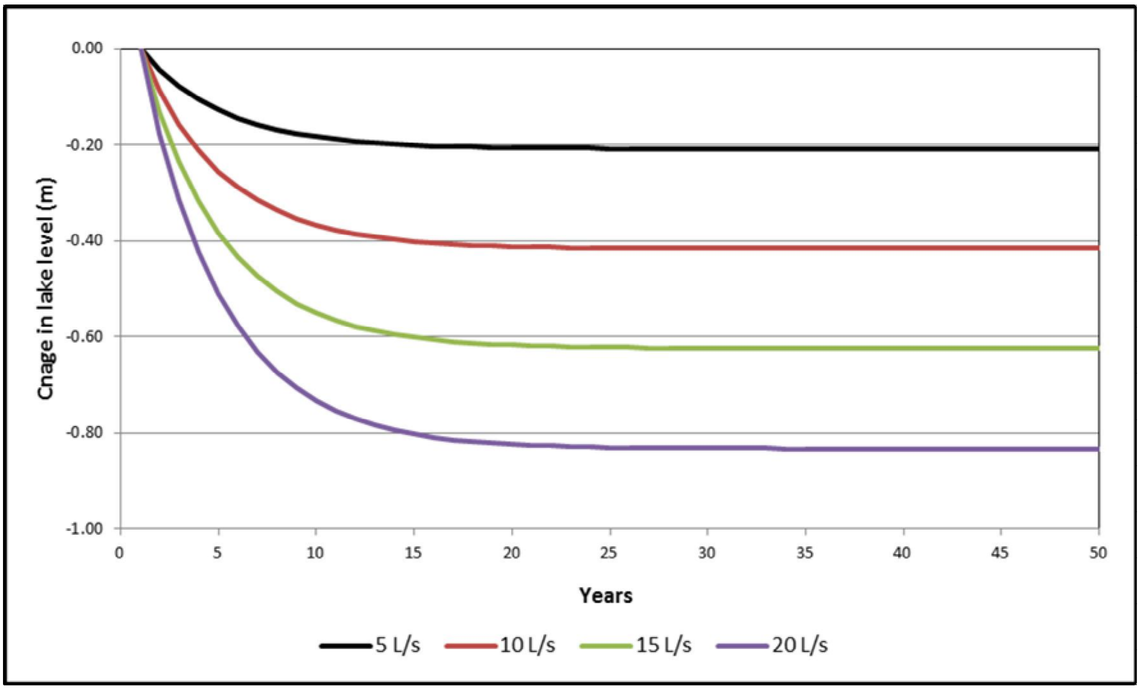

Figure 7.4: Type curves can be used to predict the impact of groundwater abstraction on lake levels. For example, if 15 L/s were abstracted from the catchment, the water level of the lake would drop by 0.62 m over a period of 30 years.

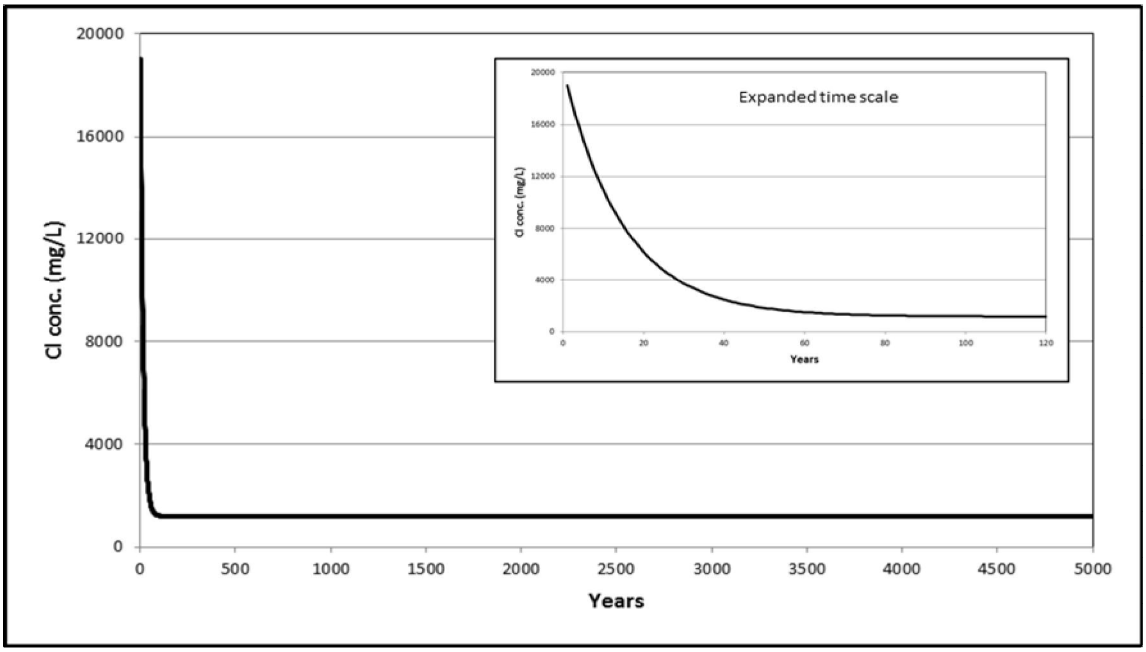

Figure 7.5: Modelled change in Cl concentration since Groenvlei disconnected from the sea to present, with the insert graph providing an expanded view of the change.

Results of salinity modelling allowed consideration to be given to the origin of salts in Groenvlei, first speculated about by Martin (1960a) (Section 2.1.1). Given the rapid evolution of the salinity of the lake (120 years), and a flushing rate of 2.7 years, it is probable all salts captured in Groenvlei when formed some 5 000 years ago have passed through the system. It is improbable salts are residual from when the lake still had marine or estuarine connections. The long disconnect from the sea and the rapid flushing of salts point to the SANBI classification of Groenvlei as an estuarine wetland being inappropriate. Similarly, the hypothesis that the water quality of Groenvlei is reflective of its marine origin is rejected.

Results of monitoring of Cl concentrations in rain (including dry deposition) by Weaver and Talma (2005) indicated concentrations and loads of sea spray are too low to play a significant role in controlling the salinity of the lake. Elevated concentrations of Cl in rain at the coast, however, play at contributory role, and this was taken into account when quantifying the Cl load contributed by rain. Martin's third possible source of salinity was disqualified by the absence of saline springs that flow into Groenvlei. Rather, it is a balance of salts coming into the lake and those leaving the lake that account for Groenvlei's salinity, with the reed collar understood to play a significant role in retaining salts in the system.

The salt balance was also used to predict the impact of groundwater abstraction on lake salinity levels. Using the same example of abstracting 180 m^3/d from the Groenvlei catchment, it was predicted the Cl concentration of the lake would reduce from 1 200 mg/L to 1 182 mg/L over a period of 85 years (Figure 7.6). Type curves were produced to predict lake salinity changes resulting from the abstraction of groundwater from the Groenvlei catchment (Figure 7.7, but expert advice has to be sought to appreciate the effect of these changes on the ecology of the wetland.

