
GROUNDWATER / SURFACE WATER RELATIONSHIPS WITH SPECIFIC REFERENCE TO MAPUTALAND

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by

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EXECUTIVE SUMMARY

South Africa lies in the subtropical region aligned with all the major deserts of the southern hemisphere. This places it the 11th country, out of 150 countries, having the least amount of renewable water available per person¹. Consequently, water resources play a fundamental role in the sustainable development of Southern Africa and their management is therefore an issue of national importance. Informed management requires an understanding of the resources and the underlying processes that control their utilization. This is particularly important with so many social and environmental constraints on the availability of water.

Until 1998, water resource management adhered to the principle of public and private water associated with riparian rights. With some exceptions, groundwater has not featured prominently as a major water resource and nearly all national control was directed at surface water resources. In most areas, the groundwater was not deemed to be of sufficient importance and was therefore classified as private water and the rights to it were linked to the land ownership. With this management legacy, groundwater has not been developed to any significant extent in most areas of the country. The need to change the management strategy in a water-scarce country was recognized many years ago but this has only happened when new legislation was introduced in 1998 that recognizes all water as one entity (described by the hydrological cycle) under national ownership regulated by central government. However, the recent legislation does not specifically define groundwater and it makes little direct reference to geohydrological aspects. The legislation does recognize that groundwater is an integral part of the hydrological cycle in that it contributes to rivers, lakes, wetlands and estuaries as well as supplying water to vegetation in the near surface zone and it is a resource that can be abstracted (either naturally or artificially) long after recharge has taken place. This formal recognition of the continuity in the hydrological cycle requires a clear understanding of the differences and linkages between the surface water resources and the groundwater resources. **This report attempts to describe the water resources and their interaction at the interface between groundwater and surface water in an attempt to overcome differences in conceptual models and to understand the processes and methodologies that describe the interaction between components of these resources.** The project has focused on the Maputaland Coastal Plain where there is an extremely strong interaction between the surface water features and the local aquifers.

The report starts by trying to define the various components of the different resources to overcome the many different terms and definitions that are in conflict between the surface water and groundwater practitioners. This is highlighted by conflicting terms which can have opposite meanings for the same process. Such as the case when considering the flow of surface water (discharged) into the groundwater (recharged). This report includes a summary of the definition of the resource components and a description of their interactions derived from the scientific literature.

A review of methodologies and conceptual models that have been examined and applied to quantify the groundwater and surface water interactions at the University of Zululand follows the description of the various resources. Many of the methods and concepts have formed part of student studies on the

¹Molapo P, Pandey SK and Puyô S (2000) *Groundwater resources management in the SADC region: A field of future regional co-operation*. In: *Groundwater: Past Achievements and Future Challenges*. Sililo et al. (eds.), Balkema, Rotterdam

Maputaland and Mozambique coastal plain. Some of these have been included as case studies in **Part 2** of this report or have been published in journals. Some of the examples highlight difficulties with conceptual models or illustrate the implementation of methodologies. Examples include the upscaling of parameters to regional aquifers such as hydraulic conductivity, spatial and temporal interpolation of measurements and some interpretation of information.

One of the most important issues relating to surface water/groundwater interaction involves the estimation of water transport or flux through the intervening surface. Many studies have attempted to define the groundwater recharge process and methods for estimating the recharge rate. No attempt is made to repeat these methods. Instead, conceptual models and methods derived or tested in the studies on the Maputaland Coastal Plain are explained in the report. Similarly, there are many models describing the evapotranspiration process and methods for estimating the rates under different conditions. All of these relate to the unsaturated zone which are important features in surface water models but are not discussed in this report. Few examples of evaporation models or measurements from the saturated zone were found. Many studies and models of groundwater systems combine the percolation (recharge) and evaporation components into a single parameter. However, it is difficult to reconcile this method with studies of the surface water/groundwater interaction so alternative methods have been discussed and modelled in this study. In particular, methods for incorporating the percolation and evaporation in numerical models of the surface water/groundwater interaction are discussed and examples given in the text and case studies.

While many studies have been done to determine or estimate (through models) the vertical fluxes through the surface leading to infiltration, percolation and evaporation, there is less information of the methods and model for describing the interaction between aquifers and their groundwater discharge boundaries such as rivers, wetlands and lakes. Methods for examining the flux between coastal lakes and their application in the analyses of many of the Maputaland Coastal Plain lakes and streams are described.

The most pragmatic approach to describing the interaction between surface water and groundwater is through the use of models. The mass balance approach is commonly applied but is generally reliant on an estimate of at least one unknown flux such as evaporation. The main components of the mass balance approach are discussed and their estimation described. However, these studies illustrate that the most holistic approach to describing the interaction is through the use of appropriate conceptual models supported by numerical models. Nevertheless, some attempts to measure and understand the interaction are included in the text. The numerical models commonly available for simulating the interaction between surface water and groundwater systems are summarized in the report. These have been used to evaluate conceptual models of systems (lakes), and simulate mass balance components of lakes and rivers.

The difficulty in identifying unique or unambiguous terms or features will persist while the two fields of science, namely surface water and groundwater, continue to remain isolated in many forums. Consequently, a brief **Glossary of Terms** has been included in the report.

Part 1 of this report describes the main concepts in the surface water/groundwater interaction. The initial descriptions of the groundwater system and the conceptual models relating to specific surface water resources (interactive boundaries) have been described in relation to experience in the evaluation

and quantification of the Maputaland systems. **Part 2** of the report describes some specific models and analysis of the system from studies done during the project. Many studies have been conducted on the Coastal Plain, most of which involve varying levels of numerical modelling, to understand and evaluate the surface water/groundwater interaction. Time and space have limited the case studies included in this report to five. Two additional studies are listed but the results of these have been published in recognized journals and are not duplicated in this report. Many other studies are ongoing and will be reported in journals as time progresses. The individual case studies are preceded by a description of the study area.

The 1st case study summarizes an effort to describe and model the preferential pathway in a catchment when water flows from interception to discharge through stream channels. This case study forms part of the Ph.D. work done by Mrs N Snyman.

The 2nd case study examines examples of the application of numerical methods for estimating the groundwater component of streamflow at various scales.

The 3rd case study illustrates the observed impact of land use on the hydrological status of a small coastal catchment. The inferred impact of afforestation on the groundwater has led to the complete cessation of baseflow in the river that has caused the demise of the Siyaya Estuary.

The 4th case study looks at classical hydrograph separation techniques to estimate the contribution of the various preferential flow paths in river runoff in two contrasting settings. The first example uses the hydrographs from the Ntuzi catchments in the Ngoye Hills while the second example uses the river runoff from a small catchment in Richards Bay.

The 5th case study examines the simultaneous hydrographs from the river and groundwater in Richards Bay in an attempt to link the surface water and groundwater processes.

The 6th case study looks at the development and application of a three dimensional transient groundwater model for evaluating the interaction of the lake and streams with the surrounding aquifer.

The 7th case study looks at the role of groundwater seepage into Lake St. Lucia and its function in moderating the effects of salinity fluctuations that play an important role in the ecological resilience of the system.

Many references to two other case studies are included in the main text of the report. A case study looking at the interaction between Lake Mzingazi and the surrounding aquifer has been published in separate reports. The case study that examines the importance of groundwater in sustaining the ecological function of Lake St Lucia has also been published in many journal articles. Consequently, these case studies have not been described in this report.

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ABBREVIATIONS

BHN - basic human needs
CMA - Catchment Management Authority
CRD - comprehensive Reserve determinations
DRD - desktop Reserve determination
DWAF - Department of Water Affairs and Forestry
EIS - ecological importance and sensitivity
EMC - ecological management category
GMA - groundwater management areas
GRU - geohydrological response unit
HGU - hydrogeomorphic unit
HRU - homogeneous response unit
HSU - hydrostratigraphic unit
ICM - integrated catchment management
IFRs - in-stream flow requirements
IRD - intermediate Reserve determination
mbc - metres below collar
mS/ m - milliSiemens per metre
NGDB - National Groundwater Database
NWA - National Water Act (Act No 36 of 1998)
NWRS - National Water Resources Strategy
PAV - permissible annual volumes
PERM - Preliminary Ecological Reserve Method
PES - present ecological state
PRD - preliminary Reserve determinations
RDM - resource-directed measures
RQO - resource quality objectives
RRD - rapid Reserve determination
SDM - source-directed measures
SWR - significant water resource
WRSS - water resource situation study

PREFACE

This project has focused on the importance of groundwater and its relationship to surface water bodies. To assess the importance of groundwater on surface water bodies it is first necessary to identify (or classify) the difference between these entities and then to determine the relationship that links them before examining methods for quantifying the linkages and relationships. However, it has been very difficult to differentiate between the various resources because, in terms of their original definitions, a clear distinction or boundary between the resources cannot always be established. For example, wetlands are generally found to be components or extensions of the groundwater.

Groundwater historically has been classified and treated in an independent manner to surface water (Water Act of 1956). Consequently, the interactions between these resources have not been well researched or documented. The independent treatment of these two resources by hydrogeologists and hydrologists has led to variations in terminology and conceptual understanding of their interactions and relationships. However, the National Water Act (Number 36 of 1998) has unified all water bodies into one through the hydrological cycle, in which *'rain falls from clouds, flows over the land or sinks through the ground, where it may be stored as groundwater in underground aquifers, and finally flows through rivers, lakes and dams towards the sea. Evaporation from surface waters and the transpiration of plants and trees feeds the clouds and the cycle continues'*. Unfortunately, the National Water Act (NWA) does not define groundwater but does specify underground water. It is important to recognize that groundwater and underground water have different meanings to different people and represent varying components in the traditional sense. Consequently, this report attempts to unravel the terminology in the assessment of the roles and relationships of the various water resources.

The NWA, in its recognition of the continuum of water resources, defines the hydrological cycle and the interdependence of its elements, where evaporation, clouds and rainfall are linked to underground water, rivers, lakes, wetlands and the sea. The linkage between the 'underground water' and the other components forms the basis of this report. It should also be noted that the basic hydrological unit for surface water is the catchment which is seldom synonymous with the groundwater recharge area and may differ considerably in area and position for rivers and aquifers.

The NWA has still found it necessary to define several distinct legal classes of water resources for the purpose of water conservation and protection management. These have been defined by the NWA as water courses, estuaries and aquifers. What is not specified in the legislation is the (linkages) interaction and interdependencies between these classified resources. While it is specified that aquifers feed the rivers, springs, channels and wetlands (and vice versa), the nature and extent of these linkages still need to be qualified and quantified. The aquifers are fed by the surface water resources through many diverse pathways that involve multiple combinations of processes which include interception, infiltration, percolation, evaporation, transpiration, seepage and runoff. While the physical interactions form an important component of the linkages between resources, in many ecological systems it is not clear with which resources the ecosystem dependency lays, and hence which classified resource is essential in sustaining the ecological function of the aquatic environment. In order to allocate water in accordance with the requirements of the NWA, it is imperative that the interdependencies are realized and incorporated into the legal framework for the allocation of water from the various resources.

Part 1: Concepts and Methodology

CHAPTER 1: BACKGROUND - THE IMPORTANCE OF GROUNDWATER

Water resource management in South Africa has been undergoing considerable changes over the past decade. These changes are being driven by a need to protect and conserve the limited water resource while promoting sustainable development in an equitable and democratic system. The changes have been formalised through the development of the Water Policy that led to the promulgation of legislation dealing with the resource management. The democratization process has shifted the emphasis from supply management to demand management which seeks to conserve the nation's water resources by reducing demand through innovative means. The principal legislation is contained in the Constitution and in the National Water Act (36 of 1998). Principle 7, of the White Paper (DWA, 1997), states that the objective of managing the quantity, quality and reliability of the nation's water resources is to achieve optimum long-term, environmental, social and economic benefits for society from their long-term use. The National Water Policy, founded on the Water Law principles accepted by cabinet in November 1996, commits the national government to carry out its obligations under the NWA in a way which takes into account the interconnected nature of the water cycle - a process on which the sustainability and renewability of the resource depends (*RDM Manual*, 2003).

The water resources of South Africa fulfil a variety of functions and offer a range of services to the nation. They 'quench the nation's thirst, cleanse the waste, regulate the floods, drive agricultural and industrial developments, provide recreational and tourist services, and also form and sustain vital habitats that support subsistence and economic development. Society is dependent on these goods (functions) and services for our daily existence' (Schriener, 1999).

The National Water Act (NWA) of 1998 advocates that one of the ultimate aims of water resource management (WRM) in South Africa is to achieve the sustainable use of water for the equitable benefit of all users. The NWA specifies that the WRM in South Africa will be based on a National Water Resources Strategy (NWRS) that has been drafted and accepted by Cabinet in September 2004. The NWA and NWRS, amongst other things, set the principles for establishing Resource Directed Measures (RDMs) that will conserve and protect the various water resources. These RDMs contained in Chapter 3 of the NWA, will establish the means for:

- Classifying a water resource (ecological management) class
- Setting the (Ecological and Human) Reserve
- Establishing the resource quality objectives (RQOs) to protect the resource

The NWA stipulates that 'The Reserve', both human and ecological, shall be secured so that the use of water does not individually or cumulatively compromise the long term sustainability of the aquatic and associated ecosystems. This is the quantity and quality required to protect aquatic ecosystems in order to secure ecologically sustainable development and use of the relevant water resource (NWA, Ch1, para.1 xviii).

The Human Reserve is currently set by the NWA at 25 ℓ/person/day. In order to establish the Human Reserve, it is necessary to determine the demographics that depend on each water resource. However, the purpose of the Ecological Reserve is to preserve (set) the necessary requirements for maintaining a particular state (class) of a resource so that it can deliver the goods and services the stakeholders require. This means that the Reserve is linked to the state (class) that is perceived to be most beneficial by all the stakeholders. To arrive at this class it is necessary to evaluate the ecological state of the resources under the present conditions and then to determine, using best practice or 'expert opinion/knowledge', the deviation from the reference (or natural) state (ecological class) and to predict the consequences of these changes into the future in order to arrive at the Environmental Management Class in consultation with the main stakeholders (through public participation). Once the Environmental Management Class has been established, it is then necessary to set the resource operating rules that need to be established in order to maintain the goods and services that constitute the sustenance of a water resource for the identified stakeholders.

In the NWA, the protection of water resources requires the Minister to prescribe a system for classifying the water resources. The classification system should establish guidelines and procedures for determining different classes of water resources, including procedures for the determination of the Ecological Reserve. Every water resource must be classified by the Minister before any organ of state or management institution can exercise any powers in terms of the NWA. Hence the allocation of water requires some knowledge of the interrelationships between the various water resources if their ecological function is to be protected. This report will examine the importance of groundwater in the hydrological cycle and its relationship to surface water bodies.

With the present scientific knowledge and understanding of the various aquatic systems, it was necessary to establish the Ecological Reserve for the different specified water resources in an independent manner. The Reserve is presently determined for rivers, estuaries, wetlands and groundwater by separate processes. Little recognition is given to the possible interactions that occur between the various resources. This is a pragmatic approach aimed at overcoming uncertainties in our understanding of the interactions and interdependencies of the various resources. During a recent Reserve determination for the Lake St. Lucia Estuary (Van Niekerk, 2005), the 'experts' recognised the importance of groundwater in sustaining the ecological function of the system but there was no methodology for including it in the formal assessment process.

This project was established to look at the importance of groundwater in relation to other surface water bodies in an effort to understand more about the interactions to support management functions. It was initially perceived that the focus would be on the interaction between groundwater and rivers. However, the importance of groundwater to the other resources is now recognized and needs to be incorporated into the report. The project was also established to enhance capacity, particularly amongst the previously disadvantaged communities, for the management of South African Water Resources. Consequently, the project has relied heavily on student participation and some of the case studies presented in this report have been done by these support students in the Department of Hydrology at the University of Zululand.

INTRODUCTION

To understand the processes and interaction between the different water resources it is necessary to understand their main features, particularly the features that differentiate one resource from another. The title of the report implies that there are easily definable and classifiable resources that interact through specifiable processes. However, the experience of the research team in the coastal region of Zululand coastal plain highlighted the difficulties in differentiating between the two main resource units (namely: surface water and groundwater) in many situations. Consequently, the report attempts to describe the basic units on which to judge the importance of the interactions through a series of case studies.

1.1 Classification of Water Resources

The National Water Act (NWA) recognises the continuum of water resources defined in the hydrological cycle and the interdependence of its elements, where evaporation, clouds and rainfall are linked to underground water, rivers, lakes, wetlands and the sea. It also considers that the basic hydrological unit is the catchment. However, the NWA still finds it necessary to define several distinct legal classes of water resources for the purpose of water conservation and protection management. These have been defined in the NWA (Table 1). Other approaches prefer to use the concept of a water resource as an aquatic ecosystem for the purpose of the RDM (*RDM Manual*, 2003). The *RDM Manual* describes the key component of the RDM as the integrated assessment of the water quality and quantity, the surface water and groundwater, and the rivers, lakes, wetlands and estuaries.

While aquifers feed the rivers, springs, channels and wetlands, they are also fed by these surface resources through many diverse pathways. In many ecological systems, it is not clear what the ecosystem dependency on water resources is likely to be, and hence which classified resource is essential in sustaining the ecological function of the aquatic environment. In order to allocate water in accordance with the requirements of the NWA, it is imperative that the interdependencies are realized and incorporated into the assessments and allocations of each resource.

Table 1: Definitions of specific water resources specified in the NWA of 1998

Resource	Description
Water resource	Includes a watercourse, surface water, estuary, or aquifer
Watercourse	Rivers, springs, natural channels, wetlands, and specifically designated resources
Estuaries	Partially or fully enclosed, bodies of water which are open to the sea permanently or periodically
Aquifers	Defined as geological formations which have structures or textures that hold water or permit appreciable water movement through them
Wetland	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

An examination of the meaning of the main resource categories given in Table 1 indicates that there is considerable variation and much overlap in these definitions. In particular, the Ramsar Classification

System for Wetland Types (Table 2), as approved by Recommendation 4.7 and amended by Resolution VI.5 of the Conference of the Contracting Parties (http://www.ramsar.org/key_guide_list_e.htm)² incorporates the rivers, lakes and estuaries into the term wetland. However, for the purpose of this report, it was decided to limit the investigation of interactions to the broad categories listed in Table 1 which conform to the RDM classification unless there is compelling reason to include specific cases.

Table 2: Ramsar Classification for Wetland Types (Ramsar, 2004)

Marine / Coastal Wetlands
A Permanent shallow marine waters in most cases less than 6 m deep at low tide; includes sea bays and straits B Marine sub-tidal aquatic beds; includes kelp beds, sea-grass beds, tropical marine meadows C Coral reefs D Rocky marine shores; includes rocky offshore islands, sea cliffs E Sand, shingle or pebble shores; includes sand bars, spits and sandy islets; includes dune systems and humid dune slacks F Estuarine waters; permanent water of estuaries and estuarine systems of deltas G Intertidal mud, sand or salt flats H Intertidal marshes; includes salt marshes, salt meadows, saltings, raised salt marshes; includes tidal brackish and freshwater marshes I Intertidal forested wetlands; includes mangrove swamps, nipah swamps and tidal freshwater swamp forests J Coastal brackish/saline lagoons; brackish to saline lagoons with at least one relatively narrow connection to the sea K Coastal freshwater lagoons; includes freshwater delta lagoons Zk(a) Karst and other subterranean hydrological systems, marine/coastal
Inland Wetlands
L Permanent inland deltas M Permanent rivers/streams/creeks; includes waterfalls N Seasonal/intermittent/irregular rivers/streams/creeks O Permanent freshwater lakes (over 8 ha); includes large oxbow lakes P Seasonal/intermittent freshwater lakes (over 8 ha); includes floodplain lakes Q Permanent saline/brackish/alkaline lakes R Seasonal/intermittent saline/brackish/alkaline lakes and flats Sp Permanent saline/brackish/alkaline marshes/pools Ss Seasonal/intermittent saline/brackish/alkaline marshes/pools Tp Permanent freshwater marshes/pools; ponds (below 8 ha), marshes and swamps on inorganic soils; with emergent vegetation water-logged for at least most of the growing season Ts Seasonal/intermittent freshwater marshes/pools on inorganic soils; includes sloughs, potholes, seasonally flooded meadows, sedge marshes U Non-forested peatlands; includes shrub or open bogs, swamps, fens Va Alpine wetlands; includes alpine meadows, temporary waters from snowmelt Vt Tundra wetlands; includes tundra pools, temporary waters from snowmelt W Shrub-dominated wetlands; shrub swamps, shrub-dominated freshwater marshes, shrub carr, alder thicket on inorganic soils Xf Freshwater, tree-dominated wetlands; includes freshwater swamp forests, seasonally flooded forests, wooded swamps on inorganic soils Xp Forested peat lands; peat-swamp forests Y Freshwater springs; oases Zg Geothermal wetlands Zk(b) Karst and other subterranean hydrological systems, inland
Note: 'floodplain' is a broad term used to refer to one or more wetland types, which may include examples from the R, Ss, Ts, W, Xf, Xp, or other wetland types. Some examples of floodplain wetlands are seasonally inundated grasslands (including natural wet meadows), shrublands, woodlands and forests. Floodplain wetlands are not listed as a specific wetland type herein.

² Reprinted from the Strategic Framework and guidelines for the future development of the List of Wetlands of International Importance, Appendix A. http://www.ramsar.org/key_guide_list_e.htm.

Human-Made Wetlands

1. Aquaculture (e.g. fish/shrimp) ponds
2. Ponds, includes farm ponds, stock ponds, small tanks (generally below 8 ha).
3. Irrigated land; includes irrigation channels and rice fields
4. Seasonally flooded agricultural land (including intensively managed or grazed wet meadow or pasture)
5. Salt exploitation sites, salt pans, salines, etc.
6. Water storage areas; reservoirs/barrages/dams/impoundments (generally over 8 ha)
7. Excavations; gravel/brick/clay pits; borrow pits, mining pools
8. Wastewater treatment areas; sewage farms, settling ponds, oxidation basins, etc.
9. Canals and drainage channels, ditches
10. Zk(c) Karst and other subterranean hydrological systems, human-made

1.2 Legislated Interdependencies

The National Water Policy recognizes that the water resources cannot be managed in isolation from other natural resources. 'Water availability is governed by the water cycle, in which rain falls from clouds, flows over the land or sinks through the ground, where it may be stored as groundwater in underground aquifers, and finally flows through rivers, lakes and dams towards the sea. Evaporation from surface waters and the transpiration of plants and trees feeds the clouds and the cycle continues'.