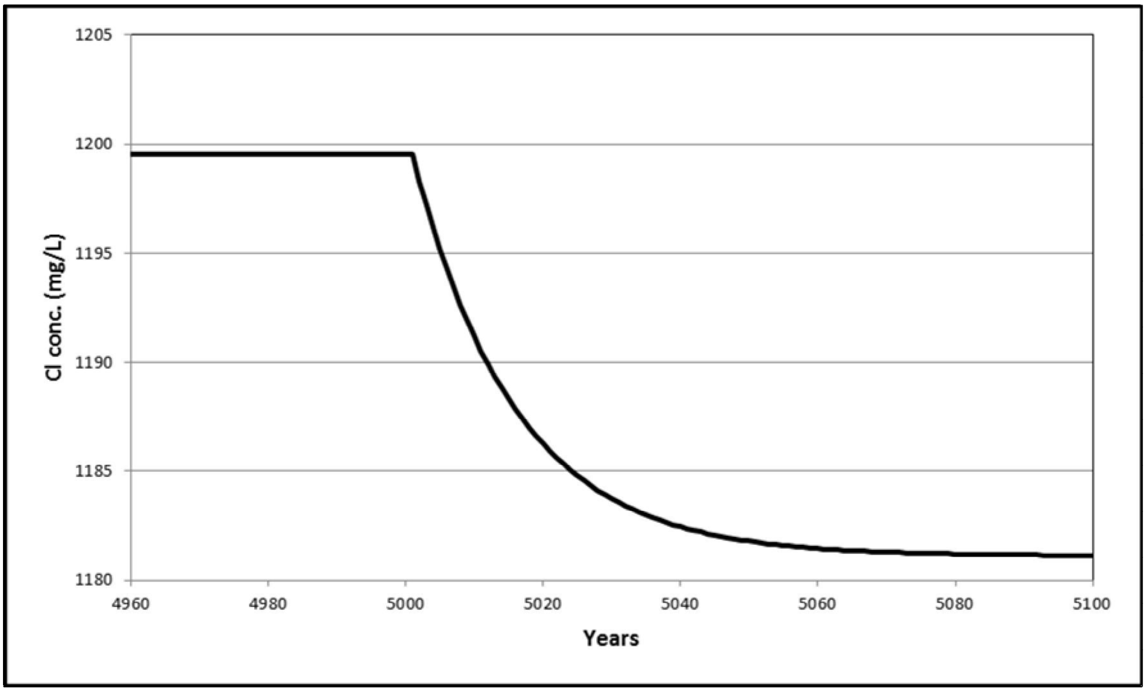

Figure 7.6: Modelled change in Cl concentration of Groenvlei resulting from abstracting 180 m^3/d groundwater from the lake catchment.

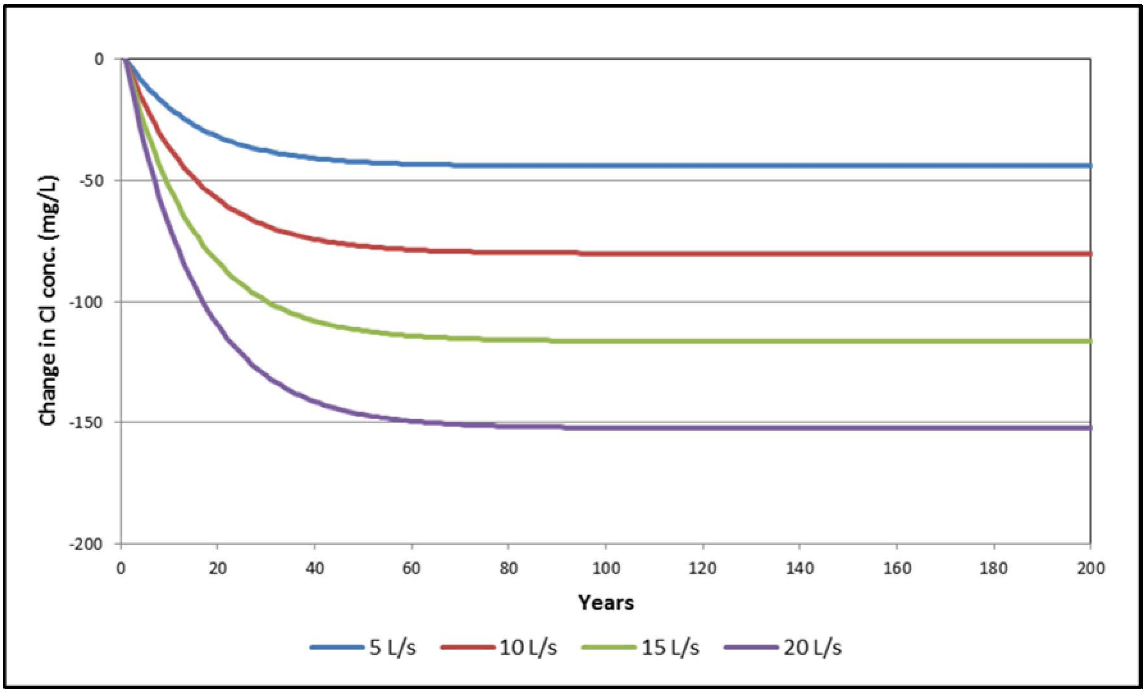

Figure 7.7: Type curves can be used to predict the impact of groundwater abstraction on lake chemistry. For example, if 15 L/s were abstracted from the catchment, the Cl concentration would reduce by 120 mg/L over a period of 80 years.

8 GUIDELINES FOR ASSESSING THE GROUNDWATER CONTRIBUTION TO LAKES AND WETLANDS

An objective of this research was to provide a basis for assessing the role of groundwater in sustaining important coastal wetland systems. From the experience of researching Groenvlei, guidelines have been developed that aim to improve the ability to understand and quantify the role of groundwater in sustaining lakes and wetlands. These guidelines - based on the scientific method where observations or hypotheses are tested against experimental results, measured data or modelled outputs – are graphically illustrated in Figure 8.1.

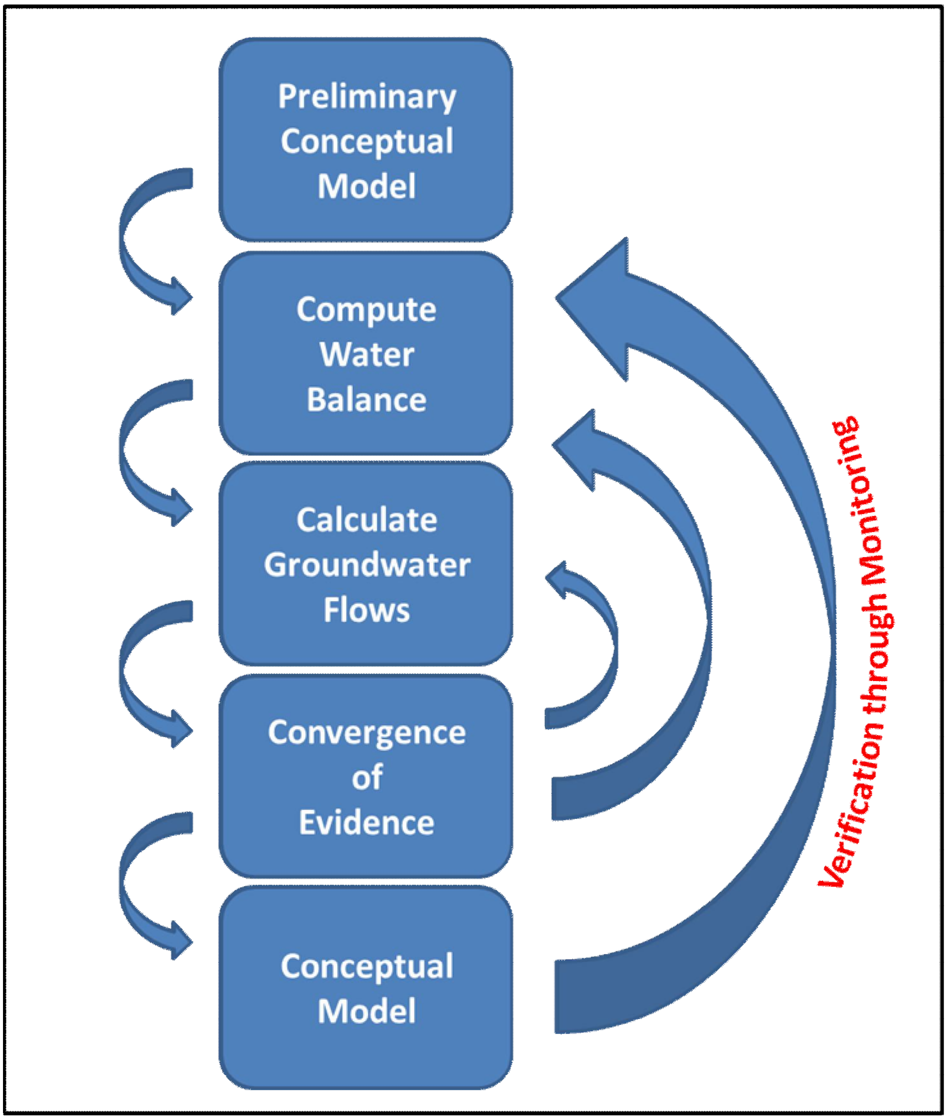

Figure 8.1: Process for quantifying the groundwater contribution to lakes and wetlands.

8.1 Preliminary Conceptual Model

The objective of this first step is to formalise observations or hypotheses regarding the role of groundwater in supporting a particular lake or wetland. The preliminary conceptual model could be in the form of a statement, a sketch or even a simple numeric model and should identify different sources of water, flow directions and quantities and other relevant information. The conceptual model is typically based on the knowledge and experience of the person developing the model, and supported by readily available information.

A sound conceptual understanding of the geohydrological regime is essential. This conceptualisation must be based on accepted geohydrological principles and supported by all available information, including that presented in the literature, site-specific reports and, if available, field measurements. It is best if this aspect is addressed by an experienced geohydrologist.

8.2 Compute Water Balance

The objective of this step is to compile an initial water balance of the hydrological system of interest and quantify the groundwater contribution. The water balance is to include all components that contribute to inputs, outputs and a change in storage and could include:

- Rainfall
- River flow – in and out
- Groundwater – in and out
- Tidal exchange – in and out
- Evapotranspiration
- Surface water abstraction
- Groundwater abstraction
- Lake or wetland water levels
- Other

The water balance is to be based on researched information, expert input and motivated argument. The detail of the balance would usually be dictated by available data, and may range from average annual data to daily or hourly time steps. Often the groundwater component is the unknown component, and the water balance is used to quantify its contribution.

An important activity in this phase is the assessment of the accuracy and reliability of the data used in the water balance. During the Groenvlei study, for example, missing and erroneous data was identified in both the rainfall and S pan data sets. Multiple data sets from different gauging stations could be used to verify data. Alternatively, patching could be used to address missing data.

Note: It should never be assumed that data sourced from databases or other institutions are without error. Checks should be made to verify the data and identify problems.

In some instances modelled data or estimates will be used to inform the water balance. This may be particularly true when rivers are ungauged and when upscaling pan evaporation data to lake evaporation. It is important to assess the applicability of the model used and the certainty of the model output. It is advisable to seek expert input in this process to ensure the best possible approach is used.

Recommendation: Evaporation from open water can be estimated using S pan evaporation measurements adjusted by monthly lake evaporation coefficients provided by Midgley et al. (1994).

Recommendation: If a lake or wetland includes a fringe of reeds, then the vegetative area should be included in the total area of the lake or wetland.

Recommendation: In summer, evapotranspiration from the reed collar is between 10% and 30% more than evaporation from open water. During winter, no (or very little) water is evapotranspired by the reed collar.

The unknown component of the water balance is computed as a residual of the other components. This means errors associated with the other components manifest in the answer. As the groundwater component is often a small part of the water balance, it is important to be aware of the uncertainty associated with the different data sets and ensure the outcome of the water balance study is not over-interpreted.

8.3 Calculate Groundwater Flows

This step usually entails determining the groundwater contribution to the lake or wetland using Darcys Law. As this method is not linked to the water balance, it provides an independent means of quantifying the groundwater contribution. Like surface water, the groundwater component can be modelled using empirical or numeric approaches. Geohydrological data is to be collected to facilitate this:

- Groundwater level data
- Aquifer parameter data
- Direction of groundwater flow and aquifer boundaries
- Recharge
- Groundwater chemistry and isotope data

A hydrocensus is a cost-effective means of collecting groundwater information, but this can be supplemented by geophysical surveys, borehole drilling and aquifer testing, groundwater sampling and ongoing monitoring.

> **Recommendation:** Environmental isotopes can be used to identify or delineate different waters, and waters subjected to different processes or flow paths.

Darcys Law can be used to quantify groundwater flow into or from a lake or wetland. To do this, the direction of groundwater flow needs to be determined from a groundwater contour map and inflow and outflow boundaries delineated. Groundwater flow can then be determined using the transmissivity of the aquifer, hydraulic gradient and the width of the flow boundary.

> **Note:** While Darcy's Law may be easy to apply, the devil lies in the detail. Accurate delineation of inflow and outflow boundaries and aquifer transmissivity are key if the role of groundwater is to be quantified with any certainty.

Recommendation: Unless suitably skilled with contouring software and sufficient data are available, groundwater level contour maps should be prepared by hand to reduce the numeric control of the contouring process and increase a degree of bias introduced by surface water bodies, geohydrological considerations and topography.

8.4 Convergence of Evidence

Because groundwater is intangible and the measurement of flow indirect, this phase of the process aims to ensure all available information and outcomes support the results of the assessment. A Chemical Mass Balance is a particularly useful tool as it is based on both flow and chemistry; and it is difficult to obtain a balance if one or more components are poorly quantified. The process is iterative, and if the evidence is in conflict or in poor agreement, data and parameters used in the previous stage must be checked and adjusted. Application of scientific rigour during this step will promote meaningful results being obtained.

8.5 Conceptual Model

Presentation of the conceptual model after (a) data collection, analysis and modelling and (b) verification against other approaches and tools should reflect the best understanding of the hydrology of the lake or wetland. In addition to highlighting the veracity of the data, the model must acknowledge its weaknesses and areas where additional data or verification is required.

The correctness of the conceptual model is governed by the knowledge and expertise of the person compiling the model, the volume and quality of data on which the conceptual model is based and the scientific rigour of the assessment. Avoidance of conceptual bias is crucial if the system is to be well understood.

8.6 Verification Through Monitoring

The scientific method dictates that there is always room to improve knowledge through ongoing observation, monitoring and testing. Also, the outcome of a particular assessment could become the basis of a new or future project. The monitoring of both groundwater and lake response to hydrological conditions provides a means of verifying the outcomes of assessments and improving thereon.