The National Water Policy goes on to state that 'it is in the public interest to ensure that the development and use of all surface, ground and unconventional water resources are undertaken in a way that is sensitive to the environment.' It is also in the public interest to make sure that the development of water resources, which is often the key factor in enabling economic development to take place, is carefully planned so as to promote rather than restrict such development. An example of the consequences of allocating a forestry licence for abstraction of water by *Eucalyptus* plantations (streamflow reduction activity) that led to the complete depletion of river flow which decimated the hydrological function of the Siyaya Catchment is presented as a case study in **Part 2** of this report.

The legal framework within which water allocation will be made has been described in the NWA. The approach to the implementation of this framework will have to take account of the specific circumstances surrounding the different hydrological systems and the impact of intervention which is likely to vary considerably over the extent of the different areas of the country. The framework must look at the various resources and identify the dominant processes that lead to the specification of control measures. This is a very extensive and complex system which requires the conceptual understanding by hydrological specialists working in a team approach. Consequently, this report focuses on a very specific part of the system involving the interaction between the groundwater and surface water resources while recognising the scope of the resource. This report strives to provide a conceptual understanding of the dominant features of all these interactions.

1.3 Groundwater Management Strategy

Groundwater has traditionally been a private resource with little development or Government interest because it contributed only about 15% of the total bulk water supply in South Africa. In the early years of democracy, groundwater was regarded as a cheap source of water, requiring little or no planning or management (Braune, 2000). Because of the highly distributed nature of water demand in South Africa, particularly in the rural area, regional bulk water supply schemes from surface water resources are often

uneconomical and there is a great reliance being placed on groundwater. This presents groundwater with a completely new role as the most feasible option for rural water supply (Braune, 2000).

Surface water resources have traditionally been well defined and managed. Methods for the Reserve determination of the surface water system are developing rapidly. However, the groundwater Reserve determination methodology (GWRDM) is being hindered by a lack of understanding of the role of groundwater in the sustainability of ecological systems. This is generally a feature of the interaction between surface water and groundwater systems.

Groundwater contributes to streamflow in most physiographic and climatic settings. The amount of water that groundwater contributes to streams has traditionally been estimated by analysing streamflow hydrographs that determine a component that is termed baseflow (Winter *et al.*, 1998). There are numerous methods for analysing streamflow hydrographs (Smakhtin, 2001). Winter *et al.* (1998) used one of these methods to determine a conservative estimate of the groundwater contribution to the streamflow in 24 regions of the contiguous United States. In these regions, Winter *et al.* (1998) found that the groundwater contribution ranged between 14 to 90% of the total runoff. The median was 55%. Consequently, the withdrawal of groundwater from the catchment areas in these regions can have a significant effect on the movement of water between the aquifers and surface water resources. Withdrawing water from shallow aquifers near surface water bodies can diminish the available surface water supply by capturing some of the groundwater flow that would otherwise have discharged into the surface water resource or by inducing flow from the surface water body into the surrounding aquifer.

Snyman (2005), summarized in Case Study 1, has developed a method for predicting the groundwater contribution to river runoff using the unique spatial characteristics of the catchment to derive the flow component hydrograph that make up stormflow. The model is calibrated against observed flow hydrograph using traditional flow-separation methods.

1.4 Aims and Objectives

Conceptualisation of this project has been difficult and progression has been slow. The initial proposal was accepted by the WRC with a proviso by the project manager to revise the title and aims. These were changed by the steering committee to those described below. These changes were accepted by the Hydrological Research Unit, University of Zululand, at the initial steering committee meeting but the consequences of the change were not fully appreciated by the project team at the time. The enormity of the project aims became apparent when staff left and capacity training was necessary to continue to meet the deadlines set in terms of the revised aims. This has had a huge impact on the project achievements.

AIMS:

To develop a conceptual understanding of typical groundwater/surface water processes in South Africa in general, and in shallow primary aquifers in particular.

To critically review (assess) methodologies (and models) to quantify the groundwater /surface water interaction.

To establish compatible methods to quantify time series of surface water and groundwater flow rates for comparative analysis.

To develop monitoring systems to contribute to the measurement and interpretation of the groundwater / surface water interface.

CHAPTER 2: THE CONCEPTUALIZATION OF WATER BODY INTERACTION

INTRODUCTION

The conceptualisation of the groundwater/surface water interaction can be approached in different ways that focus on specific features. A description of the interaction can be based on the hydrological processes, the individual resources, the interdependencies of specific systems, or the basis of regulations and controls. The processes driving the hydrological system that link specific resources which are mentioned in the NWA have been described earlier in the report. However, the interdependencies of many water resources and other components of the system (terrestrial, ecosystems, etc.) are important but not specifically mentioned in the NWA. The general approach for setting the Ecological Reserve has largely been based on hydrological responses that regulate the resources while sustaining the ecological function of the system. Consequently, this report has opted to focus on the hydrological processes that are principally responsible for sustaining the surface water and groundwater resources but it is still necessary to define the different resources. The NWA recognises the continuum of water resources defined in the hydrological cycle. It qualifies some of the processes such as evaporation, clouds, rainfall, runoff, etc. that link the various resources. This section will describe the main resources and the processes that create the necessary linkage between the surface and sub-surface (aquatic) systems (Table 3).

Table 3: Principal hydrological processes

Hydrological processes	Common unit of flux
Evaporation and transpiration	(mm/d)
Precipitation (rainfall)	(mm/d)
Groundwater recharge and discharge	(m³/d or mm/d)
River runoff and runoff	(m ³ /d or mm/d)
Lake recharge and discharge (seepage)	(m ³ /d or mm/d)
Wetland recharge and discharge (seepage)	(m ³ /d or mm/d)
Estuarine fluxes (river and groundwater seepage)	(m ³ /d or mm/d)

The main processes that are directly or indirectly involved in the surface water to groundwater interaction are shown schematically in Fig.1. The magnitude and importance of the interactions between groundwater and other 'surface' water systems vary considerably in both scale and locality. Deep confined aquifers are unlikely to make a major contribution to groundwater discharge (surface water recharge) while small localised groundwater seeps/springs may be the only source of freshwater for human consumption and ecological sustainability in a local setting. Consequently, it is necessary to understand the linkages and interactions in terms of both the spatial and temporal scales when

describing the various processes in detail. However, the first step is to establish a common and unifying description of the resources.

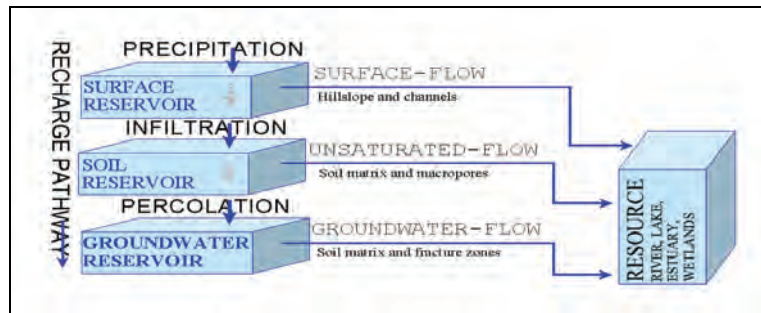


Figure 1: Conceptual Model

2.1 Water Resource Description

The initial stage in the conceptualisation of the interaction between surface water and groundwater systems is to define the difference between the resource categories. The NWA introduces 5 basic water resources, namely underground water, rivers, lakes, wetlands and the sea. It may be argued that the rivers, lakes, wetlands and the sea all constitute surface water resources that are distinct from the underground water. The difficulty with this concept becomes apparent when an attempt is made to

determine when underground water, for example in wetlands, can be classified according to whether the water is visibly above or remains below the ground surface. According to this definition, one class of wetlands would also be classified as groundwater! Consequently, it is necessary in this study to define the difference between the two 'surface and ground' water systems in order to define their interactions. However, the principal need is to support the classification and allocation process as defined in the NWA.

Box 1 : Definition of groundwater references

Bowen (1986): Water in the zone of saturation.

Parsons (1995): All subsurface water occupying voids within a geological stratum.

Driscoll (1995): Underground water that can be removed by wells. All other water in the ground is termed subsurface water and is not available for man's use directly.

Fetter (1994): Water stored in the zone of saturation.

Todd (1980): Water occupying all the voids within a geologic stratum. This saturated zone is to be distinguished from an unsaturated, or aeration, zone where voids are filled with water and air.

Vegter (1995): Water in the zone of saturation. It flows into boreholes/wells or debouches as springs.

Ward and Robinson (1989): Subsurface water in soils and rocks that are fully saturated.

[Synonym: underground or subterranean water]

2.2 Groundwater - The Main Distinguishing Features

Unlike surface water, groundwater has been defined in many different ways by many different practitioners depending on the specific application (**Box 1**). Some of the definitions of the various resources (**Box 1**) that have been cited in the literature and the circumstances in which they were used have created confusion when the issues of surface water and groundwater interactions are discussed. Recharge to groundwater is often a discharge from surface water and vice versa. It is also interesting to note that groundwater is generally written as a single word unlike

surface water. This suggests that the term 'ground water' refers to all the water in the ground, whereas 'groundwater' is reserved for a specific component of the groundwater. For the purposes of clarity, the term subterranean water is used to indicate all the water in the ground instead of the term 'ground water'.

Traditionally in South Africa, surface water and groundwater have been treated as separate entities and the practitioners have, over time, established terminologies that have different interpretations. Consequently, it is considered necessary to establish a common terminology for the processes and features under review in this report. Specific terms used in this report are defined below in Table 4. The conceptualisation and quantification of the importance of 'groundwater' and its interaction with 'surface water' are dependent on the definition of these resources.

Table 4: Specific terms relating to groundwater

Recharge	Flow INTO the aquifer
Discharge	Flow OUT of the Aquifer
Storage	Specific storage

Driscoll (1995) has subdivided subterranean water (which he calls sub-surface water) into several components shown in Fig. 2.

The first division is on the basis of the water table which represents the top of the zone of saturation in a porous medium. This is a well-recognized hydraulic interface in hydrology that separates the (unsaturated) vadose zone from the saturated phreatic zone. However, Luckner and Schestakow (1991) consider this to be a fictitious interface in geochemical terms. Alternatively, Todd (1980) and Boulding (1995) refer to the unsaturated section above the water table as the zone of aeration.

At the macro scale, the water-table interface is usually measurable in boreholes and wells and will continue to be the basis of subterranean water division in most situation analyses. However, at the local or micro-scale in the unsaturated zone near the water-table interface, capillary rise will be induced by the porous media to raise the level of saturation above the freely draining zone in this section of the unsaturated zone. The saturation level in the capillary zone will decrease 'linearly' from saturation at the water table (atmospheric pressure) to field capacity at some vertical elevation above the water table. This capillary zone, within the vadose zone, ranges in depth from millimetres to metres, depending on the nature of the porous media (Boulding, 1995; Todd, 1980) and may be an important consideration when the capillary zone is near the topographical surface. For situations with thick layers of fine-grained material that supports significant capillary rise, such as clay and peat, there may be considerable connectivity between the saturated zone and the topographical surface thereby providing ready access to water for use by surface features. However, it is very difficult to differentiate between the capillary effect and the natural extent of saturation in these sub-surface features, particularly in transient systems.

The surface water level in boreholes and piezometers is equivalent to the elevation of the saturated zone and not the capillary rise. It is believed that the capillary zone may play an important role in sustaining evapotranspiration (Lubczynski, 2009) under extremely dry soil conditions. Most terrestrial

plant roots need to breathe and consequently, the capillary zone provides both air and water which is constantly replenished, by the air from above and the water below due to capillary action from the phreatic zone.

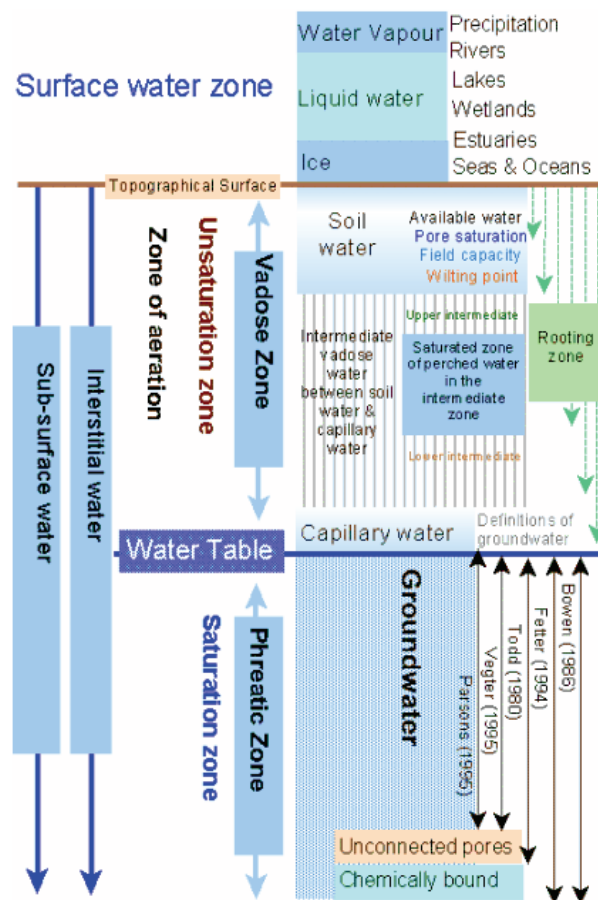


Figure 2: Definitions of subsurface water components of the water resources derived from various sources

2.2.1 The Vadose Zone

The vadose zone (or zone of aeration) above the water table, can be further subdivided into three zones on the basis of their position in the vertical profile (Fig.3). These include the water in the capillary zone (capillary water), the water held in the soil profile (soil moisture), and the intermediate zone (intermediate vadose water) that connects the capillary water to the soil moisture.

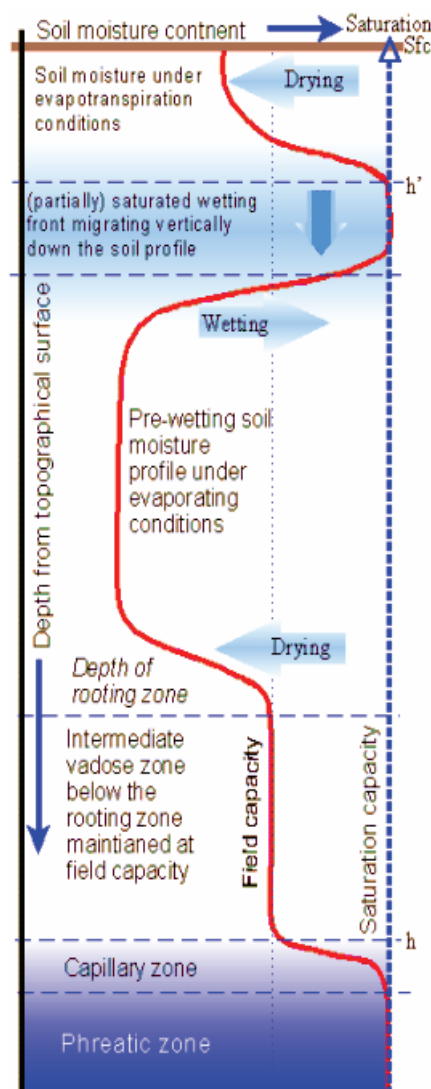


Figure 3: Typical vadose moisture profile

The soil moisture is an important component of recharge and discharge between the subterranean water resources and the surface water resources. The intermediate zone serves primarily as an unsaturated region connecting the soil moisture zone and capillary zone through which water moving vertically downward must pass. In this zone, some water is held by capillary tension and hydroscopic forces (Todd, 1980). This zone may or may not be beyond the rooting profile and hence (un)available for evapotranspiration. Consequently, the changes in moisture content would be transitory as excess soil water (water held above the field capacity of the soil) would percolate through the intermediate zone into the groundwater. An example of this was found in the high coastal dunes (>100 m a.m.s.l.) along the St. Lucia coastline where the water table was at an elevation of about 10 m a.m.s.l. The maximum rooting depth was estimated to be <20 m, leaving an intermediate zone of about 70 m.

The thickness of the three vadose sub-zones is dependent on several factors that include the rainfall, climate, vegetation (rooting zone) and soil profile hydraulic characteristics. An example of a deep intermediate zone can be found in the very high coastal dunes along the Maputaland coast near Cape Vidal. Even with an average annual rainfall of over 1.20 m, the water table can be as much as 100 m below the topographical surface and the capillary zone is likely to be only millimetres thick. In these high

dunes, this intermediate zone is likely to be at field capacity most of the time as the soil is frequently recharged and the rooting zone of vegetation is unlikely to extend deep into the profile. However, the rooting depth of many of the dune forest species is unknown and some trees may be harvesting the water in the intermediate zone.

While intermediate vadose water is conceived to flow vertically downwards while draining under gravity, the stratification and lineation of sub-surface layers could cause considerable lateral movement of percolating water. This could divert flow laterally to surface water resources before entering the phreatic zone through the water table. The concentration of lateral flow along the lineaments in a sloping stratigraphy due to cross bedding planes could lead to the emergence of surface springs. Hughes (2000) has indicated the emergence of water from unsaturated profiles in the Eastern Cape Region that could be attributed to this process. The authors have witnessed water gushing from macropores in the Ntuze catchment of Zululand long after rainfall (Plate 1). This lateral flow is discussed in more detail in later sections of this report.



Plate 1: Photograph of macropore flow in the Ntuze catchment

2.2.2 The Phreatic Zone

The phreatic zone represents all the water below the water table. This zone also contains the water held in unconnected pores and the water held in chemical combination with the rocks. However, most of the water in unconnected pores and bounded in the rock matrix is unavailable for abstraction and consequently is of limited concern for this report.

This classification of subterranean water into zones is a convenient approach for assessing specific hydrological systems but is not used consistently for the definition of groundwater (see some definitions in **Box 1**). It also ignores the transitory nature of the (partial) saturation zones (Fig. 3). The zonation of subterranean water becomes more complex when significant recharge events cause transitory zones of saturation (at level *h* in Fig. 3) that move down the soil profile and through the intermediate vadose and capillary zones into the saturation zone. There are also numerous situations where hydraulic barriers retard (slowdown) or stop the vertical movement of water to cause localized (perched) saturated sections in the vadose zone. These zones are distinguishable from the phreatic zone by an unsaturated section between the perched and regional saturation zones (level *h* in Fig. 3). Most of the definitions of

groundwater cited in **Box 1**, by implication, include the perched zone of saturation and the transitory wetting front. However, the transitory wetting front in the vadose zone will generally move vertically downward toward the water table and may not play a significant role in surface water dynamics but will become available for abstraction from the groundwater if it moves into the phreatic zone. Where hydraulic barriers or preferential pathways exist, this migrating front may move laterally and could intersect topographical features. If this transitory saturation zone is persistent (perched), it can be exploitable and consequently it can be classified as a separate (ground/surface) water resource available for abstraction. Also, perched water tables frequently form wetlands (Plate 2) and consequently, they often play a very prominent role in surface water systems and cannot be ignored in a study of surface water/groundwater systems. Because this perched water may form components of surface water features such as wetlands and springs it must be considered an important part of groundwater/surface water interactions.

For the purpose of groundwater allocation it is necessary to define the temporary scales that affect the nature of the saturation zones. At the scale of days, it may be prudent to define all non-transitory, subsurface saturated water zones for groundwater allocation if the system has sufficient yield for exploitation. This includes all the exploitable water in the phreatic zone and those sections of the regional vadose zone where local perching occurs on a regular or persistent (>> days) basis. According to this definition, there will be zones that have sufficient resources to be extractable for domestic use (as happens frequently in rural communities) and there will also be zones where the potential yield is too small for abstraction. Hence, this limited definition will exclude part of the baseflow in rivers, the low-yielding system that sustains many terrestrial and aquatic ecosystems described in case studies later in the report. Plate 2 shows a perched pan that has been used for water supply on a regular basis by local inhabitants for many years.

Groundwater forms a major component of the available freshwater supply in South Africa. However, its availability, sustainability and vulnerability may vary substantially because of its location and form. This has led to numerous further descriptions or classifications of groundwater. These include terms such as deep and shallow, confined and unconfined, renewable and non-renewable, old and new, fresh and saline. Todd (1980) describes the origin of many forms of groundwater that indicate their transitory nature (or lack of) and hence their level of interaction with the hydrosphere/atmosphere. He describes those groundwater components that are bound by the geology for a considerable period of time (connate water, magmatic water, metamorphic water, plutonic water, volcanic water and juvenile water). These categories of groundwater, together with water that is in more recent contact with the atmosphere, are described in detail in Todd (1980).



Plate 2: Photograph of a perched wetland used for domestic needs

2.3 Groundwater - Time Scale and Composition (Age) of Groundwater

The classification of water can also be based on the position and average state of the water in the ground. The transitory nature of (ground) water can be analysed in terms of its residence time in various reservoirs (zones) and the characteristic time scales of important processes (Fig.4).

Subterranean water has characteristic motion/flow rates/velocities that vary from km/d to $\mu\text{m/d}$. This translates to characteristic transit (residence) times of hours to millennium (Fig.4). These variations in transit/residence times reflect the variations in flow paths and the subsequent impact on the water quality. Hence, groundwater interaction can be analysed for response times that are similar or much greater than the surface water features. Several case studies are presented that look at the hourly fluctuations in groundwater and corresponding variation in the surface water features.

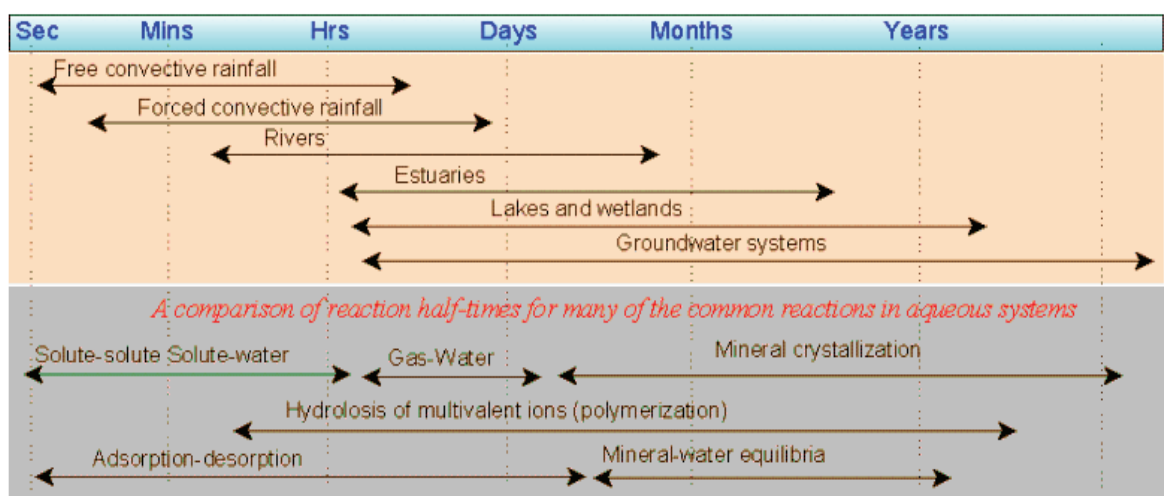


Figure 4: Characteristic response time scales of hydrological processes

2.4 Groundwater Quality and its Use in Flow-Path Analysis

Environmental isotopes of hydrogen and oxygen have been used as natural tracers in catchment studies (Sklash, 1990 quoted in Beven, 2001). Since hydrogen and oxygen are part of the water molecule they will be part of the natural flow paths in the catchment. Differences in rainfall and runoff concentrations of these isotopes have been used to estimate the source of water during runoff processes in the hydrograph (Sklash and Farvolden, 1979). Similar studies are being undertaken to investigate the groundwater processes in St. Lucia (Taylor *et al.*, 2006).