It is beyond the scope of this thesis to provide a monitoring protocol for observing aquifer and lake behaviour over time, as the detail of monitoring is governed by resources available, the size and extent of the lake or wetland and the importance of the wetland. However, the design of the monitoring program must be driven by the objectives of monitoring. If monitoring is being undertaken to improve the water balance calculations, then the following is to be monitored:

- Rainfall and evaporation
- Lake levels and chemistry
- Groundwater levels and chemistry

The number and location of groundwater monitoring stations around a lake or wetland will be a function of the resources available and the flow patterns in the vicinity of the water body, with the drilling of monitoring boreholes accounting for a large part of the monitoring budget. Use of existing boreholes can be considered if not overly affected by groundwater abstraction. Also, the boreholes need to be at a distance from the wetland so not to be influenced by the water level of the wetland.

Employment of data loggers to monitor both water levels and chemistry (EC, pH, temperature, dissolved oxygen, oxidation reduction potential) means that water level and water chemistry monitoring can be done at hourly or daily intervals. Rainfall and evaporation measurement should be done on a daily basis. The data should be downloaded, captured and verified at least every three months.

The ongoing appraisal of monitored data against the conceptual model will allow the model to be validated. If the conceptual model does not stand up to this scrutiny, then an iterative process of reassessment of the water balance and the verification thereof is required,

9 CONCLUSIONS AND RECOMMENDATIONS

9.1 Conclusions

The primary objective of this research was to develop an understanding of the hydrological functioning of Groenvlei (Section 1.2). This included quantifying the contribution of groundwater to the lake and understanding the chemistry of the system. Reviewing existing information, applying a daily water balance model, analysis of groundwater data and employment of a chemical mass balance model facilitated an improved understanding of the hydrology of Groenvlei. Key findings of the research are highlighted below:

- Groundwater contributes 0.95 Mm^3/a to Groenvlei, accounting for 28.4% of the input;
- Direct rainfall onto the wetland contributes 2.41 Mm^3/a to the system and is the main contributor to the lake (71.6%);
- No (or very little) surface run-off drains into Groenvlei, making demarcation of the lakes catchment based on topography inappropriate.
- Lake evaporation comprises losses from open water (2.12 Mm^3/a) and transpiration from the reed collar (0.74 Mm^3/a);
- During summer the rate of loss from the reed collar is 10% to 30% more than that from open water, but in winter the reed collar is dormant;
- Groundwater discharge along the southern boundary (seepage) amounts to 0.44 Mm^3/a (or 16.9% of losses), invalidating the claim that Groenvlei is endorheic in character;
- Using groundwater elevation data and interpreted directions of groundwater flow, the size of the Groenvlei catchment – or capture zone – was measured to be 25 km^2.
- Only 5.7% of rainfall in the Groenvlei catchment discharges into the lake, and 8.7% of rainfall south of Groenvlei discharges into the sea via the subsurface;
- Climatic extremes play an important role in the hydrology of Groenvlei, with extreme rainfall events causing lake levels to rise rapidly. However, extended periods of above- or below-average rainfall seem to control the overall state of the lake.
- Wind plays an important role in the hydrology of the lake, preventing deposition of fine sediments in the shallow fringes and promoting the mixing of lake water, including the higher salinity water in the reed collar.

- The dynamics of the lake are far slower than typically associated with surface water, but are in line with that of groundwater.
- Groenvlei is a relatively young, dynamic wetland system that responds quickly to change. In its short history it has adapted and responded to changes in both sea level and climate, collectively resulting in the present-day system;
- Notwithstanding relic fauna, Groenvlei has long since lost its hydrological connection to the marine and estuarine environments to which it was once linked;
- The water quality of Groenvlei reflects a balance between the incoming salts from groundwater and rain, salts stored in the lake, salts leaving the system and the re-entrainment of salts from the reed collar and hyporheic zone.
- It is proposed the reed collar and hyporheic zone play key roles in maintaining the current salinity of Groenvlei by retaining salts that are re-entrained into the lake through wind action.

Appreciation of the difference between losses from open water and the reed collar, and dormancy of the reeds during winter, are key outcomes and contributions of this research not considered in previous work. The role of the reed collar in retaining salts has also not previously been addressed, but explains the high salinities measured by Roets (2008) at the edge of the wetland.

The findings of this research can be applied to similar wetlands where the role of groundwater might be less obvious because of river flows and tidal exchange. The importance of direct rainfall onto wetlands, quantification of evaporative losses using S pan data and coefficients prescribed by Midgley et al. (1994), and the relationship between open water losses and transpiration losses are three aspects that could improve the understanding and quantification of lake – groundwater interaction elsewhere.

The ability to quantify groundwater flows using Darcy's Law is governed by the availability of data. Use of T values ranging between 150 m^2/d and 400 m^2/d affected the quantum of groundwater calculated to flow into and out of the lake, and the relative importance of groundwater in sustaining the lake. Great care has to be taken in quantifying parameters, particularly if limited data are available.

As evaporation losses dominate the water balance, sound conceptualisation dictates the ability to understand the hydrology of Groenvlei. This is well illustrated by the disproved claim that Vankervelsvlei and Groenvlei are sustained by discharge from TMG Aquifers. This study highlighted the need to consider lake hydrology beyond the immediately recognisable boundaries of surface water bodies, and that it is necessary to understand groundwater flow patterns in relation to surface patterns. The ability to conceptualise a situation is driven by training, knowledge, experience and available information. A need exists to link hydrologists and ecologists to better understand wetlands, with each contributing specific skills and knowledge.

Groenvlei appears to be in a pristine state, and conservation and protection is required to keep it so. Tools have been developed to assess the impact of groundwater abstraction on the lake, in terms of both lake levels and water quality. These tools have to be verified by monitoring. However, any management strategies and plans for the lake must take cognisance of its young and dynamic character, recognising it has already evolved through sea level and climate change. The current state of Groenvlei is reflective of its present-day hydrological drivers.

Integration of techniques (and specifically independent techniques) and continual assessment of reasonableness of outcomes allowed for a convergence of evidence. While this facilitated an improved understanding of the hydrology of Groenvlei, data errors and the absence of data result in the assessment presented in this thesis still being subjected to a degree of uncertainty. Qualitatively, the confidence of the assessment as medium to high, and recommendations presented in Section 9.2 are aimed at reducing uncertainty and increasing confidence.

The discharge of groundwater into lakes can be investigated using a range of techniques. This study used an analytical approach founded on a water balance and Darcy's Law. Paridaens and Vandenbroucke (2007) used temperature profiling and seepage meters, albeit with limited success. A host of other investigative tools are now available that can confirm and contribute to knowledge of groundwater – lake interaction, including airborne thermal imagery, sophisticated seepage meters, continuous ^{222}Rn measurement and other geochemical tracers. Application of these tools at Groenvlei – and other wetlands – will help improve current knowledge.

9.2 Recommendations

A limitation to understanding the geohydrology of Groenvlei is the lack of information pertaining to aquifer thickness. It is recommended four boreholes be drilled to either bedrock or at least 100 m in depth (whichever is reached first) to quantify the thickness of the aquifer. Drilling could be supported by geophysical surveys if there is sufficient contrast between the primary and secondary aquifers. These boreholes must be subjected to scientifically driven pumping tests with a view of obtaining transmissivity values representative of the entire thickness of the aquifer. Also, samples should be taken and subjected to isotopic analysis, with examination of ^{87}Sr / ^{86}Sr ratios to establish linkage with the underlying TMG aquifer.

This research was hampered by the uneven spatial distribution of data, with much of the data being obtained in the western parts of the study area. Some data were available for the area south of Groenvlei, but little data were sourced from the north and east. In the absence of a directed research effort, a hydrocensus of the catchment should be conducted every 5 years to source new information. Groundwater level elevations should be accurately surveyed.

The ability to model the hydrology of Groenvlei would be greatly enhanced by better information on evaporation from open water and transpiration from the reed collar. It is recommended evaporation measurements using techniques such as eddy covariance and the surface renewal method be conducted at Groenvlei, paying particular attention to the open water and reed areas and differences between summer and winter.

The role of the reed collar in retaining salts has been identified in this research as being important in the salt balance of Groenvlei. This aspect requires further research by investigating salt uptake by the reed vegetation (vegetative assimilation), and spatial and temporal variations of salt concentrations in the shallow, fringe surface waters of the lake, in the hyporheic zone and in the groundwater. Specific attention has to be given to wind conditions and the state of mixing in the wetland.

Geo-electrical profiling at the western edge of Groenvlei could provide useful spatial information on the saline zone transitory between groundwater and the lake. If successful, profiles could be undertaken to the north, east and south to determine the prevalence of the zone in those parts.

Absence of a monitored groundwater record prevents a proper analysis of the temporal relationship between groundwater and the lake. While changes in the lake can be addressed on a daily basis, steady-state conditions have to be considered for the geohydrological component. A monitoring program comprising both groundwater levels and water quality should be established to collect data to further the understanding of groundwater – lake interaction.

Of particular relevance is the impact of the Sedgefield WWTW on Groenvlei. Monitoring of boreholes at the site has detected nutrient contamination, but the rate and direction of movement of the plume is not known. This shortcoming should be addressed by the proposed monitoring.

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Appendix A

Parsons, RP (2009) Is Groenvlei really fed by groundwater discharged from the Table Mountain Group (TMG) Aquifer? Water SA, Vol. 35 No. 5, pp 657 – 662.

Is Groenvlei really fed by groundwater discharged from the Table Mountain Group (TMG) Aquifer?

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Abstract

Vankervelsvlei is a unique wetland located in the stabilised dunes east of Sedgefield. Groenvlei is one of a series of 5 brackish coastal lakes along the Southern Cape coast of South Africa, but is the only one disconnected from the sea. It has been hypothesised that discharge from the underlying Table Mountain Group Aquifer sustains Vankervelsvlei, which in turn discharges into Groenvlei. This paper critically reviews the conceptual model and information on which the hypothesis was based. It is argued that the conceptual model is flawed as it does not take account of topographical and geohydrological conditions prevalent in the area. Analysis of limited hydrochemical data did not explore other possible water sources, and the electrical conductivity characteristics used to confirm the link between the wetlands and the deeper secondary aquifer also apply to 56.3% of boreholes located in a variety of aquifer types across the Western Cape Province. No information is available that supports a link to the Table Mountain Group. Rather, it appears that Vankervelsvlei is sustained by direct rainfall and there is no hydraulic link between Vankervelsvlei and Groenvlei.

Keywords: surface water/groundwater interaction, wetlands

Introduction

Groenvlei is a brackish coastal lake known for its diverse bird life and is one of the best venues for large-mouth black bass angling. It is one of a series of 5 brackish coastal lakes along the Southern Cape coast of South Africa, but is the only one disconnected from the sea. It is located about 5 km east of the holiday town of Sedgefield (Fig. 1). Growing concern about the impact of proposed development in the vicinity of the near-pristine wetland has highlighted the need to understand the hydrological functioning of the system.

Roets (2008) researched the wetland in the context of groundwater dependence of aquatic ecosystems associated with Table Mountain Group (TMG) Aquifers. Together with Roets et al. (2008a; 2008b) and Roets (2009), he contends that discharge from the TMG Aquifer sustains Vankervelsvlei – a unique wetland located in the stabilised dunes east of Sedgefield – and thereafter discharges into Groenvlei. A critical review of this work indicates that the Roets conceptual model is flawed and not supported by basic geohydrological principles and available information. This paper evaluates the conceptual model in view of available information and presents an alternative model that takes account of existing information.

Description of study area

Groenvlei

Groenvlei has been described by Martin (1956), Fijen (1995) and Parsons (2008a). It has a west-east elongated shape being some 3.7 km long and 0.9 km wide. The surface area of the water body is 2.34 km^2 while the surrounding vegetation in and peripheral to the water body covers 1.52 km^2. The lake has a perimeter of 9 000 m, while that of the total wetland is 11 400 m. The maximum depth of Groenvlei is about 5 m, but much of the lake is less than 3.7 m deep (Martin, 1956). Groenvlei is located at an elevation of some 3 m a.m.s.l. on unconsolidated aeolian sands of Pleistocene and Recent age.

Little information is available regarding the thickness of the sands or the nature of the underlying geology. The contact between shale and quartzite of the Kaaimans Group and sandstone and quartzite of the Peninsula Formation of the Table Mountain Group is covered by sand, but is probably located directly west of the wetland (Coetzee, 1979). These rocks are at least of Ordovician age (495-443 Ma). Groenvlei started to form about 17 000 years ago during the last glacial period. Martin (1959) postulated that Groenvlei had an estuarine origin, and was connected to Swartvlei some 5 km to the west about 8 000 years ago. Wind-blown sand deposits covered the area between Groenvlei and the sea some 6 000 years ago, effectively covering evidence of the lake's earlier connectivity to the sea.

The lake does not have any influent rivers, and is fed only by direct rainfall and groundwater inflow (Parsons, 2008a). This is offset by evaporation losses from the lake surface, evapotranspiration losses from vegetation in and peripheral to the water body, and subsurface discharge along the southern shores. Long-term monitoring of the water level of the lake by the Department of Water Affairs and Forestry (DWAF, now the Department of Water and Environmental Affairs) shows levels ranging between 2.25 m a.m.s.l. and 3.40 m a.m.s.l., with a median of 2.76 m a.m.s.l. The water level displays an interannual range of about 0.3 m.

Vankervelsvlei

Vankervelsvlei is located about 2 km north-east of Groenvlei, and was described by Irving and Meadows (1997) as a floating bog. It has no open water and the peat is in the order of 10 m thick. Vankervelsvlei is a rare geomorphological feature located at an elevation of some 150 m higher than Groenvlei.

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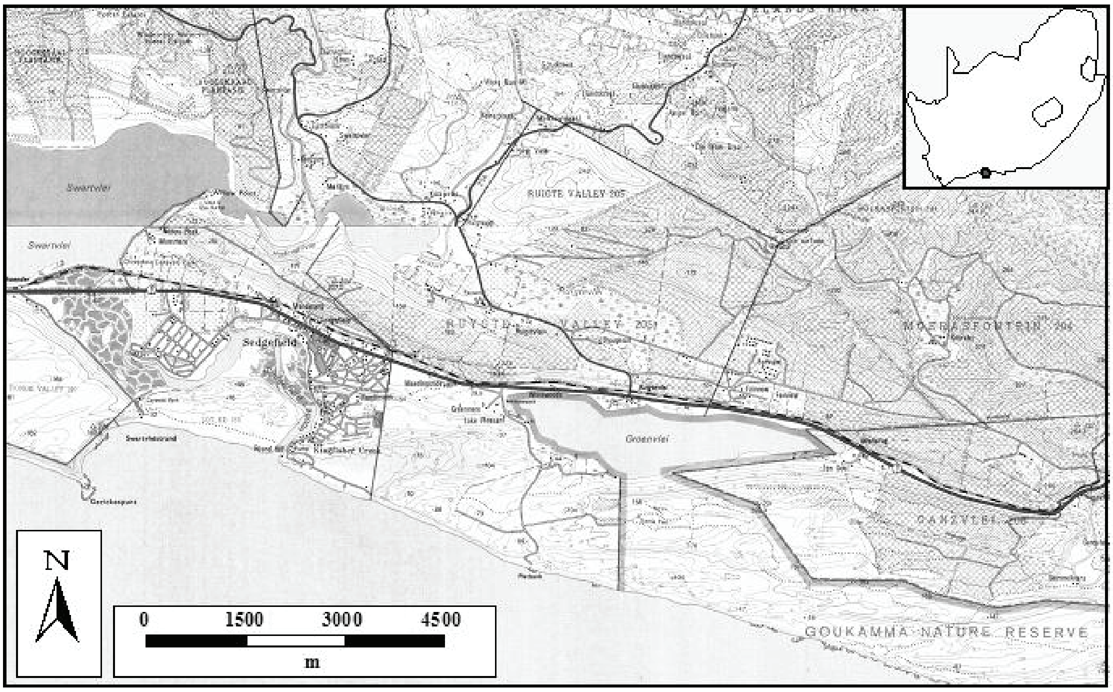