Kelbe and Germishuys (1999b) have used the variation in continuous observations of electrical conductivity (EC) to differentiate between the components of the runoff hydrograph for high-intensity short-duration storms in conjunction with conventional hydrograph analysis (Fig. 5). The initial drop in EC was associated with the rapid rise in discharge associated with quickflow (new water) that had an EC similar to rainfall. When the quickflow had dropped to almost half its maximum value the EC started to rise rapidly with the mixing process until the throughflow component had declined appreciably. At this stage the discharge was predominantly baseflow with similar EC to the pre-storm conditions (Fig. 5).

2.5 Aquifer Boundaries

Aquifers are defined as physical features that have definable boundaries containing the bulk of the groundwater. Consequently, all aquifers can be described on the basis of their physical boundaries and internal properties. These boundaries define the limit of the aquifer and the region or extent of their influence on other resources. In terms of the water balance, the storage of water in the aquifer is constrained by the inflow (recharge) and outflow (discharge) through these boundaries.

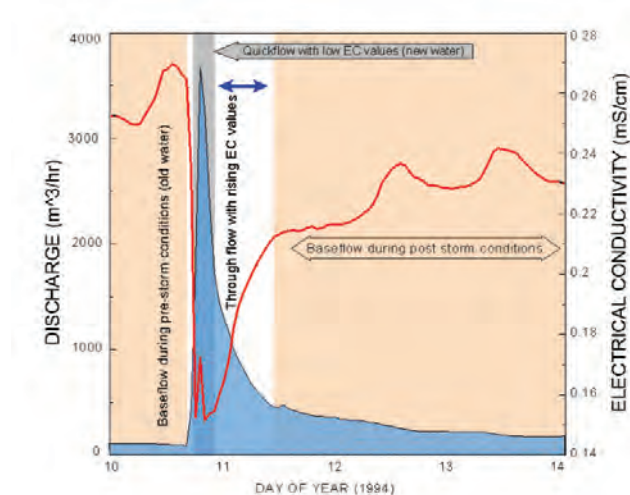


Figure 5: Typical discharge and water quality responses to storm events in the Ntuzi catchments used in identifying flow processes

Consequently it is important to define the boundaries of the aquifers. Methods for defining the boundaries are described in subsequent sections.

While the processes of recharge and discharge operate through portions of these external boundaries, there are processes that operate through internal boundaries that can induce inflow or outflow through localised sections of the aquifer. These internal and external boundaries are conceptualised in Fig. 6.

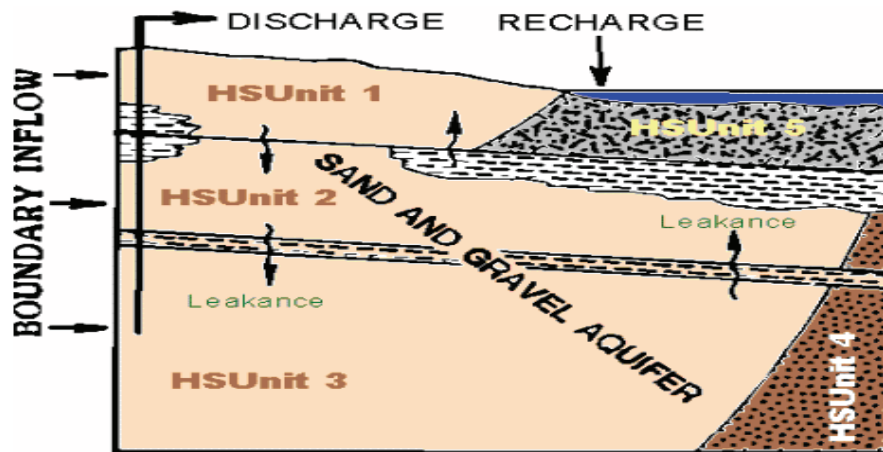


Figure 6: Conceptualized boundaries of an aquifer (adapted from Anderson and Woessner, 1992)

2.5.1 Definition of Aquifers

Aquifers are defined in the NWA 'as geological formation which have structure or textures that hold water or permit appreciable water movement through them'. Traditionally they have been classified as either a confined or an unconfined system depending on the presence or absence of a water table. These are relative terms for two classes which form the extremes of a continuum. There are many (intermediate) aquifers that fall in between these two classes which are generally referred to as leaky aquifers. The three basic types of aquifers are described below.

2.5.1.1 Unconfined (Primary) Aquifers

Unconfined aquifers are ones in which the water table varies in undulating form and slope, depending on areas of recharge and discharge and hydraulic properties of the porous medium. There is no constraint on the flux through the upper surface. Within the aquifer, there are often areas with low permeable strata where concentrations of soil moisture lead to perched water tables (Fig. 7). Most of the alluvial deposits along flood plains of large river systems comprise unconfined aquifers which are closely linked to the recharge and discharge from the rivers. The largest primary aquifer in South Africa lies along the NE coastal plain of KwaZulu-Natal (Midgley *et al.*, 1994).

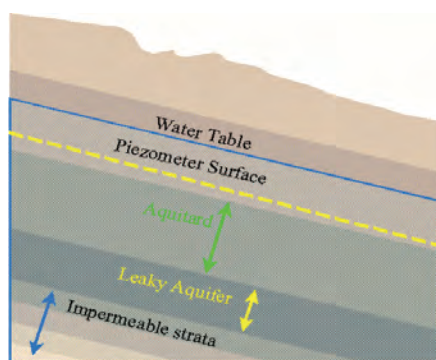


Figure 7: Leaky aquifer

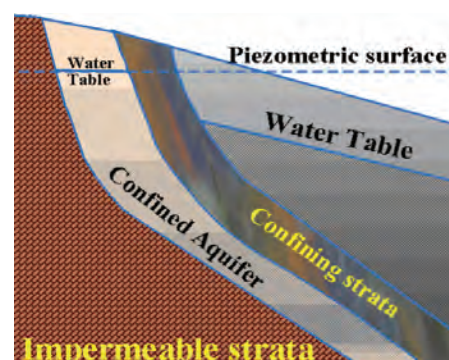


Figure 8: Confined aquifer

2.5.1.2 Leaky Aquifers

Leaky aquifers are those sections of the geological formations where there is a permeable stratum that is overlain or underlain by a semi-pervious aquitard or semi-confining layer that imposes limitations on the flux through the upper or lower surface. These leaky aquifers generally have different piezometric surfaces for the different layers (Fig. 7). The unconsolidated sediments in the Zululand Coastal plain of the Port Durnford Formation that overly the Conglomerates of the Miocene deposits are an example of these types of aquifers (Kelbe and Germishuysen, 2000b). This can cause the aquifer to have higher or lower piezometric surfaces to the overlying aquifers.

2.5.1.3 Confined Aquifers

Confined aquifers also known as artesian or pressure aquifers, occur where groundwater is confined between impermeable (confining) layers (Fig. 8). This causes the aquifer to be under pressures greater than atmospheric in sections of the strata. These aquifers are recharged in unconfined regions where there is scope for direct interaction with the surface. The recharge will induce a change in storage unless there is a compensating discharge at some boundary. The only surface water/groundwater interaction occurs in the recharge zone or at specific interface boundaries where the confined aquifer intersects the surface.

It is important to recognize that these descriptions of aquifers are a gross generalization where, in reality, the systems are more complex. Certain boundaries (e.g. faults) provide well-defined horizontal and vertical features that aid in the classification of aquifers. However, most delineation of aquifers is dependent on the scale of analysis and the regional and local geology derived from a few exploratory observation points that require interpolation or extrapolation in conjunction with expert opinion to define these external boundaries.

2.5.2 Groundwater Storage

The assessment of temporal changes in the mass balance of an aquifer requires an understanding of the storage function of the system. The amount of water that can be stored in these systems depends on the type of aquifer and its hydraulic properties. In unconfined aquifers, the ones of principal interest to this project, the storage volume within the aquifer is directly related to the porosity (i.e. those portions of the porous material are not occupied by solid material). The porosity is made up of spaces known as voids, interstices, pores and pore spaces. However, in confined aquifers, the storage volume, also contained in the interstices of the strata are generally under pressure from the overlying burden. The confined aquifer can be described in terms of its storativity (storage coefficient) which is the volume of water that an aquifer releases into storage per unit surface area per unit change in piezometric head.

The definition of storage and the parameters necessary to define it depend on the analytical and numerical methods used for analysis. These are discussed in the different sections of the report.

2.5.3 Groundwater Recharge

In order to develop a conceptual understanding of the interaction between the water resources traditionally classified as surface water and the groundwater, it is necessary to establish some terms of reference. In this resume, the issue of scales of interaction becomes important and it is necessary to

differentiate the interaction at different time and space scales. Because of the need to implement the NWA, the principal scale of focus will be on the macroscopic features that provide means for classification and allocation of the resources.

The storage capacity of an aquifer is defined by the aquifer boundaries and the hydraulic properties of the porous material that make up the aquifer. The mass balance of an aquifer unit equates the change in storage over time ($\partial s / \partial t$) to the difference between what comes in (recharge) and what goes out (discharge) as depicted in Fig. 9.

$$\frac{\partial s}{\partial t} = IN - OUT = \text{Recharge} - \text{Discharge}$$

Under average hydro-climatic conditions, it is generally assumed that the system is in a state of equilibrium so that there is no change in storage ($\partial s / \partial t = 0$) and $IN = OUT$. In transitory conditions the change in storage will determine the available water for abstraction. Consequently, the amount of recharge that is available for use (allocation) is dependent on the rate of discharge. Hence, the determination of the discharge boundaries becomes a very important component and needs to be defined and specified accordingly.

Recharge is the complex process where water enters the aquifer through various pathways to form part of the groundwater storage. It has been defined in many different ways in the literature. Some of the definitions applied by researchers are described in Table 5.

Table 5: Definitions of recharge given in the literature

Bowen (1986)	The difference between precipitation and runoff plus other losses, i.e. that part of precipitation which resides as groundwater, surface storage and soil moisture.
Rushton and Ward (1979)	Groundwater recharge is that amount of surface water which reaches the permanent water table either by direct contact in the riparian zone or by downward percolation through the overlying zone of aeration. It is this quantity which may in the long term be available for abstraction and which is therefore of prime importance in the assessment of any groundwater resource.
Besbes and De Marsily (1984)	Recharge is defined as the flux arriving at the water table, i.e. after percolation through the whole unsaturated zone. In this definition, we assume that the transfer of the "effective rainfall" through the unsaturated zone is conservative. Average infiltration and average recharge are thus identical over a long period ... Implicit in this definition is the assumption that there is no upward migration from the water table to the soil surface.... This assumption is probably valid in temperate or dry climates, where the water table is deep and the delay between infiltration and recharge therefore significant.
Vegter (1995)	The processes involved in the absorption and addition of water to the zone of saturation. Synonym: Replenishment
Anderson and Woessner (1992)	The volume of infiltrated water that crosses the water table and becomes part of the groundwater flow system. (Discharge refers to groundwater that moves upwards across the water table and discharges directly to the surface or unsaturated zone)
Sharma (1998)	The addition of water to an aquifer usually resulting from vertical percolation of water through the unsaturated zone.

The definition adopted in this study is based on the need to describe the surface water/groundwater interaction shown in Fig. 10). In this context, the recharge from the surface (upper boundary) can be defined as that flux of water which percolates through the upper aquifer boundary and into the regional groundwater body and that which flows under a hydraulic gradient from other parts of the aquifer (D).

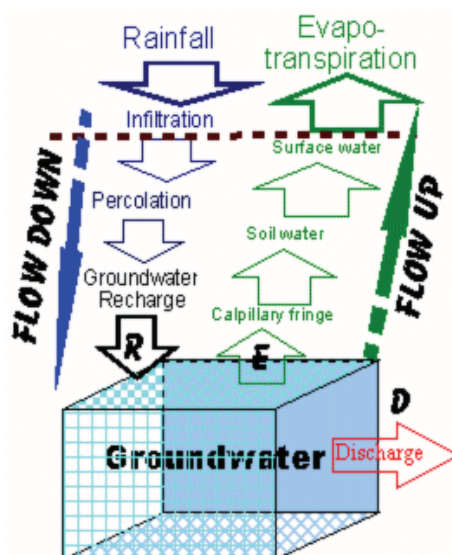


Figure 9: Schematic diagram depicting the various processes that lead to groundwater recharge and discharge

In terms of the continuity equation, the flux (m^3/d) is given by the gross recharge (R_g):

$$R_g = E + D + dS/dt.$$

To derive a relationship between recharge and rainfall, it is necessary to determine the surface and soil moisture fluxes (Fig.10). The evaporation (E) from the unsaturated zone involves soil moisture that is replenished from above and below. In this highly complex system it has become pragmatic to define several terms that are used under certain conditions.

$$R_G = \text{gross recharge (defined above)}$$

$$R_N = \text{net recharge} = \text{Gross recharge} - E - D$$

Bredenkamp *et al.* (1995) have examined most of the large-scale recharge estimation methods used in South Africa. The bulk of these methods involve estimation of the Net recharge (RN) which is applicable at the larger scales of application (regional long term studies). However, in many local, short term estimations, it is necessary to determine the gross recharge (RG). This becomes extremely important when evaluating the impact of evapotranspiration on groundwater system. It is not possible to estimate/model evaporation losses (discharge) from the groundwater if it has already been (partially) extracted from the recharge.

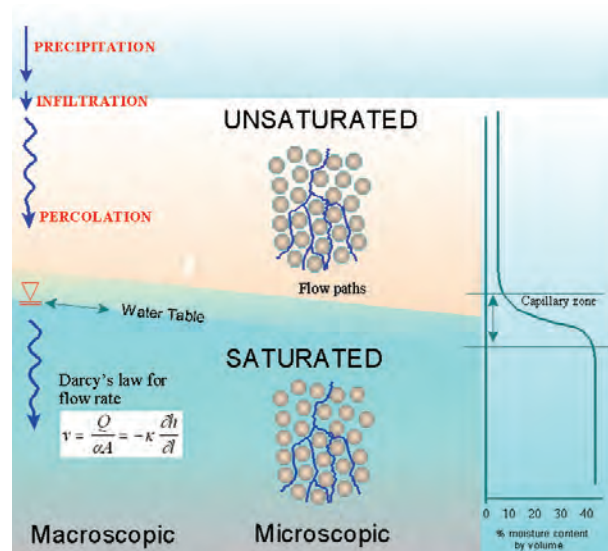


Figure 10: Uniform recharge in a homogeneous system

2.5.4 Variation in Recharge Patterns

In a similar approach to that used by Winter *et al.* (1998), the physical characteristics of the hydrological systems were broadly classified into four conceptual landscapes depicting extremes of hydrogeological features (unconsolidated sediments and hard fractured rock systems). These were used to describe the variations in vertical and horizontal flow patterns in subterranean systems through the upper aquifer boundary. For homogeneous porous media, lateral flow occurs above the water table in the capillary zone (Todd, 1980). It is generally accepted that no lateral flow occurs in the unsaturated zone at a macroscopic scale above the capillary fringe. However, lateral movement of water does occur in the unsaturated zone at a microscopic scale which becomes extremely important in chemical dispersion impacts. The flow paths will generally follow the line of least resistance through the matrix of soil particles of variable shape and size. The most probable pathway is in the vertical plane with a decreasing probability for lateral flow rates. Consequently, the dispersion is a function of matrix properties.

2.5.4.1 Vertical Matrix Flow System in a Homogeneous, Uniform, Porous Medium

The vertical flow process in a homogeneous soil medium includes capillary connection between unsaturated and saturated zones (Fig.11). No lateral flow occurs at the macroscopic scale in the unsaturated zone.

Saturated flow obeys Darcy's Law. This type of system does display spatial variability in flow (fingering) at the local and macroscopic scales. For homogeneous porous media, lateral flow occurs above the water table in the capillary zone (Todd, 1980). It is generally accepted that no lateral flow occurs in the unsaturated zone at a macroscopic scale above the capillary fringe.

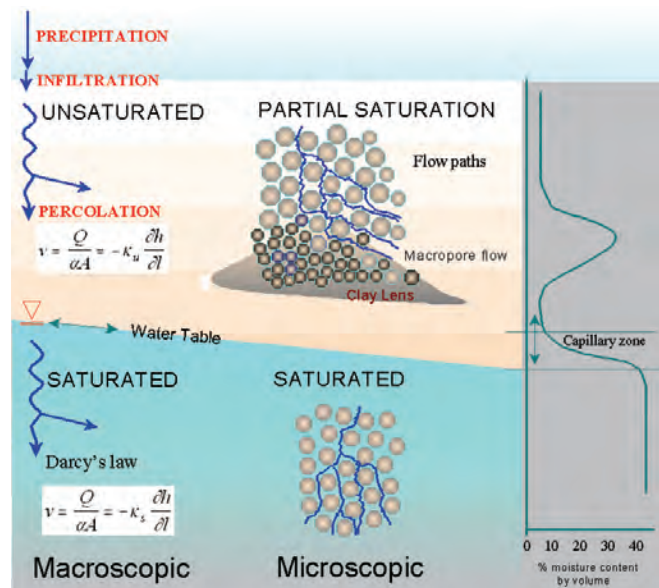


Figure 11: Non-uniform recharge in a heterogeneous system

The flow paths will generally follow the line of least resistance through the matrix of soil particles of variable shape and size. This class is unlikely to have a pronounced or prolonged 'interflow' component in the stream discharge.

2.5.4.2 Vertical Matrix and Lateral Boundary Flow System in a Heterogeneous, Non-uniform, Porous Medium

This class (Fig. 12) has lateral flow that distinguishes it from the first, purely vertical, flow system. These heterogeneous systems display large variability in hydraulic features (porosity and permeability). Dual porosity and permeability at the meso-scale(local) lead to variable flow patterns that produce lateral flow conditions. This class of recharge could lead to lateral discharge which is probably closely associated with the 'interflow' component in hydrograph analyses. Clay lenses and macropores contribute to the variability.

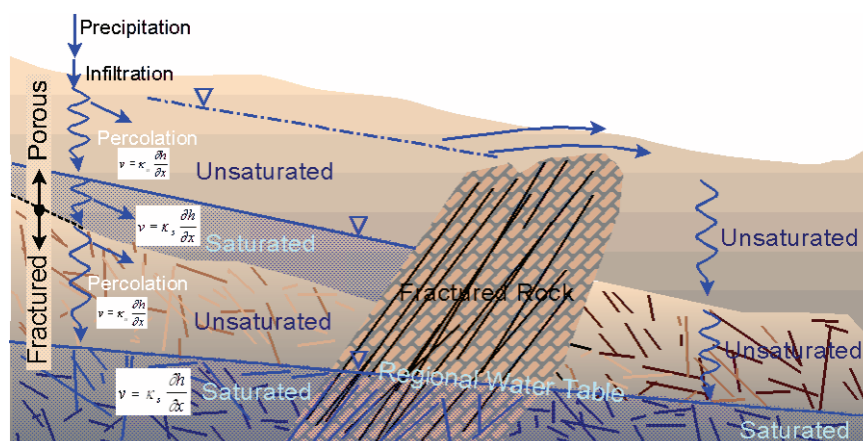


Figure 12: Complex recharge in a conglomerate system with porous and fractured components

Lateral flow in a porous medium will be likely to occur where it shows a large degree of heterogeneity in the hydraulic characteristics. This occurs from inclusion of sand and clay lenses with different

permeabilities (K), bedding planes with aerial hydraulic properties (K) and with the extent of macropore development. These heterogeneous hydraulic features produce variable zones of moisture content that may lead to localized zones of saturation which induce significant macroscopic flow rates in the lateral direction. At the microscopic scale, the dispersion still occurs but is enhanced by changes in hydraulic characteristics due to variable soil properties (clay lenses) and bedding planes that have different hydraulic conductivity. The occurrence of macropores induces preferential pathways that induce rapid lateral flow paths. The flow rate can be described by Darcy's Law but the hydraulic conductivity is a function of the degree of saturation (Todd, 1980) and the extent of macropore development. In these cases the Darcy velocity becomes an empirically derived value.

2.5.4.3 Complex Interaction of Matrix and Fractured Recharge Systems that are Distinguishable from the Fractured and Porous (Matrix) Systems

This class (Fig. 13) has intermediate (perched) recharge zones in the unsaturated strata which are likely to impacts on the interactive mechanisms for recharging the regional underlying aquifer.

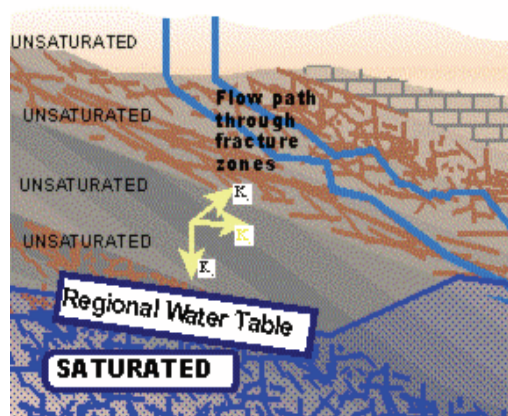


Figure 13: Non-uniform recharge in a fractured rock aquifer

The interactions involve both vertical and lateral flow paths that need to be described. Lorentz (2000) describes a situation where it is necessary to have the perched conditions to support regional groundwater recharge to fracture zones. In these situations, the recharge is not a vertical process but can be derived from lateral flow conditions.

2.5.4.4 Thin Soil Mantle Overlying Fracture Rock Recharge Zone above Regional Groundwater System

This class (Fig. 14) has vertical and lateral flow paths in the unsaturated zone that are directly associated with the hydraulic features (fracture zones) of the strata.

Preferential pathways lead to differential recharge rates for different geological features. In these situations, the recharge zone is generally displaced from the confined aquifer and the recharge process must be conceptualized as a direct flux. In all the cases presented in Figs. 11 to 14 the recharge would continue to increase storage (rising head) until it balances the discharge. Consequently, the specification of the discharge boundaries becomes critical in evaluating the aquifer and its relationship with other aquatic systems.

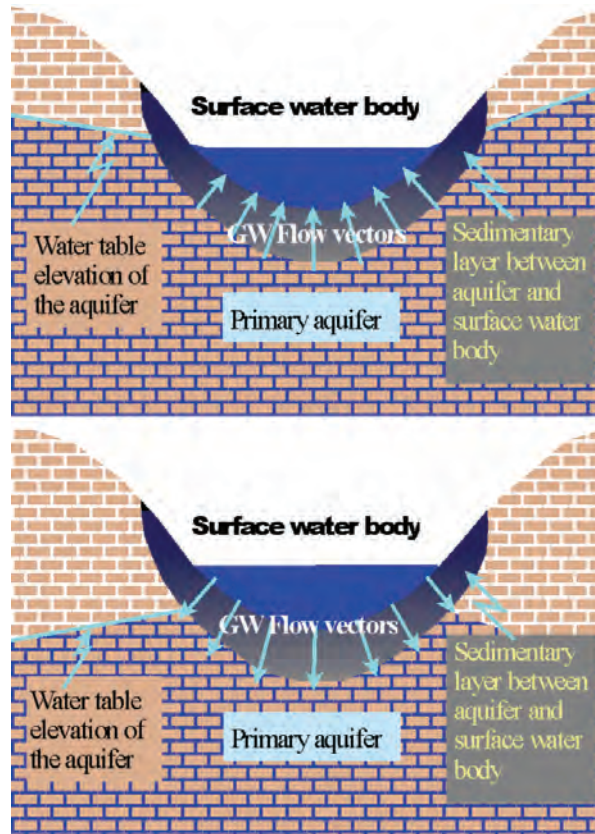


Figure 14: Conceptualisation of the interactive boundary between a primary aquifer and a surface water body (river, lake, wetland or estuary)

2.5.5 Groundwater Recharge Process within the Saturated Zone

The dominant recharge mechanism is down vertical pathways in the unsaturated zone even where there is some lateral flow. In the saturated zone, the lateral flow becomes the important feature which leads to recharge/discharge from the aquifer.

Flow conditions within the saturated zone along the flow path are a function of the hydraulic characteristics of the aquifer together with the hydraulic gradient. The hydraulic characteristics are a function of the geology while the hydraulic gradient is a function of the boundary conditions. The flow rate in an aquifer can be defined by Darcy's Law under mass balance constraints in a Cartesian coordinate system using the following Boussinesq model (from Anderson and Woessner, 1992):

$$\left[\frac{\partial \left(K_{xx} \frac{\partial h}{\partial x} \right)}{\partial x} + \frac{\partial \left(K_{yy} \frac{\partial h}{\partial y} \right)}{\partial y} + \frac{\partial \left(K_{zz} \frac{\partial h}{\partial z} \right)}{\partial z} \right] + W_B = S_i \frac{\partial h}{\partial t}$$

where:

K_{ii} is the hydraulic conductivity along the Cartesian coordinate axes which are assumed to be parallel to the major axes of the hydraulic conductivity (L/T).

h is the piezometric (potentiometric) head

$(L)W_B$ is a volumetric flux per unit volume representing sources and/or sinks at the various boundaries (B) (T-1)

S_i is the specific storage of the hydrostratigraphic unit (L-1)

T is time

Hydraulic conductivity (K) is defined for a medium which transmits a unit volume of groundwater in a unit time through a cross section of unit area (A) measured at right angles to the direction of flow under a unit hydraulic gradient. This equation is valid for laminar flow (usually $NR < 1$) up to a Reynolds number (NR) of 10 (Todd, 1980) which is within the flow conditions of most aquifer.

Unsaturated flow can also be described by Darcy's Law but the application suffers from hysteresis and uncertainty in the derivation of parameters (K). For unsaturated flow, the hydraulic conductivity is a function of the porous material, the water content and the soil moisture tension (negative pressure head). This model of groundwater flow is routinely applied in groundwater studies (Anderson and Woessner, 1992) and has been incorporated in most groundwater models currently in routine use (see subsequent sections). The sources (recharge) and sinks (discharge) are treated as separate independent functions operating through a horizontal boundary (interface zone). The recharge boundary has been described above. The discharge boundaries are described below.

Under steady state (average hydro-climatic) conditions, the change in storage term become negligible and the flow within the aquifer is balanced by the sources and sinks (WB).

2.5.6 Discharge Processes (Physical Boundaries)

Physical boundary conditions fall into one of three categories (Anderson and Woessner, 1992):

- **Specified head** (Dirichlet condition) where the head in the aquifer is set by the boundary condition such as the Oceans
- **Specified flux** (Neumann condition) where the flow rate through the boundary can be defined such as recharge
- **Mixed** (Cauchy condition) where the hydraulic gradient between the aquifer and boundary defines the flow rate through the boundary, such as rivers, lakes, estuaries and wetlands. These are often referred to as head-dependent boundaries.

2.5.6.1 Specified head boundaries

Set the head within the aquifer. A special case is the large surface water bodies that remain fairly constant in relation to the transient state or variability of the aquifer (or the duration of the study creating a 'constant head' boundary). The large ocean masses are generally considered to be constant specified (0 m a.m.s.l.) head boundaries. This boundary can be conceived as an infinite source or sink for flow into and out of the aquifer.

2.5.6.2 Specified flux boundaries

Set the rate of flow into or out of the aquifer. A special case can be described for situations where there is no flow across the boundary (no-flow boundaries). An impermeable surface will create a no-flow boundary. Similarly, the groundwater divide would constitute a no flow boundary. However, this boundary is generally dynamic and a direct response to changes within the aquifer. Direct abstraction or injection through wells and boreholes would constitute a specified flux boundary. In this case the boundary could be located within the main body of the aquifer.

2.5.6.3 Head-dependent boundaries

In most physical systems, the boundary would respond directly to the changes in the aquifer conditions. The change in one resource would create a compensating change in the other resource. In these cases, the flux to/from the groundwater systems will be determined entirely by the boundary conditions at the interface between the aquifer and the other aquatic systems. Flow within the groundwater system is determined by the hydraulic characteristics of the aquifer described in the previous section.

The flow at the boundary is determined by the hydraulic characteristics of the interface. There is flow from the aquifer to the surface water body when the piezometric pressure in the aquifer below the surface water body is greater than the stage (head or water table elevation) in the overlying surface water body. When the head difference is reversed and the aquifer head is lower than the surface water body then the flow direction is also reversed. Based on these concepts, several types of boundary conditions can be used to describe (and model) the surface-groundwater interactions.

$$C_f = K_x A \frac{\partial h}{\partial x}$$

where:

C_f is the conductance between the two water bodies.

Conductance is out of the aquifer and into the surface water body when the head gradient:

$$\frac{\partial h}{\partial x} > 0$$

These boundary conditions apply to most interfaces between surface water and groundwater systems. They have been used to describe the interaction between groundwater systems and rivers, lakes, wetlands and estuaries. Examples are given in the case studies later in the report. Other boundaries can also be described when modelling studies are considered such as distance boundaries and initial conditions.

2.6 Surface Water - Main Characteristics

Technically, surface water can be classified on the basis of its phase (gaseous, liquid or solid), or its setting (precipitation, springs, rivers, streams, lakes, dams, wetlands, estuaries, etc.). However, when the water table intersects the topographical surface the groundwater becomes visible and hence can, and often does, become classified as 'surface water'. No formal definition of 'surface water' was found in any of the references on groundwater. For many purposes it is assumed that 'surface water' is the liquid water that can be visibly observed on the earth's surface. It is separate from the water vapour that forms the atmospheric water and the subsurface water. This definition of surface water is not appropriate when considering the interaction between groundwater and surface water bodies because the water table is frequently observable in many topographical depressions, forming wetlands, pans and even lakes that could also be classified as surface water.

Surface water is a term used frequently but not described by any formal definition, yet it is a term used to distinguish all the resources on the surface from the underground water. However, the rivers, lakes, wetlands and estuaries that occur on the surface are mentioned specifically in the NWA as separate systems. Because the hydrological processes that link these and the underground water are

generally common to all resources, it is not considered necessary to independently examine the interaction between these specific resources and the underground water unless there are identified differences. Since the interaction is conducted through the transfer of water along specific pathways, this study will focus primarily on the processes and pathways, rather than on individual resources. However, it is still necessary to describe the various surface water resources that are influenced by the groundwater system.

In many classification systems, 'wetlands' is the term that covers all aquatic system. There are numerous definitions of wetlands in the literature that reflect the nature of their use. These include the Environment Canada Wetland Classification system (Environment Canada, 2004) and the US Fish and Wildlife Services Scheme for classifying wetlands and deepwater habitat (ref), amongst many others. The Ramsar Classification System for Wetland Type (Ramsar, 2004) is probably more relevant for this study because the Maputaland region has been recognized by the Ramsar Convention as having Wetlands of International Importance. This recognition has been further supported by the declaration of the Greater St. Lucia Wetland Park as a World Heritage Site, highlighting the extreme importance of these wetlands. Table 2 lists the types of wetlands defined by the Ramsar Classification Scheme. These codes are based upon the Ramsar Classification System for Wetland Types as approved by Recommendation 4.7 and amended by Resolution V1.5 of the Conference of Contracting Parties.

In evaluating the importance of groundwater to these wetlands, the main consideration is the contribution of groundwater to the wetland water balance as well as the dependence of the ecology on the different components of the water balance. In many situations, the wetlands rely on groundwater as the dominant component of the water balance, or it forms a significant proportion of the overall contribution. However, there are other wetlands where the groundwater contribution is a small proportion of the overall contribution but it either makes up a dominant and significant proportion of the freshwater contribution or it becomes dominant under specific and important situations (drought or brackish conditions). Wetlands are transitional terrestrial and aquatic forms on the interface between the surface water and groundwater systems. They are created by saturation zones above or below the surface. They include areas that are both permanently and intermittently inundated (Cowardin *et al.*, 2003). Consequently it is important to consider many of the different types of wetlands independently. Four basic groupings of wetlands (Table 6) are examined in this report.

Table 6: Classification of wetland types examined in this report

Wetland groups			
Rivers	Hillslope	Channel	Permanency
Estuaries	Closed	Mixed	Open
Lakes	Dams	Coastal	Off-channel
Other wetlands	Pans		Salt marshes

Since the UNCED Conference in Rio in 1992 it is now widely recognised that the lives of people and the environment, particularly wetlands, are profoundly inter-linked (Acreman and Miller, 2005). Wetlands support many important industries, provide ecological and cultural value through recreation and tourism

and exert a significant influence on the hydrological cycle (Bullock and Acreman, 2003). Some wetlands have isolated hydrological regimes but most are hydrologically connected to other water bodies.

Part 2 of this report describes in more detail the Maputaland wetlands and presents many case studies that investigate the interaction between groundwater and surface water system in an attempt to quantify the importance of the groundwater for this region. The more generic features are described in the following sections.

2.6.1 Rivers

There are an infinite number of pathways that water can travel from the precipitation on the hill slope to its release into the river channel. This has always made it difficult to determine what contribution of the river flow is derived directly or indirectly from groundwater. This is important because any abstraction or utilization from the groundwater zone that forms part of a preferred pathway will impact on the river flow regime.

2.6.1.1 Pathway Concept

To understand the contribution of groundwater to the riverine system it is necessary to review the pathways that water travels from source (rainfall) to the channel outlet. All these pathways involve the catchment surface. Many also follow sub-surface routes that involve the saturation zones (Fig. 15). Horton (1935) proposed that overland flow occurred when the rainfall intensity exceeded the infiltration capacity of the surface. It was subsequently recognized (Hewlett, 1982) that the convergence of flow on the lands surface created partial area of saturation that further induced overland flow. However, much of the rainfall infiltrates the surface and proceeds to flow vertically down under the forces of gravity toward the water table. In a similar manner to the partial area concept for surface flow, the subsurface also develops zones of saturation that induce lateral flow. These zones, often associated with macropore developments that are equivalent to miniature surface pathways, create flow paths to the river through the vadose zone. Finally, the percolating water enters the saturation zone to raise the hydraulic gradient and induce lateral flow toward the river in the phreatic zone creating baseflow.

Beven (2001) has proposed modifications to this model for situations where there is substantial perching of water at surface interfaces in the vadose zone (Fig. 15). This pathway may be indistinguishable from the macropore flow path that is conceived to create the throughflow component observed in storm hydrographs. However, it is clear that the length of the pathway and the zones through which it passes will directly influence the time it takes for the water to reach the channel and its physical and chemical properties. Snyman and Kelbe (2006) have used these concepts to model the hill-slope processes and derive an estimate of the groundwater contribution to the river flow (see the next section).

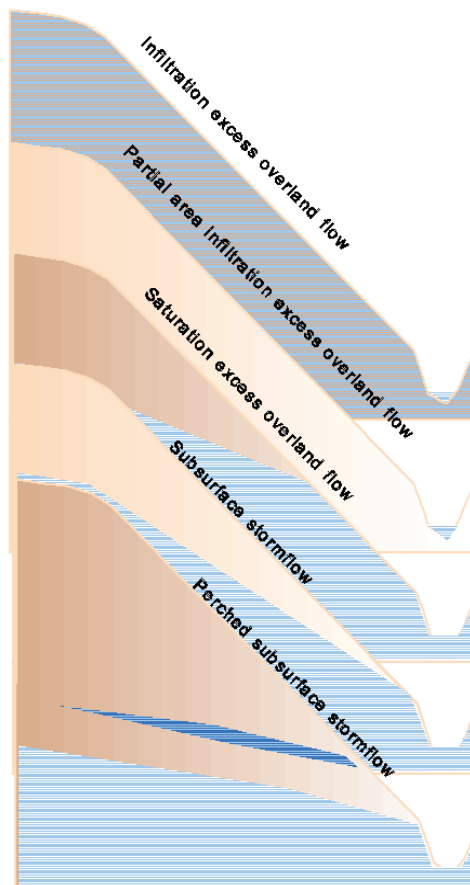


Figure 15: Conceptualized pathways for hill-slope processes (adapted from Beven, 2002)

2.6.1.2 Channel flow

The flow of water in a river or stream generally increases as the flow progresses downstream from the headwater catchments as hill-slope processes continue to contribute to flow into the channel via the various pathways. However, water may flow out of the river channel through transmission losses through various pathways that may or may not return to the channel at some later stage.

The pathway approach has describe the river response to different hydrological processes that induce various flow paths resulting in various runoff conditions that have been observed in the discharge hydrograph.

Rivers have been classified in many different ways depending on selective characteristics (Table 7). Generally, it is pertinent to derive a classification system that is consistent with identifiable conditions. The classification system for ecological purposes is unlikely to be consistent with a classification system required to define the groundwater dependency unless it incorporates the hydraulic characteristics of the interface. Consequently, this section aims to identify a classification system suitable for determining the discharge characteristics of groundwater systems in order to provide information for the ecological reserve determination. On the basis of the physical interaction described in the previous section, the definition of these boundaries is not important. Rather, the hydraulic conditions at the boundary define the level of interaction.

Table 7: Examples of river classification systems

Classifications	Reference
Flow permanency	Joubert and Hurly (1994)
River morphology	Wadeson and Rowntree (1994)
Channel stability	Brierly (1994)
Biotic	Eekhout (1994)
Ecological function	Hart and Campbell (1994)

Stream and river morphology together with the soils and slope of the hillside form the main physical boundaries that constrain surface runoff. These boundaries change over geological time scales. However, the flow of water within these boundaries fluctuates rapidly over a very large range (low flows to floods) which is controlled by climate, weather and other catchment conditions. During extreme events, the physical boundary may change, covering the extent of the entire flood plain. For ephemeral rivers and under drought conditions in perennial streams, there may be no flow in the channel. Under certain conditions there may even be flow in the 'upstream' section of a 'dry' river bed (Todd, 1980). Methods are described in subsequent sections that model the groundwater interaction within the stream channel and the flood plain when river morphology is known.

An example of the simulation of streamflow in eShikawini is shown in Fig. 16 for a river channel that can change from inflow to outflow along the length of the stream channel.

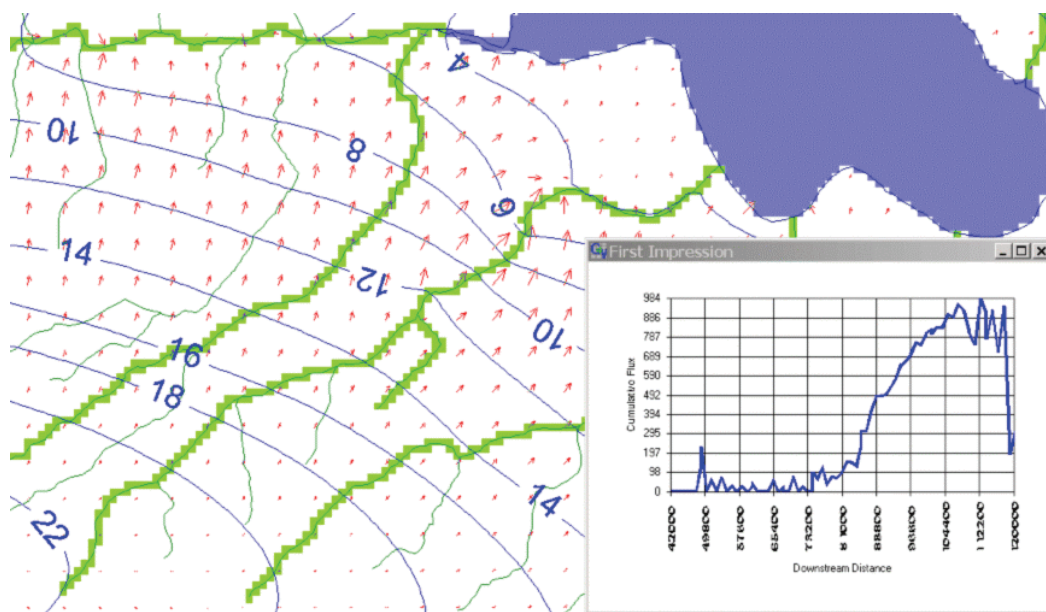


Figure 16: Examples of the contours illustrating the stream sections where groundwater discharge and groundwater recharge occur simultaneously in a stream channel

The groundwater contribution to the river run-off (baseflow) is dependent on the groundwater reservoir (head) and boundary conditions described in the previous section. However, the change in the groundwater reservoir is a direct response to surface processes. Hydrograph analysis and tracer studies

have shown that a large proportion of rainfall can be diverted to surface and near surface flow processes (quickflow and intermediate flow that will limit the recharge to the groundwater reservoir.

While the physically based approach to defining the surface water/groundwater interaction has been described in a previous section, the groundwater contribution to river runoff can be estimated from the analysis of the runoff hydrograph.

The conceptual processes and analytical assessment of the river runoff for small headwater catchments in the Ngoye Hills have been described by Snyman (2005) and summarised in Case Study 1. Hydrograph analysis techniques described by Schulz (1976) were used to partition the river runoff into three identifiable components that portray preferential flow pathways that are generally referred to quickflow, throughflow and baseflow. While it is recognized that there are an infinite number of possible pathways that the water can travel from source to outlet in a catchment, the hydrograph analysis suggests that there are three preferential groupings of these pathways. These have been linked to Horton/Hewlett; unsaturated soil; and groundwater processes (Snyman, 2005). The preferential groupings of pathways through various reservoirs (surface, soil and aquifer) as depicted in Fig. 15 can be described by surface and subsurface flow methods.

These reservoirs (groupings of pathways) are conceptualised as three 2D planes that route the water to the catchment outlet (Fig. 17). The upper reservoir is associated with the topographical surface processes, the middle one with unsaturated soil flow patterns and the lower reservoir is assumed to be the principal groundwater zone. Flow paths through these surfaces are constrained by a plain that is a unique function of the geomorphic features of the surface. The direction of steepest slope (derived from surface elevation - DEM) defines the flow path and its length. The frequency of flow-path lengths creates a unique geomorphic response function (GRF) for the flow pattern in that plain (reservoir). An example for three small headwater catchments in the Ngoye Hills in Zululand is shown in Fig. 18. Applying Manning or Darcian principles (Snyman, 2005) to the GRF, catchment characteristics and cumulative flow patterns, the velocity profile along each flow-path will describe the travel times required to flow from start to end along each pathway.

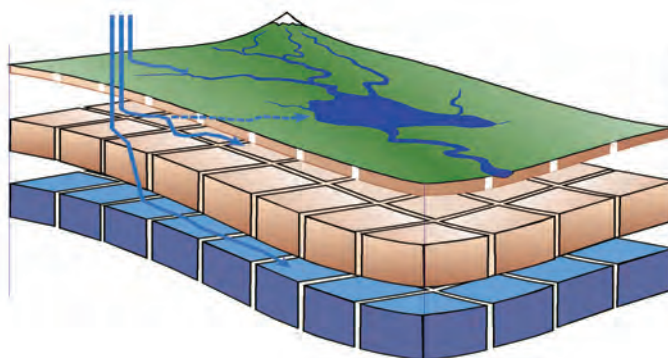


Figure 17: Representation of quasi 3D flow plane where three preferential pathways dominate the flow pattern (adapted from Abbot et al.)