Figure 1
Locality map of the study area, showing Groenvlei and Vankervelsvlei in relation to Sedgefield

The wetland is located on a stabilised aeolian dune described by Illenberger (1996). Irving and Meadows (1997) reported that clay layers at least 3.5 m thick underlie the upper peat layers. The wetland is thought to overlie rocks of the Peninsula Formation of the Table Mountain Group, but the depth to the fractured aquifer system is not known. Vankervelsvlei is at an elevation of 150 m a.m.s.l. and covers an area of only 0.5 km^2. It is an enclosed interdunal depression with no surface water inflows. Water in the wetland is completely concealed by a dense covering of matted sedge vegetation to a depth of approximately 2 m below the surface. The basal sediments of Vankervelsvlei have been dated at 40 000 years old.

Roets conceptual model

Roets et al. (2008b) presented a hypothetical cross-section through the southern Cape coastal belt to illustrate their understanding of surface water/groundwater interaction and the role that the TMG Aquifer plays in sustaining both Vankervelsvlei and Groenvlei (Fig. 2). Their thesis is that groundwater is discharged from the confined TMG Aquifer up into Vankervelsvlei, and then flows into Groenvlei. Analysis of groundwater chemistry data is used to cement the link between the underlying TMG Aquifer and the wetlands. However, available topographical, geological, geohydrological and hydrochemical data do not support this model.

Topography

Critically, the hypothetical section misrepresented the topography of the area and incorrectly portrayed the elevation of Groenvlei in relation to both the sea and Vankervelsvlei (Fig. 3). Groenvlei is at an elevation of about 3 m a.m.s.l., while Vankervelsvlei is at an elevation of about 150 m a.m.s.l. If there is a hydraulic link between the 2 water bodies, the average hydraulic gradient would be in the order of 0.054. This is an extraordinarily steep hydraulic gradient, the likes of which are not reported in scientific literature. Typically, hydraulic gradients range between 0.0005 and 0.01

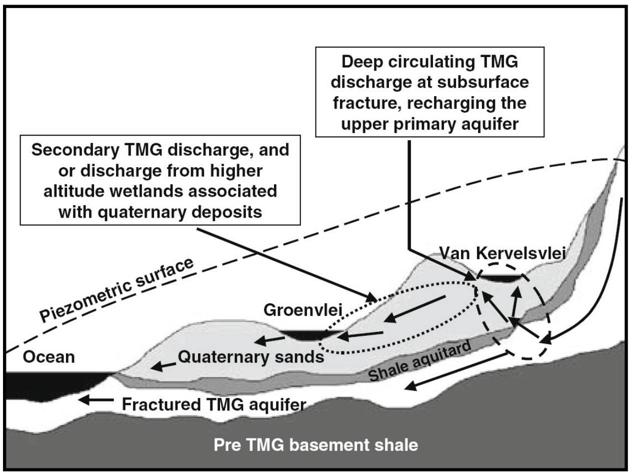

Figure 2
North/south hypothetical cross-section through the southern Cape coastal belt presented by Roets et al. (2008b) to illustrate their understanding of surface water/groundwater interaction and the role of the TMG Aquifer in sustaining Groenvlei and Vankervelsvlei.

(Hartner, 2003). The hydraulic gradients of major primary aquifers in South Africa conform to this range (Table 1). The steep hydraulic gradient postulated in the conceptual model would also result in significant groundwater discharge along the northern slope of the valley between Groenvlei and Vankervelsvlei, a phenomenon not observed in the field. Given the highly transmissive nature of the Sedgefield Aquifer and the measured hydraulic gradients ranging between 0.001 and 0.004 reported by Parsons (2005), a hydraulic gradient in the order of 0.05 is refuted.

Groundwater levels

Seven deep boreholes have been drilled into the stabilised dunes within 2 km of Groenvlei (Parsons, 1997; 2004). All are

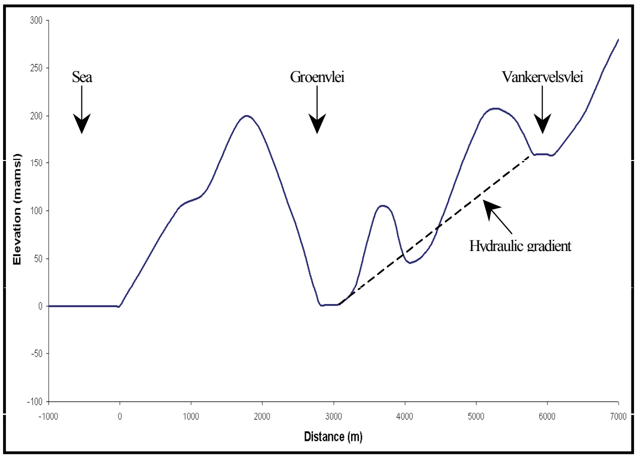

Figure 3
North/south topographical cross-section through Groenvlei and Vankervelsvlei, with a 10x vertical exaggeration. The steep hydraulic gradient of 0.054 required to link Vankervelsvlei with Groenvlei is refuted.

Table 1 The average hydraulic gradient of major primary aquifers in South Africa		
Aquifer	**Hydraulic gradient**	**Information source**
Atlantis	0.015	Fleisher (1990)
Cape Flats	0.004	Wright and Conrad (1995)
Grootfontein (Yzerfontein)	0.014	Timmerman (1985)
Langebaan Road	0.004	Dyke (1992)
Saldanha	0.003	Du Toit and Weaver (1995)
Zululand	0.004	Meyer et al. (2001)

Table 2 Deep boreholes drilled into the stabilised dunes around Groenvlei				
Borehole	**Elevation (m a.m.s.l.)**	**Borehole Depth (m)**	**Groundwater level (m b.g.l.*)**	**Groundwater level (m a.m.s.l.)**
BH1	59.3	72	55.6	3.7
BH2	66.9	78	63.2	3.7
BH3	53.9	66	50.4	3.5
BH4	144.1	162	142.2	1.9
BH5	144.5	162	143.1	1.4
BH6	79.9	84	77.8	2.1
BH7	41.3	47	39.4	1.9

(from Parsons, 1997; 2004)
**m b.g.l. - metres below ground level*

located west of the water body. In all instances, the groundwater level was measured to be within 3.7 m of mean sea level (Table 1). Further, groundwater level monitoring has revealed that groundwater levels display little interannual variation (~0.2 m) (Parsons, 2006). Given that Vankervelsvlei is located in the same geological and geohydrological setting as that into which the boreholes were drilled, the water levels of 148.6 m a.m.s.l. and 148.8 m a.m.s.l. measured by Roets (2008) in piezometers VKA and VKB at Vankervelsvlei are in all likelihood perched water levels, and not representative of a regional water table or piezometric surface. The perched condition is in line with the description of Vankervelsvlei presented by Irving and Meadows (1997) and means that the water body is disconnected or detached from the underlying groundwater system. This also refutes the hydraulic link between Groenvlei and Vankervelsvlei.