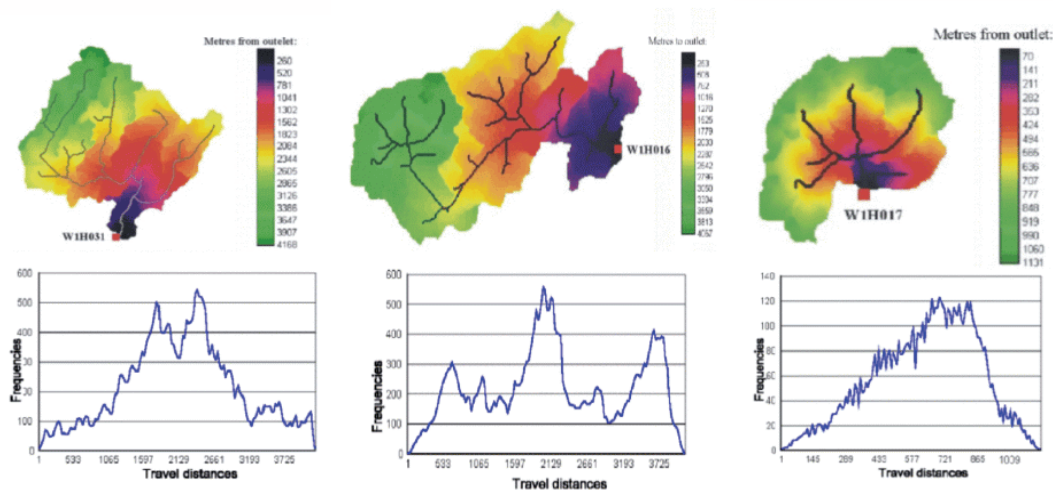


Figure 18: The DEM and associated geomorphic response function for catchment W1H016, W1H017 and W1H031 in the Ntuzi catchment

These Travel Time Response Functions (TTRFs) are scaled to conform to the measured time to peak and recession rates of observed storm hydrograph components through a calibration process. These calibrated response functions are equated to the unit hydrographs for each flow component (Fig. 19). The concept of the unit hydrograph is based on the assumption of a unit of effective rainfall/recharge uniformly dispersed over the plain which is routed to a common outlet. The plain represents a reservoir of water that discharges at specific rates based on the properties of the catchment.

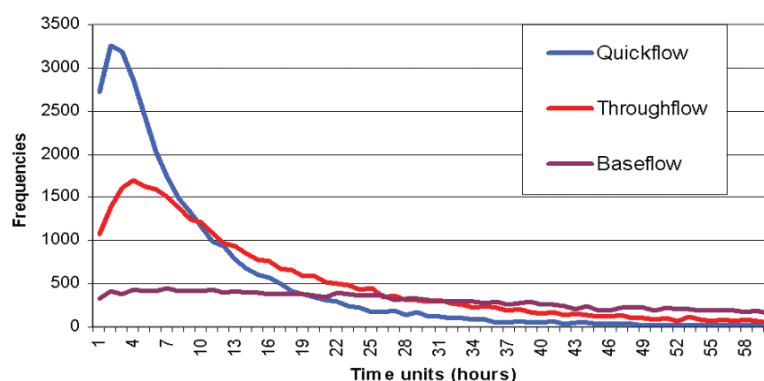


Figure 19: Example of the travel time response function for the three flow components described in the text

To determine the actual contribution to stormflow hydrograph, it is necessary to partition the effective storm rainfall to the different reservoirs. It is well recognised that high intensity rainfall will often exceed infiltration and lead to surface runoff (quickflow) while low intensity storm will generally infiltrate and contribute mainly to groundwater storage (see Case Study 5) by Kelbe and Shezi). Consequently, it is necessary to partition the storm rainfall to the three reservoirs in order to create a convolute storm hydrograph from the three flow components. The convolution process involves the calibration of the storm hydrographs to preserve the cumulative flow for the simulation period (Case Study 1).

This model has been proposed as a method of evaluating the impact of abstraction on the different sources (reservoirs) of water contributing to streamflow. It is proposed that the direct abstraction from

the groundwater will change the flow pattern that will alter the Travel Time Response Function (Fig. 19) leading to a change in the baseflow component. Likewise, any change in land use will affect the partitioning of rainfall into the three reservoirs leading to a change in the storage and subsequently discharge through the three pathways. This method is described in greater detail in Case Study 1 and in Snyman (2005).

2.6.2 Estuaries

The groundwater contribution to estuaries is derived from two principal sources, namely the baseflow component of rivers and streams flowing into the estuary; and the direct seepage into the estuary along the banks and shoreline (Fig. 20).



Figure 20: Example of groundwater seepage from the St. Lucia Lake shoreline

The contribution from baseflow has been described in the section on **Rivers** and is not discussed any further in this section on estuaries. However, the direct seepage has been identified as a very important component of freshwater inflow into the St. Lucia Estuary (Taylor *et al.*, 2006; Kelbe *et al.*, 2005; Kelbe and Taylor, 2005 and Van Niekerk, 2005). Two features of groundwater seepage have been identified by these researchers as important and significant factors in the sustainability of the ecological function of estuaries in Maputaland, particularly Lake St. Lucia:

- Seepage is relatively consistent and persistent over long periods of time in comparison to the highly variable nature of river runoff.
- Seepage zones can be the only source of freshwater and form micro-refugia sites for the survival of aquatic species, both fauna and flora (Taylor *et al.*, 2006).

Estuaries have been classified by Whitfield (1992) into five categories shown in Table 8. A recent national spatial biodiversity assessment of estuaries is presented by Turpie (2004) given in Table 8. The importance of the Maputaland estuaries in terms of these assessments is presented in **Part 2**.

Table 8: The type (Whitfield, 1992) and number of estuaries in each biogeographical region (Turpie, 2004)

Estuary type	Cool temperate	Warm temperate	Subtropical
Estuarine bay	0	1	3
Permanently open	2	29	16
River mouth	2	6	4
Estuarine lake	0	4	4
Temporarily closed	6	87	94

The groundwater is not considered formally in the Ecological Reserve determination of estuaries. However, Kelbe *et al.* (2005) have described the importance of the groundwater contribution to the ecological resilience of Lake St. Lucia which was acknowledged but not incorporated into the St. Lucia RDM desktop study. Kelbe (2002) has also highlighted the role of the reduced groundwater (baseflow) contribution to the demise of the Siyaya Estuary. These two case studies are presented in **Part 2** of this report.

The interaction between groundwater and estuaries can be evaluated in the same way as rivers. The interaction can be evaluated as a boundary to the aquifer or as a process involving various flow paths. The flow paths feeding directly into the estuary will be from contributing hill slopes. However, the river feeding the estuaries will also contain the baseflow component which can be derived in the same manner. Similarly, the estuary and river can form the discharge boundary to surrounding aquifers. An example of the changes in the hydrological features of the Siyaya Estuary and feeder streams is presented in Case Study 3. This is examined using historical hydrological data and numerical models simulating aquifer boundaries.

An example of the importance of groundwater contribution to the St. Lucia Estuary was derived from a mass balance model by Kelbe and Taylor (2005). Figure 21 shows the fluctuation in estuary salinity from June 2002 when the mouth is closed. The red line indicates the simulated conditions with the calibrated groundwater contributions. The blue line indicates the simulated decrease in salinity due to a doubling in groundwater contribution.

Within most estuaries there are a variety of wetlands ranging from salt marshes to mud flats. These wetlands within the estuary may have strong interactions with groundwater seepage (Fig. 22). If the shoreline seepage into these wetlands is similar to those estimated by Kelbe and Rawlins (1992c) for the eastern shores of Lake St. Lucia ($\sim 500\text{--}1\,000\text{ m}^3/\text{d}\cdot\text{km}$) then the groundwater contribution could be a significant proportion of the total flow through these systems on a daily basis.

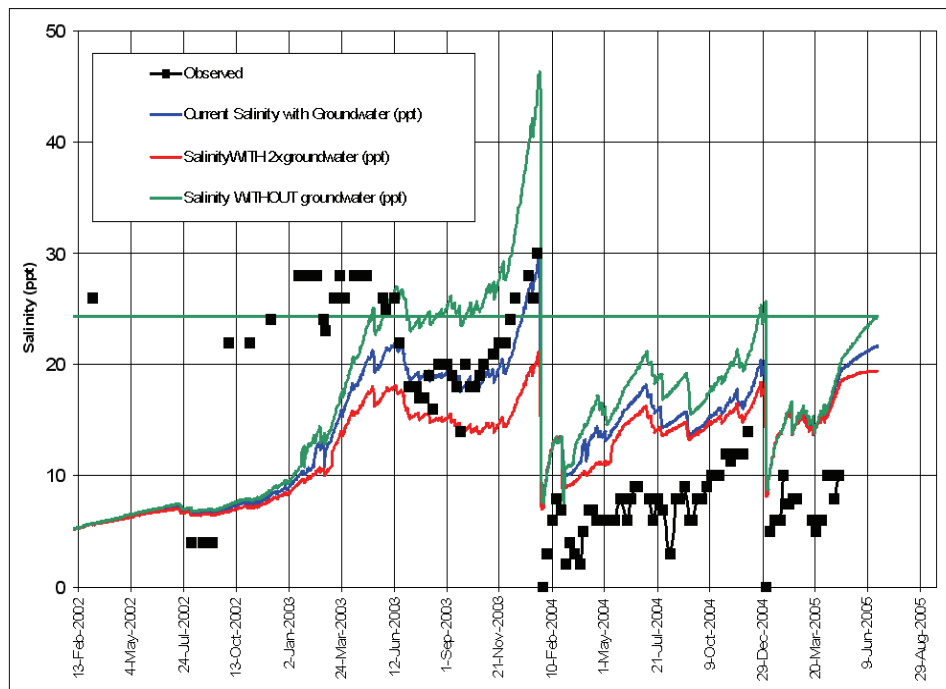


Figure 21: Simulated impact of doubling of the groundwater seepage on the salinity profile for the St Lucia Estuary under mouth closure since June 2002 (Kelbe and Taylor, 2005)



Figure 22: Groundwater seepage into mud flats on the Umlalazi Estuary (Kelbe, 1993)

2.6.3 Lakes

Lakes are large open bodies of water that may or may not have river inflow or outflow. There is no clear distinction between lakes and wetlands and the terms are often interchangeable. Because the definitions are too general to distinguish between all resources, this study has adopted the principle that the different (geo)hydrological pathways are more pertinent to defining the level of interaction between groundwater and surface water. This involves the identification (classification) of the different pathways and then a description of those which constitute the important ones for implementation of the NWA. However, before describing the different pathways, it is necessary to define/describe the various components that make up the underground water that will involve interact through the different pathways. While there are numerous pathways involving the flow of water from the atmosphere through the ground and back to the atmosphere, the probability of selective pathways predominating is

determined entirely by local conditions. To develop a generic conceptual model of the GW-SW interaction, it was necessary to examine the components of the water resources and the main hydrological processes involved in the interaction.

2.6.4 Other Types of Wetlands

The Ramsar Classification System (Table 2) identifies many other forms of aquatic environments other than those covered in the above three classes (rivers, estuaries and lakes). An assessment of the groundwater contribution and its importance requires an understanding of the hydrological features and their dynamic functions. Acreman and Miller (2005) indicate that an assessment should start with a simple conceptual model of how wetlands work and then adopt more complex concepts if necessary. They identify the main components as the wetland soils, the underlying geology and their characteristics such as permeability that control their interaction with the groundwater. However, Restrepo *et al.*, (1998) added several other important features in their hydrological description of wetlands. They indicate that models of wetland hydrodynamics needs to give fundamental consideration to the physics of surface flow processes as well as the groundwater interaction. Surface water flow in wetlands can be represented by sheet flow through dense vegetation but that overall water movement is often dominated by wetland slough channel flows (Restrepo *et al.*, 1998). The sheet flow occurs when the surface is inundated with water, but these wetland areas are also prone to frequent drying that requires an understanding and modelling of the rewetting process. Restrepo *et al.* (1998) describe a model for simulating the groundwater interaction with wetlands that is discussed in the next section.

Acreman and Miller (2005) provide a summary of 14 different ways that water can move into or out of wetlands which they call water transfer mechanism. These mechanisms occur individually or concurrently but they do not necessarily dictate the distribution of water within a wetland but rather they define the hydrological interface with the surrounding aquatic environment and groundwater in particular. However, several underling features can be identified that provide simple models of the interactions:

- The difference in water table elevation (head) between the wetland body and the surrounding aquifer will determine the direction and rate of flow of water between these two systems
- The flow of water in a wetland can vary by orders of magnitude between the matrix flow in the wetland soils to sheet flow and channel flow in the sloughs

This introduces considerable complexity in modelling the interactions which are discussed in the models presented in the next section. Krause and Bronstert (2005) indicate that it is further necessary to delineate the interaction zone between the lowland river (or slough) and the flood plain. This interaction zone is specified as the 'Direct Catchment' which they define as the part of the wetland in which the water balance is mainly affected by surface water dynamics.

CHAPTER 3: REVIEW OF METHODOLOGIES AND CONCEPTUAL MODELS TO QUANTIFY THE GROUNDWATER/SURFACE WATER INTERACTION

Understanding the processes that develop as a result of surface water / groundwater (SW-GW) interactions are becoming more important to protect the integrity of related surface water ecosystems, such as rivers, lakes and wetlands. In many situations, aquifers are fed from surface water bodies (influent conditions), while in other locations the groundwater feeds streams, rivers, lakes and wetlands (effluent conditions). All these resources are used in some way to support social functions such as water supply or land use that directly and indirectly impacts on the system. In order to manage these systems effectively, it is necessary to measure and monitor the impacts in an effort to conserve the resources for future generations. A review of methodologies and models for quantifying the level of interaction is a precursor for establishing monitoring systems.

Primary aquifers occur in areas of increasing importance for development and exploitation. Many important water resources, such as rivers, off-channel lakes and coastal freshwater lakes, are dependent on groundwater which is threatened by surface developments. Agriculture and forestry use and contaminate groundwater to varying extents. Urban, industrial and commercial (tourism) developments with their associated waste also use and contaminate the groundwater. National and international legislation and regulations are placing greater and greater restrictions on the use of these resources to sustain their long-term use. Consequently, there is a great need to develop suitable methodologies for the monitoring and management of these resources.

With the advent of the National Water Act (Act 36 of 1998), the requirements are that a 'reserve' be set for basic human needs and also to protect aquatic ecosystems, in order to secure sustainable development and use of the relevant water resource. To sustain functionality of an aquatic ecosystem, instream flow requirements (IFRs) must be set. The IFRs are generally described in terms of maintenance flows, drought flow and floods. These are transformed into time-series that can be compared with natural historic flow series. The contribution of groundwater to surface water bodies has also been identified as an integral component for sustaining their functionality.

Groundwater monitoring is expensive and consequently, simulation models are providing a quick and pragmatic method for assessment of groundwater systems. These models have been used extensively for many purposes such as environmental impact assessments (EIAs) (Kelbe and Rawlins, 1992a; 1996; 2002; Rawlins and Kelbe, 1997; Germishuyse and Kelbe, 1999b; Kelbe and Germishuyse, 1999a; 2000; Kelbe *et al.*, 2003), Lake Water Requirement (LWR) (Germishuyse and Kelbe, 1999a and Kelbe and Germishuyse, 2000b) studies and waste site investigations (Kelbe and Germishuyse, 1996). However, there are serious inadequacies in the application of these models and further development is required, particularly in the regions where the surface influence is greatest on the groundwater system. The Water Research Commission (WRC) recognized the need for research into the surface water/groundwater interaction.

In a first attempt to quantify the interaction and variants in surface water and groundwater systems, it is generally pragmatic to classify specific types of aquatic systems as either surface water or groundwater systems. This separation is particularly relevant when the resources promote interactions

that are unidirectional and the storage components are easily identifiable and measurable. However, when there is no clear demarcation between surface and subterranean components the classification is extremely complex and requires a different approach. This is particularly relevant when the surface water bodies are mere extensions of the primary aquifers. These systems are common in many parts of South Africa and their importance in sustaining the ecology is not well known. This project has been established to develop conceptual models of the different processes and identify methodologies for their quantification.

It is best to conceptualize the groundwater/surface water interaction occurring through several pathways that include recharge (inflow) and discharge (outflow) zones. However, the internal flow dynamics is important for linking the recharge and discharge zones when considering the interactions as part of a holistic system. The difference in flow rates between these zones creates imbalances in the storage that can be exploited by the environment and/or used for anthropogenic needs. A review of assessment methods and models will include the recharge and discharge as well as water balance methods but the anthropogenic abstraction is beyond the scope of this report. Because of large differences in transit time and residence times in aquifers, the water quality has also been used to estimate the level of interaction between groundwater and surrounding systems. Water availability and water quality have a large impact on ecological systems and consequently, the biodiversity of a system may also provide a method of assessing the level of interaction between surface water and groundwater systems.

The quantification and qualification of the groundwater interaction with the surface water resources is based on a conceptual understanding or model of the systems. The purpose of developing the conceptual model is to simplify the system to achieve the intended purpose of analysis. This means simplifying the field problem and organizing the associated field data so that the system can be analysed more readily. It is recommended that the conceptual model should strive for parsimony by keeping the model as simple as possible yet retain sufficient complexity for the subsequent analysis and assessment (Anderson and Woessner, 1992). In general, the three principal steps are to define the aquifer (hydrostratigraphic units), prepare a water budget and define the flow system components. Anderson and Woessner (1992) go on to suggest that the basic data requirements for developing the conceptual models should be derived from the data sets as shown in Table 9. In this chapter we focus on the delineation of the aquifer and the specification of the hydraulic properties.

Table 9: Basic data requirements for evaluating groundwater importance and its interaction with surface water bodies (from Anderson and Woessner, 1992)

A) Physical framework	
1	Geological maps and cross-section showing the areal and vertical extent and boundaries of the system
2	Topographical map showing surface water bodies and divides
3	Contour maps showing the elevation of the base of the aquifer and confining beds
4	Isopach maps showing the thickness of the aquifer and confining beds
5	Maps showing the extent and thickness of streams and lake sediments

B) Hydrogeological framework	
1	Water table and potentiometric maps for all the aquifers
2	Hydrographs of groundwater head and surface water levels and discharge rates
3	Maps and cross-sections showing the hydraulic conductivity and/or transmissivity distribution
4	Maps and cross-sections showing the storage properties of the aquifer and confining beds
5	Hydraulic conductivity values and their distribution for stream and lake bed sediments
6	Spatial and temporal distribution of rates of evapotranspiration, groundwater recharge, surface-groundwater interaction, groundwater pumping and natural groundwater discharge.

3.1 Mass Balance

The conceptual methods and techniques used to evaluate the groundwater interaction with the surface water bodies were described, for several situations, in the previous chapter. This section will look at the details of the methods and their incorporation in system approaches. In the assessment or measurement of the interaction between groundwater and surface water bodies a simple conceptual model is required that considers the main features of the aquifer and the response in the receiving waters. For the aquifer, a measure of recharge (inflow), discharge (outflow) and storage are essential features that need to be determined. If two components of the system are known then the third can be estimated from simple mass balance principles. Similar principles apply to the receiving waters but in these situations, the aquifer outflow equates to receiving water inflow (and vice versa). In this section, the type of model and associated parameters are presented for evaluating the interactions between surface water and groundwater systems. This chapter will focus on the aquifer properties while the subsequent chapters deal with the recharge and discharge components of the mass balance.

While it is the difference between the inflow and outflow that determines the change in storage in any mass balance model, the inflow and outflow zones need to be established before the differences can be determined. In an equilibrium state, the inflow is assumed to be equal and opposite to the outflow with no significant change in storage. This equilibrium state, one that is often equated to a steady state conditions where no change in time occurs (i.e. no change in storage), is used to estimate discharge from an estimate of the recharge, and *vice versa*. However, this is not a true reflection of the interaction between the aquifer and other water bodies as there are many inflow and outflow paths involving competing processes. The movement of water from the recharge zone to the discharge zone involves internal flow dynamics within the aquifer that can only be evaluated with an understanding of the mass balance. Consequently, it is necessary to determine the spatial extent of the aquifer, incorporating all the inflow and outflow zones. Obviously these physical boundaries also define the spatial extent (volume) of available water for abstraction.

An investigation of the interaction between groundwater and surface water involves the delineation of the intervening surface (boundary). Similarly, the transport of water within the aquifer to the interactive surface boundary requires the delineation of the aquifer unit boundaries. The delineation and specification of these boundaries involves mapping techniques using sparse and incompatible or divergent

data sets. Common techniques and methods for delineating and mapping these features are described in the next sections.

3.2 Hydrostratigraphic Units (HSU)

In the previous chapter, the different types of aquifers were described and some of their hydraulic properties discussed. However, the boundaries between these aquifers are not always clear. In many cases, the aquifers are delineated in terms of their geology. However, many geological features have been described for purposes unrelated to the aquatic setting and this delineation may not be entirely suitable for defining aquifers. Most geological studies in South Africa have been related to mineral exploration and not geohydrological investigation. Consequently, it is not only necessary to delineate hydraulically similar units from geological logs but also to establish their physical boundaries using data from unrelated studies. These units are referred to as hydrostratigraphic units (HSU) in this report.

The experience and knowledge of the different hydrostratigraphic units of the research team is limited to work on the unconsolidated coastal plains of Maputaland. However, the conceptual models and methods presented here are assumed to be equally applicable to other geological settings. The concept of a hydrostratigraphic unit or HSU provides a way of grouping property zones into a single unit with delineated boundaries that can be used to evaluate geological formation that have similar hydraulic features. This is analogous to the 'lumping' of surface features in catchments for rainfall-runoff modelling. Within an HSU, some geological and hydrological properties may vary but the unit is treated as one aquifer for the purposes of assessment. For example a layer confined between impermeable strata would constitute an HSU for the purpose of evaluating the flow and yield of that layer although it may have differing mineralogy.

The first step in defining the HSU is to establish its spatial extent (boundaries) and its basic hydraulic properties. Where the HSU interacts with surface water bodies, it is also essential to establish the extent of the interaction and hence the domain of the complete system. The spatial features involve mapping the important features such as the HSU boundaries described in the next section.

The delineation of the HSU boundaries is generally determined from geological information such as geological maps, geological cross-sections, and well logs. The mapping of the bounding surfaces can also involve the interpretation and interpolation of other hydrogeological and geomorphological information such as bathymetric surveys, river cross-sections and inference from conceptual models.

3.3 Mapping Hydrostratigraphic Units

The mapping of hydrostratigraphic unit boundaries from point sources and other limited spatial data generally involves numerical interpolation and extrapolation methods. Several of the more common methods are Kriging, spline fitting and nearest neighbour techniques. These surface fitting techniques have been described in the literature and have been incorporated into many of the programs that are discussed in more detail in the next section. These surface fitting techniques are seldom sufficient on their own in defining the aquifer boundaries. The project team (Rainaldi, 2005) have used average spatial fields from the different methods to create surface maps that match the conceptual model developed for the system. It is also necessary to force the mapping to conform to conceptual models by subjectively adding data to overcome apparent deformities in the surface profile. This is particularly important when

there are large areas/zones where there is limited or no data. A map of the aquifer basement for Maputaland was created by adding modified bathymetric surface of deep lakes and rivers as well as assumed depth for areas with no data to the other known geological data (surface contours, borehole logs and cross-sectional transects).

Similar approaches are applied to all the surface boundaries to create a model of the aquifer. Once the data set has been acquired and compiled in a suitable form(at) it can be converted to a spatial model (contour map), DEM or TIN for defining the aquifer boundaries at the appropriate spatial resolution. Generally, there is insufficient data for all zones and it is necessary to use intuition or subjectivity based on the conceptual understanding to infill these zones for the creation of the entire boundary. There are large areas of Maputaland where there are no borehole logs that intersect the basement of the aquifer. Several approaches may be considered to create these maps.

- For selective formations, the bedding angle may be assumed from alternate information such as scientific journal and reports. For Maputaland, the basement has been described as a sloping surface of 3-5° by many geologists (King, 1972; Maud, 1980)
- Information about palaeochannels and other geological features that can set upper and lower limits on the elevation of the surface. The bathymetry of the deep coastal lakes is being used to set the upper limit of the basement of the primary aquifer in Maputaland. Several palaeochannels have been proposed for Richards Bay by Worthington (1978), Maud (1968), Lynn (1992) and other geologists that were used in setting up the conceptual model and subsequent map of the subsurface layers by Kelbe and Germishuys (1999a) and Germishuys (1999).
- Variants of the topographical surface also set an upper limit to aquifer boundaries and those areas initially estimated to lie within specified depth of the upper surface can be adjusted accordingly. Fig.25 shows where selected zones have topographical contours extracted to assist in creating the lower boundary.

This entire process is the quantification of the conceptual model boundaries that will be used for analytical studies. This process is used to create the upper and lower surfaces of each aquifer. It also defines the boundary position of vertical flow processes such as recharge and evaporation. The method has been used to try and create other hydrostratigraphic units within a large primary aquifer by Kelbe and Germishuys (1999a) for the Uloa formation in Richards Bay that is hydraulically different to the overlying burden (Fig. 23). Hydrochemical analysis and pump testing has indicated that these hydrostratigraphic units in Richards Bay are not well connected and form leaky aquifers.

One of the most frustrating issues in setting the upper and lower boundaries of aquifers in Maputaland has been the total lack of geological information from borehole logs. The few that have been identified with sufficient information to estimate the depth of the HSU is shown by the set of few boreholes displayed as a red triangle. These borehole logs have been supplemented with geophysical measurements by Meyer and Godfrey (1995) and Worthington (1978).

The techniques for capturing the data set and analysing it are far advanced and commonly available. The more common techniques and system for this report are described in the next section. Sourcing and acquiring the data in an appropriate form(at) is a bigger challenge. DWAF have a substantial database of maps and borehole data that is freely available. However, there is a huge amount of data that has not been captured on their database that needs to be included in the development and quantification of the

conceptual model. Figure 25 gives an example of a large number of independent geological exploration studies, some of which have not been captured in the national data base.

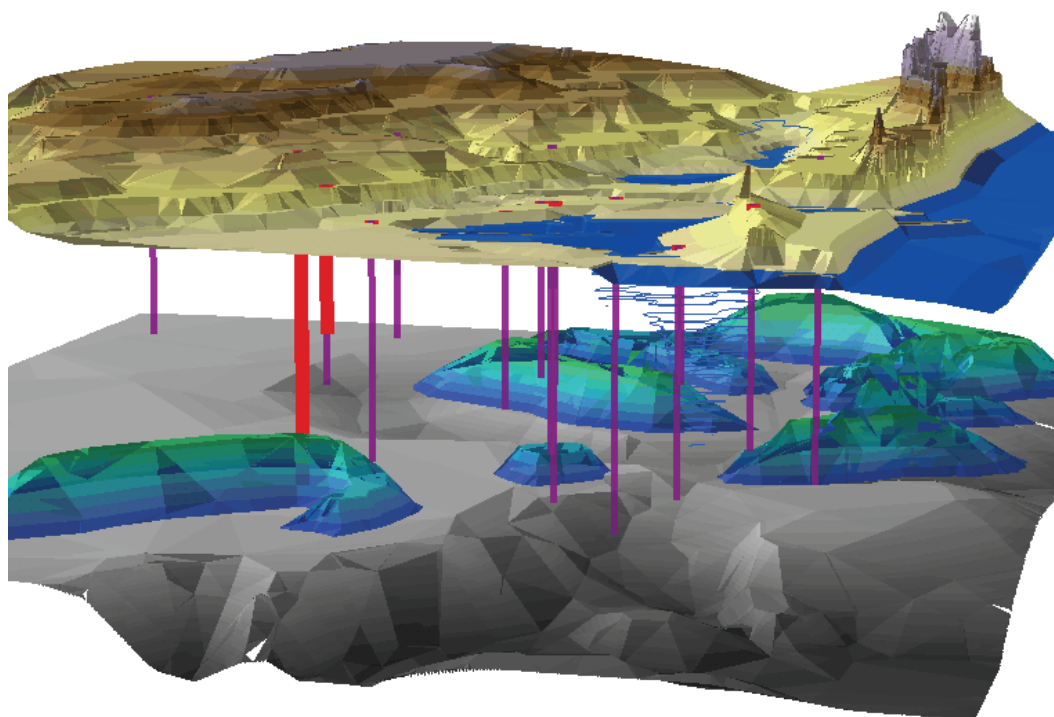


Figure 23: An example of the visualization of a conceptual model of the primary aquifer boundaries (basement in grey scale), surface topography, Uloa units and the location and depth of monitoring boreholes in Richards Bay. Note the exaggerated vertical scale

While the vertical extent of the aquifer is defined by the upper and lower surface, the horizontal extent of the aquifer is controlled by other boundaries which are discussed in the next section. The data requirements for specifying the surface water bodies that interact with the groundwater are also described in the next sections.

3.4 Techniques and Tools for Specifying the Upper and Lower Boundaries of HSUs

Many tools and methods are available for the specification or creation of the upper and lower physical boundaries of the hydrostratigraphic units. There are equally as many methods of displaying and analysing these surfaces. The more common methods of analysis and display is using 2D maps with transects through selective sections of the study area. However, there is a growing use of computer graphic techniques that provide 3D and video forms of display.

3.4.1 Mapping Surface Elevation and Hydraulic Properties

GIS provides many of the necessary tools and functions to assemble the primary data, evaluate its potential contributions to the mapping purposes and then organize the data into a conceptual model of the hydrostratigraphic units. There are many other tools that can perform the same functions but few have been assembled into an integrated system.

The availability of data and their suitability for systems analysis varies radically between aquatic system and between locations. Figure 24 provides an example of the range of data used in the delineation of the primary aquifer in Maputaland. This data set was captured from different sources into vectors (points, lines and polygons) and raster images (aerial photos, satellite imagery, etc.) and then translated into a common coordinate system with similar attributes before transforming them into a map of the study area. The vector form depends on the intended map type. Examples of the different types of maps that can be created by GIS and other systems are shown in figures throughout this report. The most common types of map are based on elevation contours. These can be extended (extrapolated and interpolated) to cover the entire study area using all the available contours and other pertinent data such as river channel elevation, lake bathymetry, borehole elevation, to form a Triangulated Irregular Networks (TIN). TINs can be updated using new data as they become available and TINs can be converted into contours or rasterised to create a Digital Elevation Model (DEM). These maps can often be imported directly into numerical groundwater models such as the finite difference or finite element models discussed later in the report.

Recognizing that the surface is a continuous field of values that may vary over an infinite number of points, it is necessary to consider both scale and resolution of the surface description. The resolution of the grid is generally set by the study requirements and the available data:

A quick and relatively easy method of creating a surface from an irregular set of data values is to form a TIN. The TINs represent the surface as a set of **irregularly** located points linked to form a network of triangles with elevation values stored at the nodes. TINs consist of nodes that store z-values, connected by edges to form contiguous, non-overlapping triangular facets. The edges in TINs can be used to capture the position of linear features that play an important role in the surface such as ridgelines or stream courses. Because the nodes can be placed irregularly over the surface, TINs can have a higher resolution in areas where a surface is highly variable or where more detail is desired and a lower resolution in areas that are less variable or of less interest. The input features used to create a TIN remain in the same position as nodes or edges in the TIN. This allows a TIN to preserve all of the precision of the input data while simultaneously modelling the values between known points. You can include precisely located features on a surface such as mountain peaks, roads, and streams by using them as input features to the TIN. TIN models are widely available and used in finite element models. The TIN can be converted to a raster form commonly known as a DEM. The DEM represents the continuous elevation surface as a **regular** grid of small areas (cells) with sampled or interpolated values. A grid consists of a rectangular array of uniformly spaced cells with z-values. The smaller the cells, the greater the locational precision of the grid. The techniques for interpolating between the irregular arrays of values are discussed in the next section.

A set of contours at the desired interval can be generated from the TIN or DEM. These forms of the surface features (such as the elevation maps shown in Fig. 24) can then be imported directly into programs and models of the groundwater system. This is discussed in greater detail in subsequent sections.

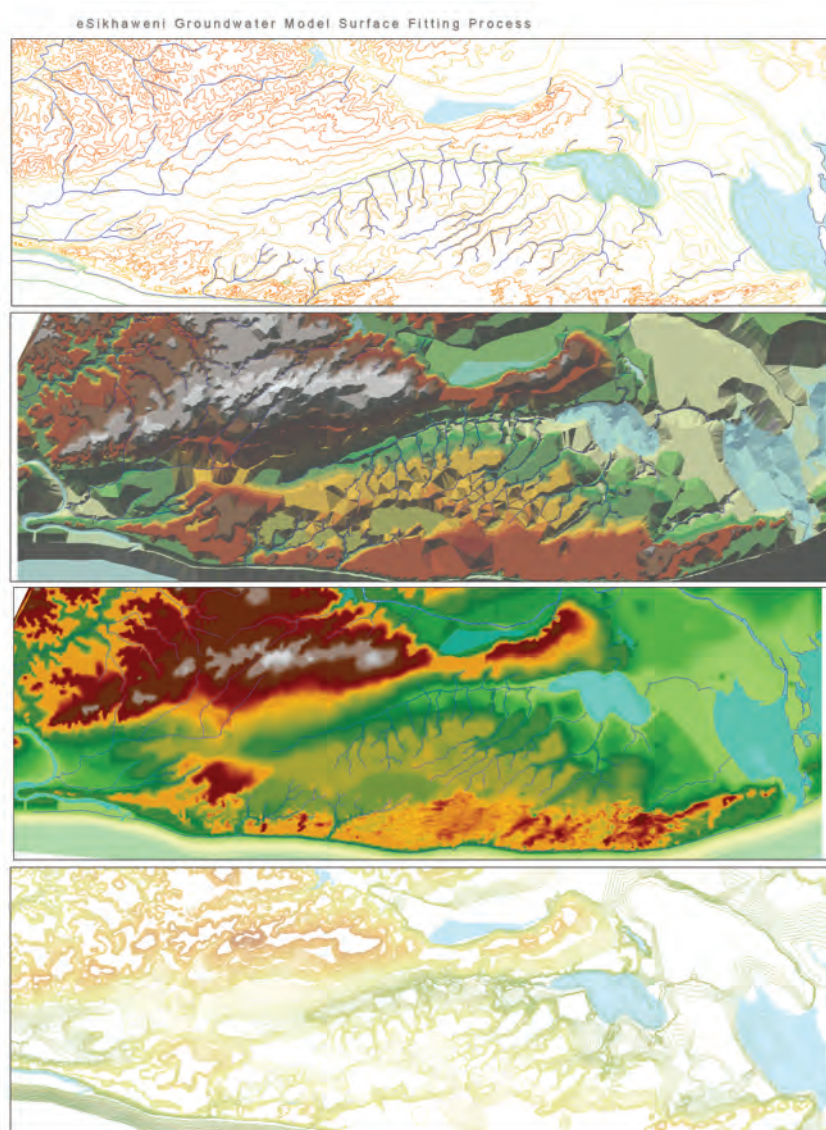


Figure 24: Example of GIS generated surface maps

3.4.2 Interpolation

Many different methods are available for interpolating spatial data to create raster images. The following are examples of techniques in common use in GIS and other spatial modelling programs:

3.4.2.1 Inverse Distance Weighted (IDW) Interpolation

The Inverse Distance Weighted (IDW) interpolation method (Fig. 25) weights the set of points closest to the processing cell more heavily than those farther away using one of several techniques. The IDW method is appropriate when the variable being mapped decreases in influence with distance from the sampled location, e.g. when interpolating a surface of rainfall. The IDW method estimates grid cell values by averaging the values of sample data points in the vicinity of each cell. The closer a point is to the centre of the cell being estimated, the more influence, or weight it has in the averaging process. The weighting is done in several different ways:

- With IDW one can control the significance of known points on the interpolated values based on their distance from the output point weighting the value using a power function. By defining a high power, more emphasis is placed on the nearest points, and the resulting surface will have more detail (be less smooth). The characteristics of the interpolated surface can also be controlled by limiting the number of input points used for calculating each interpolated point.
- Limiting the number of interpolation points is done using either-all points in a specified range (minimum and maximum number of point) within an increasing radius from the surrounding the focal point.-a fixed number of points within the variable radius from the focal point. A line or polygon dataset can be used as a break that limits the search for input sample points. A line can represent a cliff, ridge, or some other interruption in a landscape. Only those input sample points on the same side of the barrier as the current processing cell will be considered.

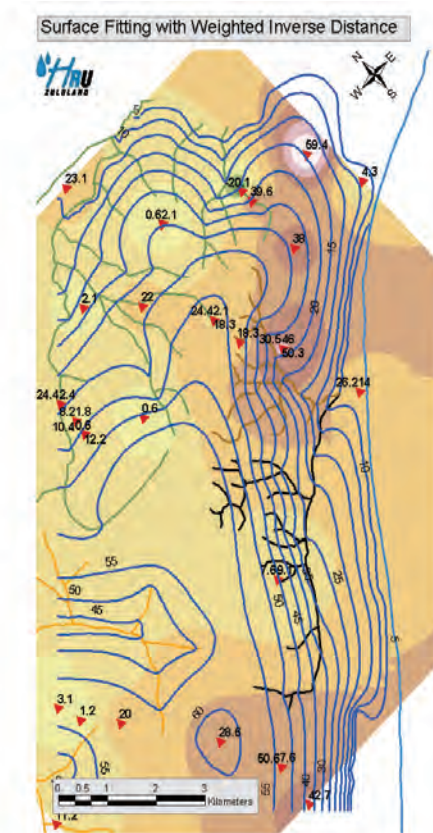


Figure 25: Example of the Inverse Distance Weighted methods of surface fitting using limited monitoring points

3.4.2.2 Natural Neighbours Interpolation

The Natural Neighbours scheme is a weighted-average interpolation method that creates a Delauney Triangulation of the input points and selects the closest nodes that form a convex hull around the interpolation point, then weights their values by proportionate area (AcrGIS, 2005). This method is most appropriate where sample data points are distributed with uneven density. The interpolation method first creates a Delauney Triangulation of the input points (part of the TIN creation process); the points are nodes in a triangulation that follows the rule that a circle around each triangle can contain no other nodes. This means that, collectively, the triangles are constrained to be as close to equilateral triangles

as possible. For each data point, the natural neighbours are the minimum number of nodes in the triangulation that connect to form a convex hull around the point. The weight of each neighbour point is determined through the use of a Thiessen/Voronoi technique that evaluates its area of influence.

3.4.2.3 Spline Interpolation

The spline technique is a general-purpose interpolation method that fits a minimum-curvature surface through the input points (Fig. 26). It fits a mathematical function (a minimum curvature, two-dimensional, thin-plate spline) to a specified number of the nearest input points while passing through all input points. This method is best for gradually varying surfaces such as elevations, water-table depths, or pollution concentrations. It is not appropriate when there are large changes within a short horizontal distance because it can overshoot estimated values (ArcGIS, 2005).

There are two spline methods, namely regularized and tension. Each method has a weight parameter, which changes depending on the spline method used.

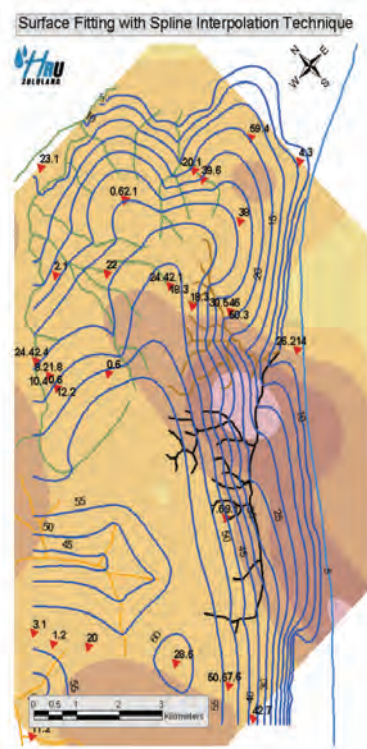


Figure 26: Spline Interpolation Method of surface fitting using limited monitoring point

3.4.2.4 Kriging Interpolation

Kriging interpolation (Fig. 27) is part of a family of geostatistical interpolation methods. Geostatistical methods are based on statistical models that include autocorrelation. The Kriging interpolation method assumes that the distance or direction between sample points reflects a spatial correlation that can be used to explain variation in the surface. Kriging fits a mathematical function to a specified number of points, or all points within a specified radius, to determine the output value for each location. Kriging is a multistep process; it includes exploratory statistical analysis of the data, variogram modelling, creating the surface, and (optionally) exploring a variance surface.

This function is most appropriate when you know there is a spatially correlated distance or directional bias in the data. It is often used in soil science and geology. A detailed discussion of these methods can be found in Isaaks and Srivastava (1989).

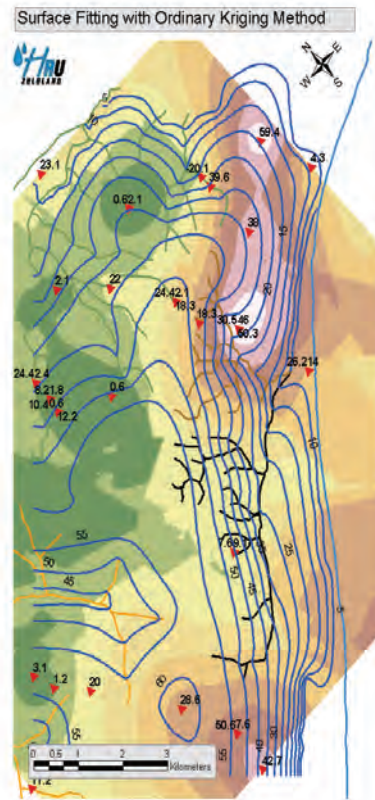


Figure 27: Surface fitting with ordinary Kriging methods

It is essential to use these methods in conjunction with a clearly defined conceptual model of the surface or boundary. Flat areas are problematic and need additional data to infill for the interpolation so it is often advisable to subjectively add data points in flat areas to force the surface to conform to the conceptual model. Figure 27 shows a section in the top right where additional contours have been added to force the model to maintain the elevation in these regions. On the other hand, features such as rivers and lakes provide the upper level limits for such surfaces as topographical elevation and even subsurface layers. GIS and other spatial modelling software provide the necessary tools to create, modify and present these interpolated surfaces in 2D and 3D graphics. Underlying this surface (shown in grey scale) is the lower boundary of the primary aquifer created from borehole logs, published maps and other relevant information (from Germishuys, 2000). Superimposed on the lower boundary are the interpolated surface profiles of the Uloa formation. The vertical lines represent the monitoring boreholes and their depth of penetration.

The same software can often be used to create the volume, area and perimeter of each stratigraphic unit. The application of the use of the volume, area and perimeter is presented in a subsequent section when the discharge from an aquifer is evaluated from recession rates.

3.4.3 Data Sources

There are common sources of data for establishing specific surfaces/boundaries. More importantly, there are numerous other sources of data that need to be obtained for the complete analysis. The most common source of elevation data is the set of 20 m height contours and spot heights that can be obtained from the Department of Surveys and Mapping.

These data are supported by many other attributes that are important for evaluating the interaction between groundwater and receiving water such as rivers, lakes, coasts, etc. The incorporation of these other features is discussed in detail in subsequent chapters. While many groundwater studies do not need to specify the surface topography, any attempt to evaluate the surface - groundwater interaction will inevitably need to establish this boundary. These contours are often the only data available to estimate the elevation of the receiving boundaries that are extremely important for determining the interaction between water sources (see Chapter 5 on Discharge Boundaries and their Associated Hydrological Processes).

3.5 Phreatic and Piezometric Surfaces

One of the most common sources of information for the assessment and evaluation of groundwater resources and their importance to surface water bodies is the piezometric surface or water table elevation. With the available monitoring data (boreholes), draining surface water bodies and other related information, a conceptual model of the water table can be created using the same spatial modelling techniques described in the previous section with the same level of variability. However, it is generally much more pragmatic to develop a numerical 3D model of the aquifer to create the water table surface (Fig. 28) based on the conceptual model of the aquifer(s) that is calibrated against the few irregular network of monitoring points (head and flux). Either way, the water table surface can be incorporated into the software with the stratigraphic unit boundaries to estimate the full extent of the volume, area and perimeter of the water in the aquifer.

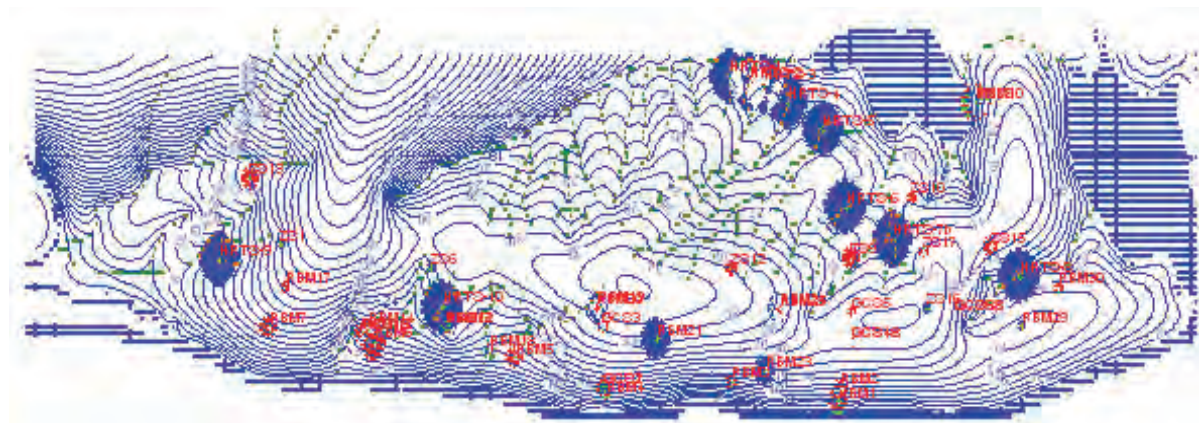


Figure 28: An example of a simulated water table elevation and calibration target for eSikhawini region of Richards Bay

The approach taken to establish the piezometric surface will depend on the available data. Where there are a very limited number of monitoring sites, then it is far more pragmatic to use 3D numerical models calibrated against the few monitoring sites, than to try and create a map from the few sites. This is particularly relevant in the Maputaland region where the aquifer is comprised of relatively few

stratigraphic units that are well drained by clearly recognized surface boundaries (described in detail in subsequent chapters). Figure 28 shows the simulated water table contours for eSihkawini calibrated against less than 10 boreholes and ad hoc measurements of stream discharge (blue circles). It would not have been possible to generate the same level of detail using the interpolation methods described in the previous sections. Typical calibration curves are shown in Figs. 29 to 31. Figure 29 gives an example of a plot between all observed and simulated heads for different groupings (regions) of a model domain for a six year period. Figure 30 shows a comparison between calibrated heads and observed heads for a typical monitoring point over 6 years.