It is noteworthy that the groundwater levels measured in piezometers VKA and VKB do not support the piezometric surface illustrated in the Roets conceptual model (Fig. 2). Except for water levels measured by himself in shallow piezometers at 10 locations around Groenvlei and Vankervelsvlei, Roets (2008) and Roets et al. (2008a; b) fail to present or refer to any groundwater level data in support of the groundwater flow patterns illustrated in Fig. 2. The piezometric surface indicated in the conceptual model would result in widespread artesian conditions if artesian conditions existed in the manner required by the conceptual model to discharge into Vankervelsvlei. It is recognised that the piezometric head can only be measured if boreholes penetrate through the confining layer and into the confined aquifer. None of the boreholes in the vicinity of the wetlands achieve this. Further, artesian conditions are not reported in any of the regional geohydrological assessments of the area (Meyer, 1999; Parsons and Veltman, 2006). Given the absence of artesian conditions in the region and the measured borehole data presented in Table 2, it is not possible for discharges from the TMG Aquifer to lift or push up more than 150 m through permeable sand to sustain Vankervelsvlei.

In addition to TMG groundwater discharging from depth upwards into Vankervelsvlei, the conceptual model also requires water to be discharged from the 0.5 km^2 wetland downward into the subsurface to create and maintain the hydraulic link with Groenvlei. It is simply not possible for the opposing groundwater flow directions to co-exist in such a small area.

Geology

Roets (2008) chose to rely on national geological shape files presented in ENPAT and used by Fortuin (2004) rather than the published 1:250 000 geological map of the area (Coetzee, 1979). The position of the contact between the Kaaimans Group and the Table Mountain Group presented by Fortuin (2004) does not correlate with that presented by the more authoritative and site-specific Coetzee (1979). Use of the simplified 1:1 000 000 scale geological shape file from the WR90 data set (Midgley et al., 1994) to define geology at a local scale is problematic. Some correlation exists with the ENPAT data set and that presented by Coetzee (1979), but the principle lithology of a geological unit is presented as opposed to the stratigraphic groupings. This may account for the apparent confusion regarding whether the Peninsula Formation or the Nardouw Subgroup underlies the wetlands.

Notwithstanding these differences, the geological detail presented in Fig. 2 does not conform to any of the geological information referred to above. In all likelihood, both Groenvlei and Vankervelsvlei are underlain by rocks of the Table Mountain Group (and more specifically the Peninsula Formation). Neither the thickness of the sand (depth to hard rock) nor the lithology of the underlying aquifer is known. The presence of a laterally extensive 'shale aquitard' is pure speculation and without foundation. The aquitard is required to support the thesis that Groenvlei and Vankervelsvlei are sustained by discharge from the underlying TMG Aquifer as the confining layer provides the mechanism needed for groundwater from depth to flow upwards and into the wetlands. The relatively thin and distinctive Cederberg Formation, which could act as the aquitard, has not been mapped in the vicinity of the 2 wetlands (Coetzee, 1979), and interpretation of the geological map indicates that they are underlain by arenites of the Peninsula Formation. The map also indicates that the Table Mountain Group is folded. Consequently, it is improbable that the Cederberg Formation would have the near-horizontal orientation indicated in Fig. 2. The absence of the Cederberg Formation further undermines the validity of the conceptual model, and portrayal of the TMG Aquifer as a confined aquifer is a misrepresentation of prevailing geohydrological conditions.

Roets (2008, p 123) states that 'Groenvlei and Vankervelsvlei are lowland wetland systems associated with major east-west running fault systems located to the south of the Outeniqua Mountains.' At no point does he indicate where these major fault systems are, and none are indicated on the geological map of the area (Coetzee, 1979). Faulting of the Cederberg Formation is a key component of the conceptual model, as it provides the mechanism for groundwater in a confined aquifer to discharge into overlaying primary aquifer system and Vankervelsvlei. The absence of any indication of faults in the area advocates this aspect of the conceptual model to be speculative and the above statement to be without foundation.

Chemical character

Roets (2008) argues that because the electrical conductivity (EC), pH and concentrations of Na, Cl and Fe of water sampled from shallow wellpoints at 2 sites at Vankervelsvlei and 3 sites around Groenvlei – all collected on 30 July 2006 – fall within ranges presented by Brown et al. (2003) as being typical of groundwater from the TMG Aquifer, the sampled water must originate from the deeper secondary aquifer system. In their report, Brown et al. (2003) presented a table from Smith et al. (2002) that displayed the mean, minimum and maximum for 13 chemical parameters from 75 boreholes drilled into the Nardouw Subgroup and 28 boreholes drilled into the Peninsula Formation in the Klein Karoo.

By simply comparing 5 chemical parameters measured at the 5 sites at Vankervelsvlei and Groenvlei to ranges considered by Brown et al. (2003) as being typical of groundwater from TMG aquifers, Roets (2008, p. 138) concludes that 'the hydrochemical data of the groundwater from this study suggests that Vankervelsvlei and Groenvlei are dependent on groundwater from the TMG Aquifer.' DWAF's hydrochemical database contains 14 377 EC records from boreholes located in a variety of aquifer types across the Western Cape Province. Of these, 56.3% have an EC in the range of 9 to 155 mS/m. Consequently, this EC range is not unique to TMG Aquifers and cannot be used for hydrochemical 'finger printing' purposes.

A review of the range of EC from 4 major coastal primary aquifer systems across South Africa shows that the EC range displayed by the small data set collected by Roets (2008) is within the range of EC displayed by the Sedgefield Aquifer, the Atlantis Aquifer, the Cape Flats Aquifer and the Zululand Coastal Aquifer around St Lucia (Fig. 4). Only the Sedgefield Aquifer is partially underlain by rocks of the Table Mountain Group, with the other aquifers having no connection to the Group at all. It is far more likely that groundwater quality characteristics displayed by Roets's data are typical of coastal primary aquifers rather than TMG Aquifers. As the chemical

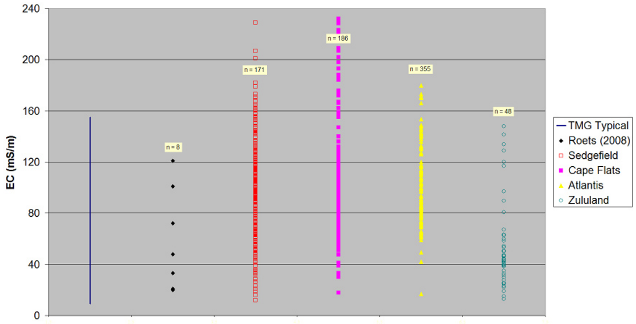

Figure 4
The range of EC displayed by groundwater in four major coastal primary aquifers, compared to data presented by Roets (2008) and the EC range considered by Brown et al. (2003) to be typical of groundwater from the Table Mountain Group Aquifer

character typical of coastal primary aquifers across South Africa overlaps with that of TMG groundwater, the relation is coincidental rather than causative. The interpretation that the chemical character of groundwater adjacent to the 2 wetlands is indicative of a TMG Aquifer link is wrong.