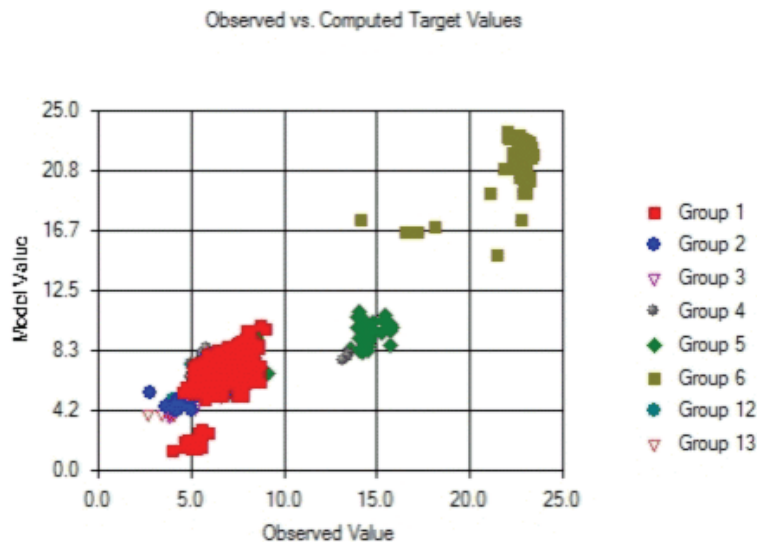


Figure 29: Example of the comparison between simulated heads and observed heads

3.6 Storage and Yield

A change in storage is a transitory feature of the reservoir. Average hydro-climatic conditions generally relate to an equilibrium state in the aquifer where the inflow (recharge) is balanced by the outflow (discharge) and the storage is assumed to remain constant (steady state). This condition assumes that the rate of change of the components is much shorter than the averaging period (t) in the mass balance Eq. (2). Figure 4 (see Section 2) indicates that the groundwater may have characteristic time scales that are longer than a year which could cause annual averages to deviate from steady state conditions in certain systems.

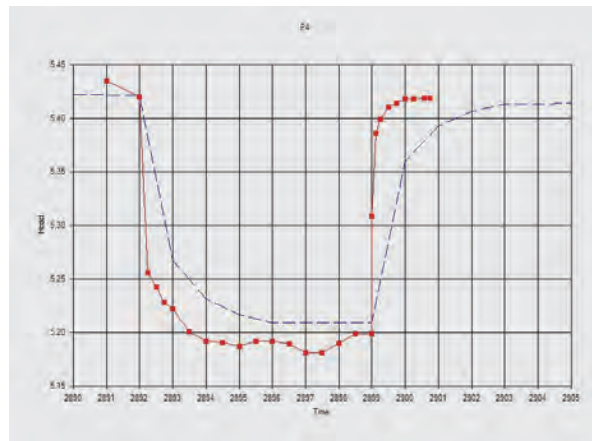


Figure 30: Example of the comparison between simulated and observed heads during a pump test

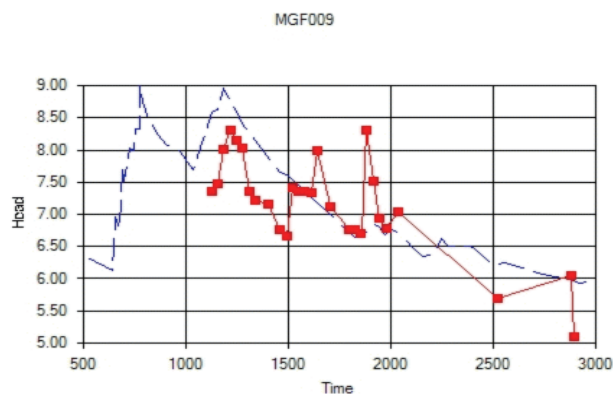


Figure 31: Example of the comparison between simulated and observed heads over 2 500 days

The amount of water that can be held in the pore spaces in an aquifer is specified by the porosity which can be defined as the volume of water in the aquifer pores (interstices) to the total volume of the media. However, the amount that is available for abstraction (specific yield) is limited by the specific retention of the porous media.

Consequently, the amount of water released/absorbed by the aquifer is dependent on the aquifer properties. The change in storage is generally measured by a change in the piezometric elevation (head) of a section of the aquifer which is related to the volumetric change by the storativity (S) of the porous medium comprising the aquifer. The storativity is defined as the specific storage (S_s) which is the volume of water released from storage within a unit volume of porous material per unit decline in head. It is often pragmatic to use the storage coefficient (S) which is the vertically average value of the volume released per unit area of aquifer per unit decline in head ($S = bS_s$ where b is the aquifer thickness). Both S and S_s measure the volume released from an aquifer from aquifer compression or expansion. Consequently, the storativity is closely linked to the specific yield. Ward and Robinson (1990), Todd (1980) and other texts give examples of the relationship between storativity and aquifer composition.

The capacity of the primary aquifer to transfer water from storage is a property of the system known as the storativity and is described by several parameters. Specific yield (S_y) is the volume of water released per volume of porous material ($A \cdot \Delta h$) released by gravity drainage in response to decline of water table (Δh). Specific storage (S_s) is the volume of water released from storage within a unit volume of porous material per unit decline in head. The storage coefficient (S) is the vertically averaged volume of water released per unit area per unit decline in head. Thus, specific storage is related to the storage coefficient in the same way that transmissivity is related to hydraulic conductivity $T=b \cdot K$ and $S=b \cdot S_s$ where b is the aquifer thickness. Pump tests can yield estimates of the storage coefficient (S) and specific storage (S_s) but are subject to large errors (Neuman, 1979 referenced in Anderson and Woessner, 1992). Typical values for S_s are given in Table 10. This has been estimated from grain-size analysis by Morris and Johnson (1967) on the basis of the relationship shown in Fig. 32. It has also been estimated from a relationship with hydraulic conductivity developed by the Bureau of Reclamation (1978) shown in Fig. 33.

Table 10: Typical values for S_s

Material	S_s
Plastic clay	0.0200000-0.002600
Stiff Clay	0.0002600-0.001300
Medium-hard clay	0.001300-0.000920
Loose sand	0.001000-0.000490
Dense sand	0.000200-0.0001300
Dense sandy gravel	0.000100-0.0000490
Rock, fissure, jointed	0.000069-0.0000033
Rock sound	Less than 0.0000033

Mapping the spatial variability of storativity often assumes zones of homogeneity where a demarcated section of the hydrostratigraphic unit has the same storativity. This is normally captured in GIS as a polygon and treated as a lumped parameter in modelling the hydrodynamics of the system.

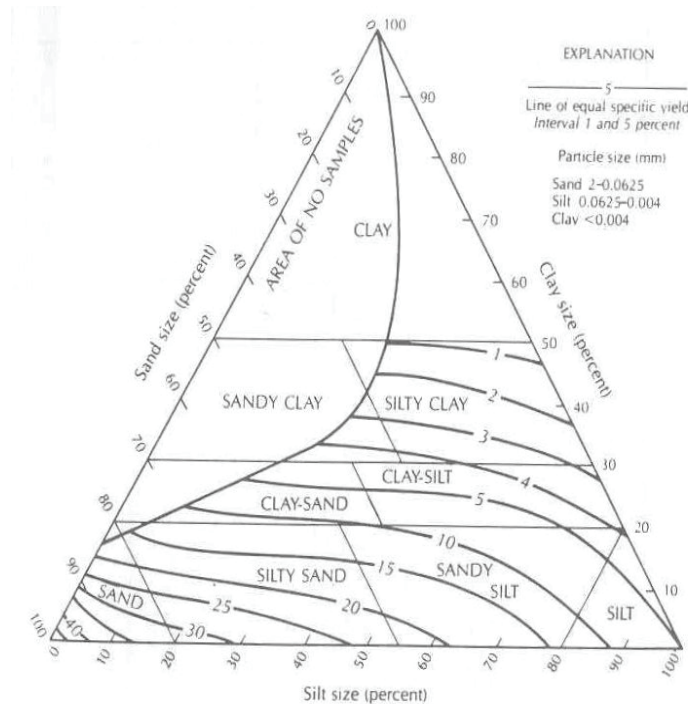


Figure 32: Textural classification triangle for unconsolidated material showing the relationship between particle size and specific yield (Morris and Johnson, 1967)

3.7 Hydraulic Conductivity or Permeability

The flow conditions within an aquifer have been described by Eq. (3). In steady state conditions that are generally assumed to prevail during average hydroclimatic periods, the change in storage over time is zero, so the right-hand term vanishes and the storativity factor is irrelevant. However, in these equilibrium states, the hydraulic head profile will be directly related to the permeability if the recharge and discharge conditions are adequately specified. Consequently, the hydraulic conductivity or permeability of a hydrostratigraphic unit is the main hydraulic factor that determines the flow within an aquifer according to Darcy's Law. The isotropic nature of the permeability is well known in most formations but is very difficult to measure. Consequently, it is often derived through calibration in 2D or 3D numerical models.

3.7.1 Measurements of K

Laboratory measurement of hydraulic conductivity using permeameters requires undisturbed core samples of the aquifer material. This can be obtained with modern drilling techniques but the representativeness of the core sample to the overall hydraulic nature of the aquifer is not well understood. According to Anderson and Woessner (1992) the laboratory measurement of permeability may be several orders of magnitude smaller than the values measured *in situ* (Tanaka and Hallowell, 1966; Herzog and Morse, 1984; White, 1988 quoted in Anderson and Woessner, 1992). They go on to state that 'large-scale features such as fractures, gravel lenses or other types of bedding that may impart transmission characteristics to the hydrogeological unit as a whole are not captured in a sample the size of a laboratory column. Hence, even intact core samples taken from the field for laboratory analyses tend to yield lower hydraulic conductivity values than are measured in the field'. The hydraulic conductivity may also be estimated for unconsolidated sand-sized sediment from grain size analysis

(Masch and Denny, 1966 referenced in Anderson and Woessner, 1996). Probably the most reliable method of measuring the hydraulic conductivity of an aquifer is from pump tests of monitoring boreholes in conjunction with groundwater modelling.

3.7.2 Modelling Hydraulic Conductivity

Steady state models of hydrostratigraphic units usually commence with a calibration of the hydraulic conductivity using various methods to establish a good agreement with the piezometric/head profile from available measurements (target heads). Calibration usually commences with a conceptual model that is adjusted subjectively (within specified limits), using trial-and-error methods, to achieve reasonable agreement with the targets. This can be followed by inverse modelling where the hydraulic conductivity is varied over a range of values to achieve the best fit to the targets. Kelbe and Germishuys (2000b) demonstrated the importance of specifying the different layers in situation where the hydraulic permeability may induce significant changes in the interaction with surface water bodies such as the Maputaland coastal lake system, particularly Lake Mzingazi. By linking the Uloa formation with its higher permeability to the lake, a water balance was achieved that showed a loss from the system of $>20\,000\text{ m}^3/\text{d}$ that could not be explained without this layer.

All estimates of static water level and drawdown from abstraction require an estimate of the hydraulic conductivity. The role and importance of the hydraulic conductance at the interface/boundary between groundwater systems and the surface water bodies is discussed in the chapter on discharge processes.

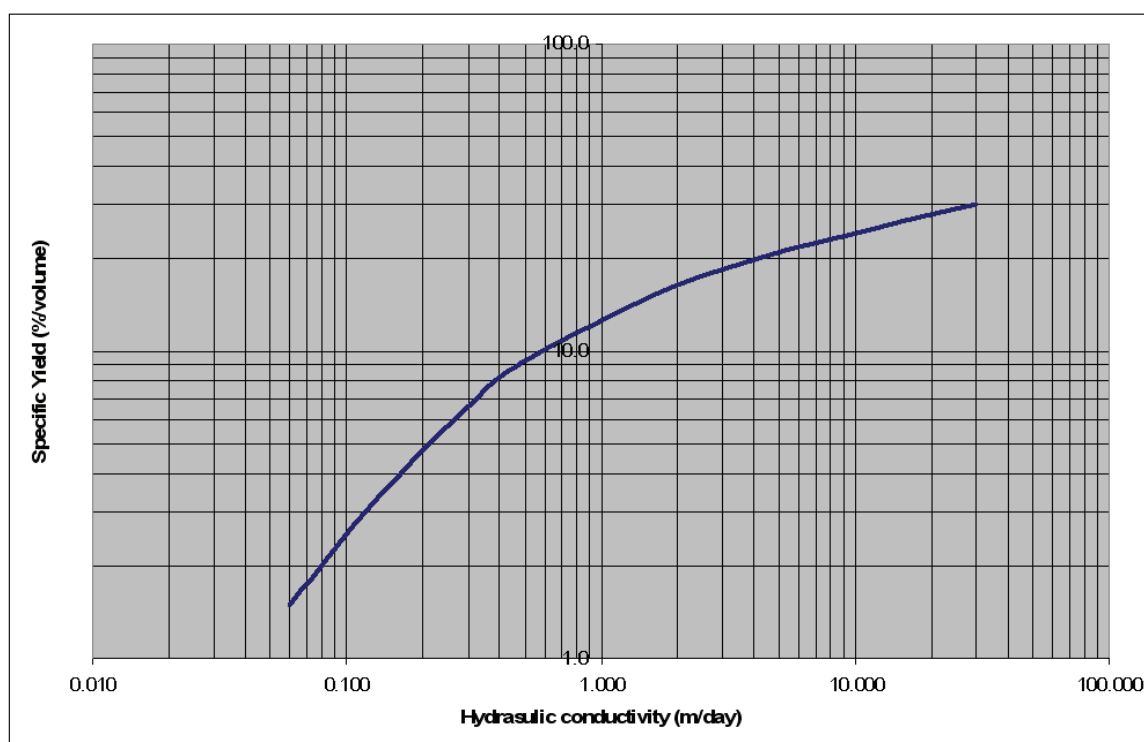


Figure 33: Relationship between specific yield and hydraulic conductivity. Adapted from Bureau of Reclamation (1978) quoted in Boulding (1995).

CHAPTER 4: RECHARGE PROCESSES

The basic mass balance model requires an assessment of the recharge to the groundwater system. Recharge is the process that transports water into the aquifer from other storage units. Excluding artificial recharge methods, the groundwater is generally supplemented by near vertical (gravitational) flows from surface water sources (rainfall, soil moisture, rivers, lakes, wetlands and estuaries) and lateral flows from adjoining aquifers and from dispersion within the aquifer. The distribution within the aquifer can be described by the model in Eq. (3). The near vertical recharge depends on the source of water and the interface or recharge boundary.

The main vertical source of recharge in the Maputaland region is derived from rainfall. However, it is possible that the coastal primary aquifers may also have significant recharge from both river runoff (transmission losses), particularly in the Mkuze wetlands, and deep groundwater flow from the adjacent hills (Ngoye and Lubombo mountains) and the interconnected shallow sedimentary layers. A plan and cross-sectional view of the Mhlathuze catchment is shown in Fig. 34 highlighting some of the interconnected units.

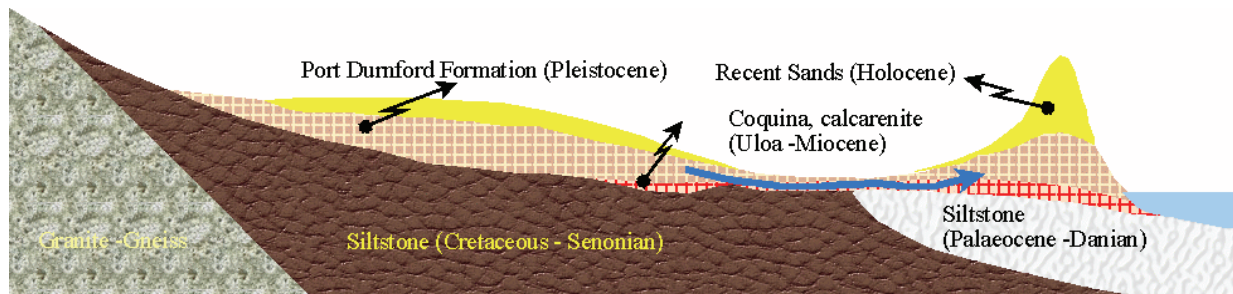


Figure 34: Schematic diagram of a shallow primary system with several hydrostratigraphic units that could conceivably be recharged from adjacent systems. An example is the flow of deep water from the western HSU (left of diagram) through the Uloa to the eastern units

4.1 Background/Historical Methods

Many investigations have been published on the issues of recharge. Beekman and Xu (2003) classify recharge- estimation methods into three categories related to scale of application (hydrogeological provinces, hydrological zones, physical and tracer approaches). They give several approaches that include those listed in **Box 2**.

The methods for quantifying the recharge to groundwater have also been reviewed by Bredenkamp *et al.* (1995) and Beekman and Xu (2003) for South African conditions. Beekman and Xu (2003) include several case studies for different methods. Most of these approaches involve mass balance and/or numerical models of some form.

Box 2: Recharge-estimation methods

(after Beekman and Xu, 2003)

CHAPTER 1	Hydrograph separation (HS)
CHAPTER 2	Channel water budget (CWB)
CHAPTER 3	Watershed modelling (WM)
CHAPTER 4	Unsaturated flow modelling (UFM)
CHAPTER 5	Zero flux plane (ZFP)
CHAPTER 6	Chloride mass balance (CMB)
CHAPTER 7	Cumulative rainfall departure (CRD)
CHAPTER 8	Extended model for aquifer recharge and moisture transport through unsaturated hard rock (EARTH)
CHAPTER 9	Water table fluctuations (WTF)
CHAPTER 10	Groundwater modelling (GM)
CHAPTER 11	Saturated volume fluctuations (SVF)
CHAPTER 12	Equal volume spring flow (EV-SP)
CHAPTER 13	Groundwater dating (GD)

These methods have been applied in different areas of South Africa by various researchers. Most of the methods are based on a mass balance of selective constituents of the system and/or use numerical models that usually invoke mass balance constraints. The mass balance is generally based on average (long term) conditions where it is assumed that the system is in equilibrium (non-transitory) state. For transitory systems, numerical models that are either physically based or empirical driven are commonly used. These methods have not been included in this report. Instead, an evaluation of simple methods that have been utilized to investigate the interaction between the groundwater and surface water bodies, using examples from studies in the primary shallow aquifer in Maputaland are presented in this section.

4.2 Groundwater Response to Rainfall

The present conceptual understanding is that the major sources of recharge to the shallow primary aquifers in Maputaland are from direct rainfall. Consequently, the relationship between the incident rainfall and the groundwater (phreatic surface) response would give an indication of the magnitude of this interaction. This section looks at the response at various scales and at different piezometric depths within the aquifers(s).

4.2.1 Annual and Seasonal Response Functions

A typical response in the groundwater elevation (BH117) to variations in rainfall over annual cycles is presented in Fig. 35. The conceptual understanding is that the aquifers have an upper limit to storage and the relationship in Fig. 35 should tend to an asymptote that reflects this upper limit.

Figure 35 illustrates an initial relationship between water level and rainfall when the aquifer is partially depleted and the asymptotic level when the aquifer reaches maximum capacity. The possibility of transient effects can be assessed from shorter time scale relationships. Figure 36 shows the groundwater response to monthly rainfall for the same aquifer. Correlation analysis shows no lag in storage (see Case Study 4) but fails to show any asymptotic conditions at the inter-annual scale

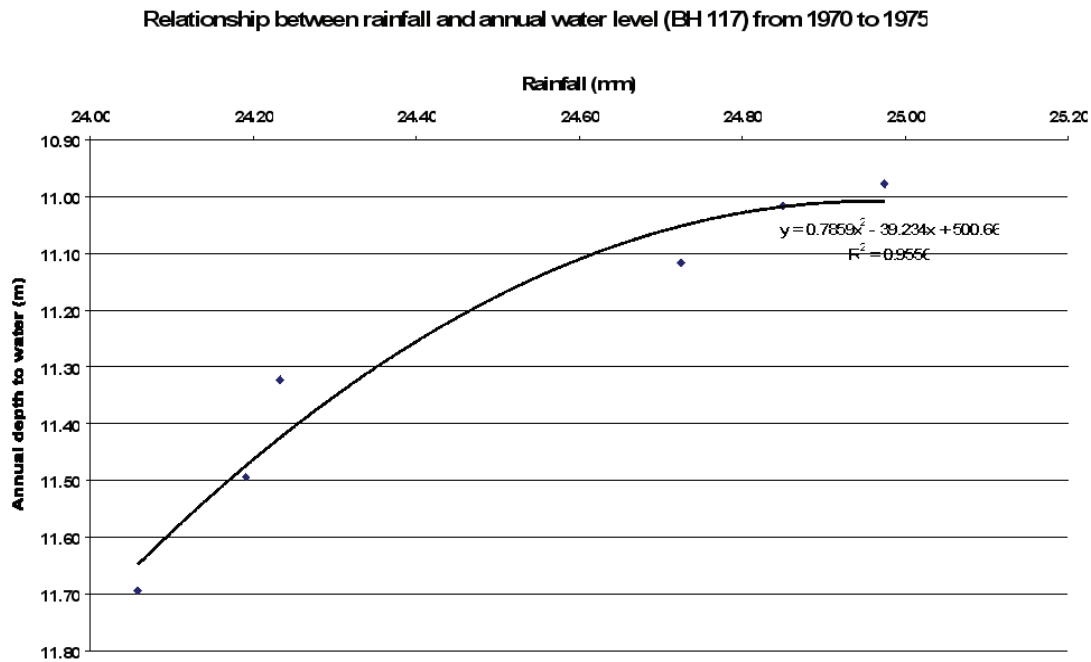


Figure 35: Annual rainfall-groundwater response function

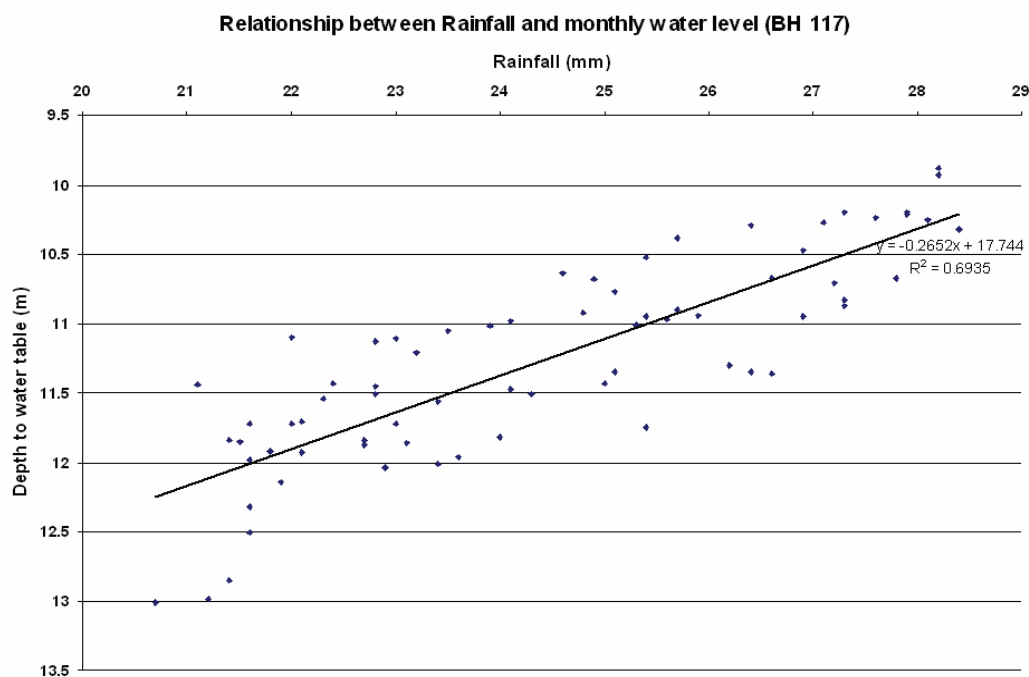


Figure 36: Relationship between monthly rainfall and water elevation

4.2.2 Event-Based Relationship

The response in the phreatic and other piezometric surfaces to short-term rainfall events have been investigated and are summarized in Case Study 4. Hourly, and in one case 15 min (Fig. 37), measurements of rainfall and piezometric head fluctuations have been examined for four sites on the Maputaland coastal plain. The phreatic surface shows very rapid response to rainfall fluctuations that varies with the piezometric elevation (depth of the unsaturated zone).