Roets (2008, p. 139) attempts to cement his argument that Vankervelsvlei is fed by groundwater from the TMG Aquifer by attributing the iron (Fe) concentration of 382.23 mg/ℓ recorded in VKB (11 m) to 'the presence of the Nardouw Formation'. Notwithstanding the fact that the geological map of the area indicates that Vankervelsvlei is underlain by the Peninsula Formation and not by the Nardouw Subgroup and significantly lower Fe concentrations were recorded at shallower depths at the same site (0.5 m = 9.5 mg/ℓ; 1 m = 7.0 mg/ℓ), Fe concentrations such as these are unprecedented. High Fe concentrations of some 30 mg/ℓ monitored by Parsons (2008b) at Arabella Country Estate are thought to represent the upper levels of Fe in groundwater, while Smith et al. (2002) set the maximum Fe concentration of the 2 geological units at 0.2 mg/ℓ and 15.4 mg/ℓ. A concentration an order of magnitude higher than these maxima is hence improbable. If the value of 382 mg/ℓ is not a laboratory-related error, then Fe contained in the vegetative mat in Vankervelsvlei deserves closer scrutiny.

Roets (2008) compares his groundwater quality data to the chemical character of groundwater from the Nardouw Subgroup. Interpretation of the 1:250 000 scale geological map of the area (Coetzee, 1979) suggests that Groenvlei and Vankervelsvlei are underlain by the Peninsula Formation and not the Nardouw Subgroup. The EC range presented by Brown et al. (2003)) for the Peninsula Formation is much narrower (3 mS/m to 26 mS/m) than for the Nardouw Subgroup (9 mS/m to 155 mS/m), and only 25% of Roets's data fall within the Peninsula Formation range.

Discussion

Roets (2008) and Roets et al. (2008b) used multivariate cluster analysis of mean EC values to group piezometer positions displaying the same or similar EC characteristics. They over-interpreted the results by stating that the groundwater at the different locations has a 'shared groundwater source', as the statistical analyses merely point to a similar character. They make similar and repeated claims that all data presented by them points to the 2 wetlands being dependent on groundwater from the TMG Aquifer, and the 2 wetlands being hydraulically linked. Based on available information and an evaluation of the information

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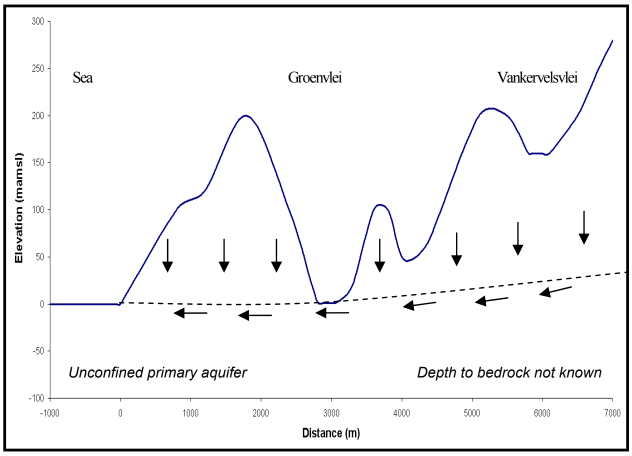

Figure 5
An alternative conceptual geohydrological model of the study area based on available topographical and hydrological data

presented by them, they have failed to provide any scientifically credible evidence that either Vankervelsvlei or Groenvlei are fed by discharges from the TMG Aquifer or that the 2 wetlands are hydraulically linked. Consequently, any conclusions and recommendations based on their hypotheses are without standing.

A major weakness of the reviewed research of Vankervelsvlei is that it did not consider the fact that the floating bog is solely sustained by direct rainfall. The absence of surface runoff, the expected depth of the regional water table and the low ECs measured by Roets (2008) support this interpretation. It is noteworthy that Parsons (2008a) estimated that direct rainfall accounted for almost 40% of the inflow into Groenvlei, while direct rainfall accounts for 38% of the freshwater input into Lake St Lucia (Van Niekerk, 2004). Both Irving and Meadows (1997) and Roets (2008) reported that Vankervelsvlei has no open water. The wetland can be sustained by a relatively small volume of water as the impermeable character of the base of the vlei prevents or restricts the downward percolation of water, and the absence of a dry season results in the vlei being continually recharged by rainfall. Water retention in the vlei is further enhanced by the dense sedge vegetation concealing the water surface (and hence reducing direct evaporation losses) while the organic nature of the soils promotes soil moisture retention. The rainwater-fed theory is both simple and plausible while the convoluted TMG Aquifer-fed theory is not supported by available geohydrological data or geohydrological principles.

An alternative conceptual model based on available geological and geohydrological information is presented in Fig. 5. It is acknowledged that the depth to bedrock is not known, but based on information presented by Coetzee (1979) it is likely to comprise rocks of the Peninsula Formation. The unconfined primary aquifer system is fed by recharge from rainfall that percolates through the vadose zone until it reaches the regional water table of the primary aquifer. The regional hydraulic gradient is typical of that of transmissive primary aquifers and is in the range of 0.001 to 0.004 reported by Parsons (2008a). Depth to the regional water table at Vankervelsvlei is predicted to be in the order of 145 m below ground level. Vankervelsvlei is fed by direct rainfall only, with a low permeability clay base preventing or retarding the downward percolation of water in the wetland to the aquifer. By contrast, Groenvlei is fed by both direct rainfall and groundwater. There is no hydraulic link between the 2 wetlands, with Vankervelsvlei being described as a disconnected system. Groenvlei is a flow-through system, as described by Born et al. (1979).

It is not claimed by Roets (2008) that Groenvlei is fed directly from the TMG Aquifer. Rather it is hypothesised that TMG groundwater discharges into Vankervelsvlei and then discharges from the wetland into Groenvlei (Fig. 2 –'secondary discharge'). However, given Groenvlei's low elevation and proximity to the coast, it is theoretically possible that deep circulation in the TMG Aquifer could discharge into the Sedgefield Aquifer, and thereby sustain Groenvlei. The study of such a theory will require the drilling of a large number of deep boreholes into the TMG Aquifer, supported by hydrochemical and isotopic examination. Until such time that there is any evidence of Groenvlei being sustained by discharges from the TMG Aquifer, such theories should be treated with circumspection.

Conclusions

It is concluded that Roets (2008; 2009) and Roets et al. (2008a; b) failed to provide any credible scientific evidence that either Vankervelsvlei or Groenvlei are fed by discharges from the TMG Aquifer or that the 2 wetlands are hydraulically linked. The conceptual model presented by them is not supported by available geohydrological data or geohydrological principles. Their thesis is based on speculation and lacks scientific rigour, having failed to consider water sources other than that from the TMG Aquifer. The interpretation that the chemical character of groundwater adjacent to the 2 wetlands is indicative of a TMG Aquifer link is wrong, as the character typical of coastal primary aquifers across South Africa overlaps with that of TMG groundwater. There is no hydraulic link between Vankervelsvlei and Groenvlei, and the former wetland is fed only by direct rainfall. Groenvlei is fed by direct rainfall and groundwater, and the possibility that it is being fed by the underlying TMG Aquifer requires further research before it can be given credence.

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Appendix B

Electronic Appendices

B1	Water balance data
B2	WISH database
B3	Summary rainfall data
B4	Borehole EC profile data
B5	Isotope and radon data
B6	Monitored groundwater level data
B7	Groenvlei models