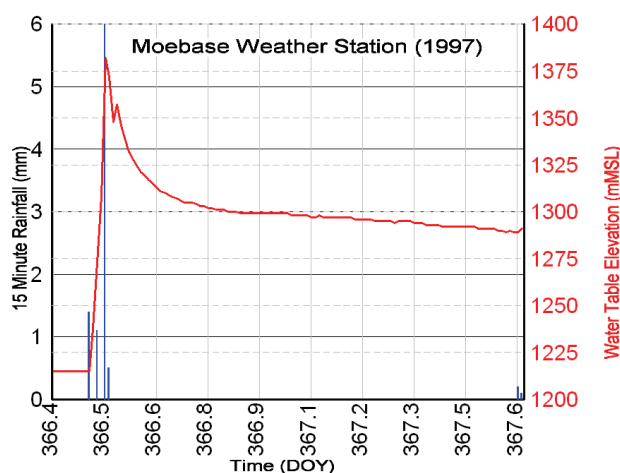


Figure 37: Hydrograph for the phreatic surface in response to 15 min rainfall for a very shallow aquifer

Similar responses can be seen within the primary aquifer in Richards Bay for three boreholes drilled strategically, one above the other, in the Maputaland coastal aquifer (Alton, Richards Bay). The upper series represents the phreatic surface (6 to 9 m screen), the middle borehole is in the Port Durnford formation (23 to 26 m screen) and the lower BH is in the Uloa formation (33 to 36 m screen). All three formations comprise sand with varying clay content (1 to 10%). Several features of the water level response are noteworthy:

- All boreholes show a sharp rise in pressure in response to rainfall. However, more detailed studies show a distinct lag with depth (see next section). In most cases the Time to Peak (TtP) varies from 2 h to 36 h depending on the rainfall condition (Fig. 38).
- The phreatic surface shows a rapid recession after the rainfall that has similar characteristic features as river runoff hydrographs. There appear to be three recession rates in the groundwater.
- The deeper formations show slower recessions with two possible rates.

These traits are observable at all the monitoring sites along the coastal plain of Maputaland and Mozambique. The time to peak elevation in piezometric head shows a linear relation to the rainfall rate and the depth of the unsaturated zone (Fig. 39). The rate of rise in the piezometric surface is related to the depth of rainfall and this relationship varies with location (Table 12).

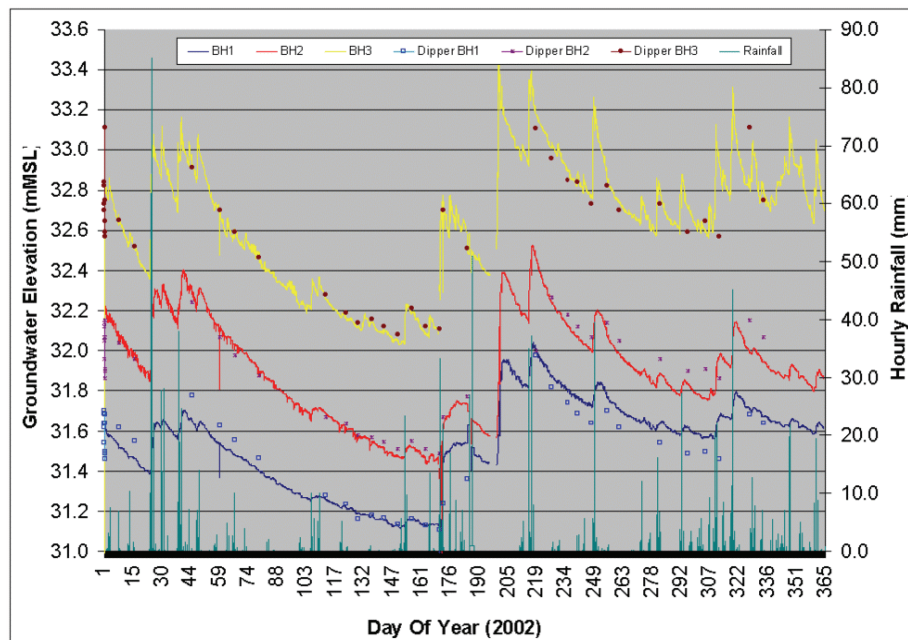


Figure 38: Hourly water level series for three boreholes in Richards Bay showing the response to rainfall at varying depths

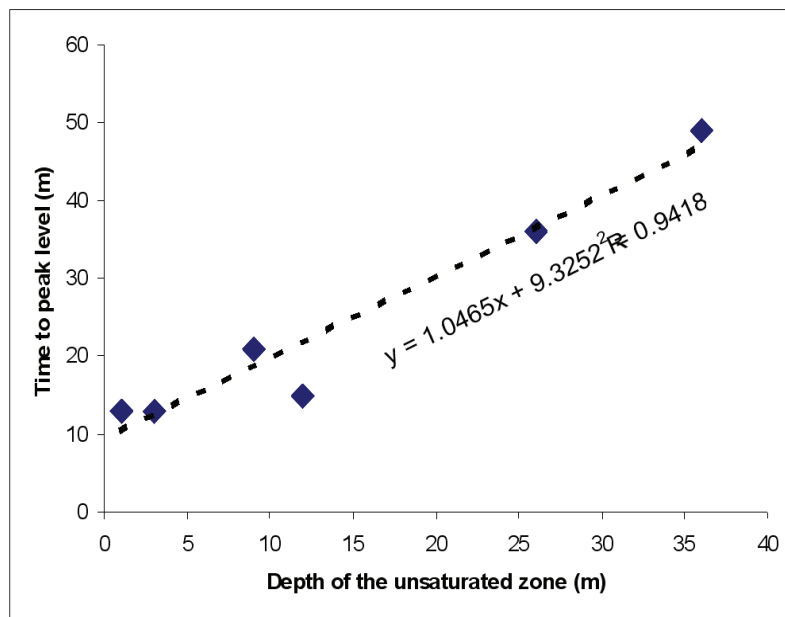


Figure 39: Relationship between depth of unsaturated zone and time to peak in head

4.3 Estimation of Recharge

In the estimation of recharge to groundwater from rainfall, it is important to know the purpose of the assessment. In an equilibrium state, the recharge equals ALL of the discharge components, including the loss to evaporation (E), surface water bodies (R) and redistribution within the aquifer (D) (Fig. 40). In most methods of regional recharge assessment, many of the components are ignored as insignificant. However, in evaluating the interaction between surface and ground water system, all components need to be estimated. This is particularly important if evaporation (E) from the groundwater occurs.

$$\text{Recharge} = P - E - R - D$$

In many situations the recharge is given as a proportion (%) of the precipitation (P). This makes a simplistic assumption that the other components are directly related to the precipitation and uniform under all conditions a situation that may only exist under long term conditions.

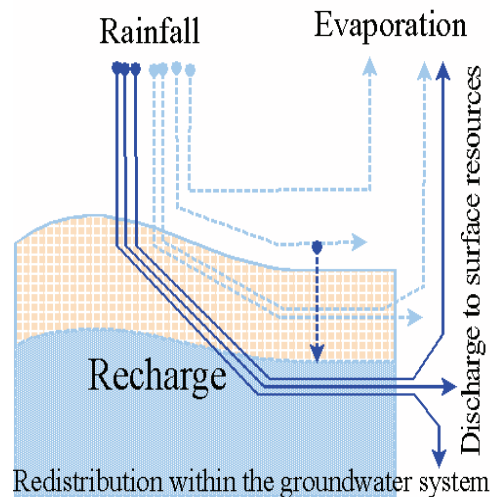


Figure 40: Conceptual pathways and partitioning of rainfall to groundwater

For these long-term conditions, Kelbe and Germishuysen (2000a) applied binomial filters to rainfall series to establish the predominant recharge episodes and then made subjective adjustments for major recharge events (Fig. 41). However, they modified their model for short-term conditions by excluding all rainfall exceeding 50mm/d as direct runoff, and then partitioned the rest over 5 d.

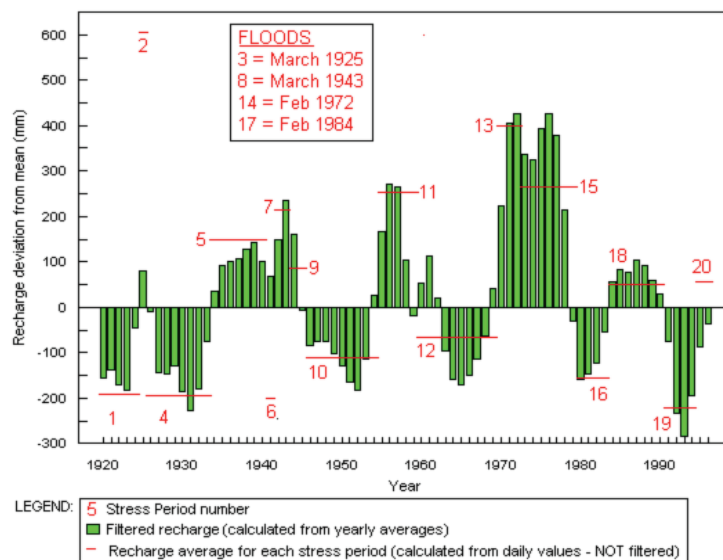


Figure 41: Example of the application of a 5-term binomial filter applied to annual rainfall to establish major recharge episodes for the primary aquifer in Richards Bay

This model was derived from studies by Rawlins and Kelbe (1991a) for St. Lucia where a linear relationship between rainfall and water table response was observed for all storms exceeding 10mm/d. Studies are continuing to determine the volume of groundwater associated with the response functions

and the volume of rainfall to get a more accurate assessment of the proportion of rainfall that enters the water table.

4.4 Variations in Head

The piezometric surface for three formations (HSUs) in Alton, Richards Bay varies by 0.5 to 1.0 m in elevation and even more at Meerensee, also in Richards Bay. These same HSUs show a direct response to the rainfall event but with increasing lag (Fig. 43) and rate of rise (Table 11). The pressure wave generated by the rainfall load takes, on average, up to 40 h longer to propagate to the deepest unit.

Table 11: The relationship between the rate of rise of water level and the rainfall amount for different sites (the rate is the rise/time to peak)

Site	Richards Bay			St. Lucia	Moebase	Moma
Depth	6 m	26 m	36 m	6 m	1 m	12 m
Rate	0.0009	0.0007	0.0001	0.0055	0.0047	0.0028
r ²	0.9590	0.9630	0.8050	0.9573	0.9288	0.8740

The relationship between the rate of rise in the water table and the rainfall has been described in Case Study 5 (Kelbe and Shezi). An example is shown in Fig. 42 for a borehole on the coastal plain in northern Mozambique.

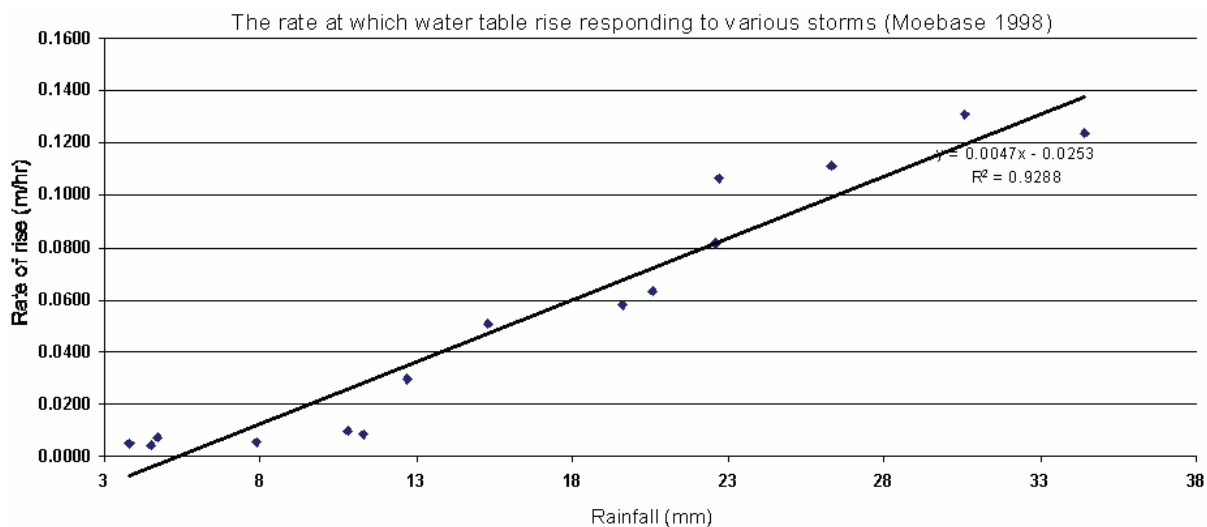


Figure 42: The relationship between rate of rise in head and rainfall for a primary aquifer on the coastal plain in northern Mozambique

4.5 Variations in Water Quality

A current investigation in Richards Bay shows that the water quality in the Uloa Formation (Unit 3 in Fig. 43) is decreasing (increasing TDS) while the overlying formations (Units 1& 2 in Fig. 43) remain unchanged. The principal source of the increase in TDS is magnesium ions. A geochemical analysis of the borehole samples by Mkhwanazi (2010) shows no appreciable levels of minerals that can be linked

directly to this increase in TDS. The initial reaction is to ascribe the increase to solute transport from unknown recharge zones.

Coupled with the variation in piezometric head described above, this is being interpreted as leaky aquifer with substantial differences in recharge and discharge rates that create a head deficit of >1m when the recharge in the lower unit is lower than the discharge.

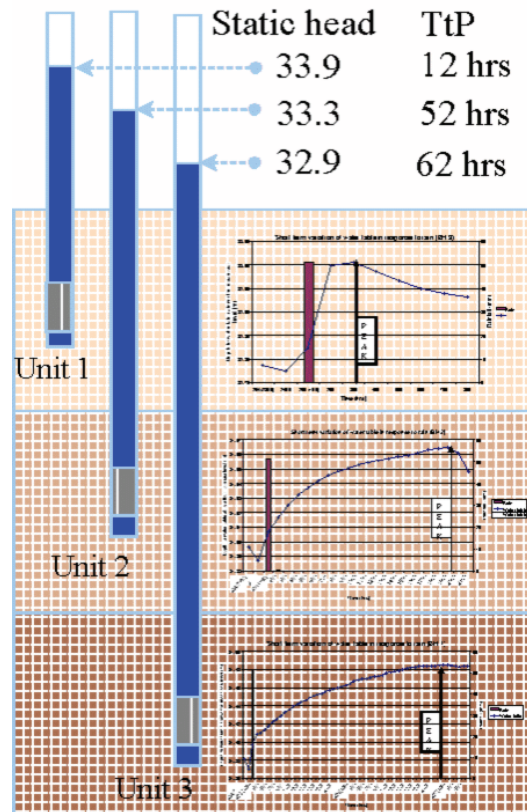


Figure 43: The composite variation in the piezometric response at three depths to a rainfall event in Richards Bay

4.6 Snyman and Kelbe Model based on Stormflow Hydrographs

The hydrological hill-slope processes have been presented by Snyman and Kelbe in Case Study 1. They identified three preferential groupings of pathways that led to the characteristic storm hydrograph often observed in the Zululand headwater catchments. Ignoring the flow in river channels where the various flow components become mixed over distances greater than about 5 to 10 km, the rainfall is partitioned into storage zones on the hill slope, referred to as hill-slope reservoirs, which are recharged and drained according to hydrological principles. The recharge to the hill-slope reservoirs depends on storm type and catchment conditions.

The discharge from each reservoir depends on the slope of the pathway and resistance to flow (conductance) related to the conditions along the pathways. This concept led to the creation of a unit hydrographs for each preferential flow component (quickflow, intermediate flow and baseflow) that were scaled for storm size and convoluted to create a storm hydrograph.

The pathways were derived from geomorphological conditions on the hill slope (direction of steepest gradient). These draining reservoirs were combined by scaling the magnitude of the discharge to match the storm runoff volume (effective rainfall) and the recession rates of the different components.

These calibrated component hydrographs were derived for four different storm types that reflected short- (long) duration and high- (low) intensity situations that created different partitioning of the effective rainfall to the three reservoirs. The results of the model were compared to the observed storms in the Ntuze catchments.

The time to peak in this model is when all of the hill-slope paths are contributing to the runoff for the particular component. If the storm duration is shorter than the longest time taken to flow to the channel (outlet) then the time to peak is determined by the shape of the travel time response function created from the geomorphic response function. Consequently, the time to peak is a complex relationship between catchment conditions and storm type.

CHAPTER 5: DISCHARGE BOUNDARIES AND THEIR ASSOCIATED HYDROLOGICAL PROCESSES

The last section of the recharge description highlighted the relationship between the recharge and discharge in primary aquifers where most the surface-groundwater interactions occur. Consequently, it is considered imperative that any study of the groundwater recharge must be accompanied by an assessment of the discharge in order to establish the magnitude and range of the recharge. This section describes many of the discharge features that were examined for the Maputaland region.

In the assessment or measurement of the interaction between groundwater and surface water bodies a simple conceptual model is required that consider the main features of the aquifer (Fig. 24) and the response in the receiving waters. Similar principles apply to the receiving waters but in these situations, the aquifer outflow equates to receiving water inflow (and vice versa). In this section, the type of model and associated parameters are presented for evaluating the discharge processes that describe the interaction between surface receiving waters and groundwater systems. The discharge through the different boundaries is dependent on the processes operating at the boundaries. Evaporation (vertical flux) involves a change in phase while the seepage through the boundary into rivers, lakes, etc., is a direct flow of water. Consequently, the different boundaries need to be evaluated separately.

The discharge of groundwater into stagnant water bodies such as the ocean, large lakes, estuaries and wetlands will induce a response in the receiving body that is generally undetectable in the standard method of observation (head) but may be measurable using specialized equipment such as seepage meters (see examples in Chapter 6). Except for tidal variation, the oceans are perceived as a constant elevation and no amount of groundwater seepage will ever be detectable in a simple mass balance approach. Similarly, it is extremely difficult to measure the response to groundwater seepage into/out of lakes, estuaries and large wetlands where the stage (head) is the standard method of observation if the groundwater is only a small proportion of the total inflow. Consequently, these large water bodies rely on the estimation of groundwater contribution from point source measurements, indirect estimates and models involving mass balance principles.

Rivers and small receiving water bodies where the groundwater contribution is a major component of the inflow will generally exhibit a response that is detectable in the standard methods of observation. In rivers, this is generally considered to be the baseflow that can be estimated from river runoff analysis. In small water bodies such as wetlands it will be reflected in the persistence in elevation (head) of the water surface.

These different responses of the aquifer boundaries to groundwater seepage require separate evaluations. The response in the receiving water bodies needs to be compared to the observation in the aquifer to establish a mechanism that can assist in the conceptualization of the boundary processes.

5.1 Aquifer response to discharge processes

The rate of discharge from an aquifer is observed in the water level hydrographs from strategically placed wells and other monitoring points. Examples of borehole hydrographs have been shown for different locations on the coastal plains of eastern Africa. Examples of the recession limb of storm hydrographs for shallow phreatic surfaces are given in Figs. 44 and 45. In the shallow groundwater hydrographs for the coastal plain at four different locations (Richards Bay, St. Lucia, Moebase and Moma), the recession limb shows three characteristic recession rates that have some similarity with the river hydrographs. However, the groundwater recession rates appear to be more linear than the exponential rates observed in most of the river hydrographs (see Case Study 1). However, it is assumed that the aquifer acts as a draining reservoir and that these different recession rates are derived from different reservoirs in the system in a similar concept to that presented by Snyman and Kelbe in Case Study 1 for hill slopes and rivers.

Schulz (1976), Todd (1980), Linsley *et al.* (1975) and Gordon *et al.* (1993) all suggest that a draining reservoir can be described by a simple exponential decay function of the form:

$$\frac{q_1}{q_0} = K_r^t$$

where:

q is the discharge at time t in days

K_r is a recession constant

This equation will plot as a straight line on a semi-log graph of $\ln(q)$ -vs- t .

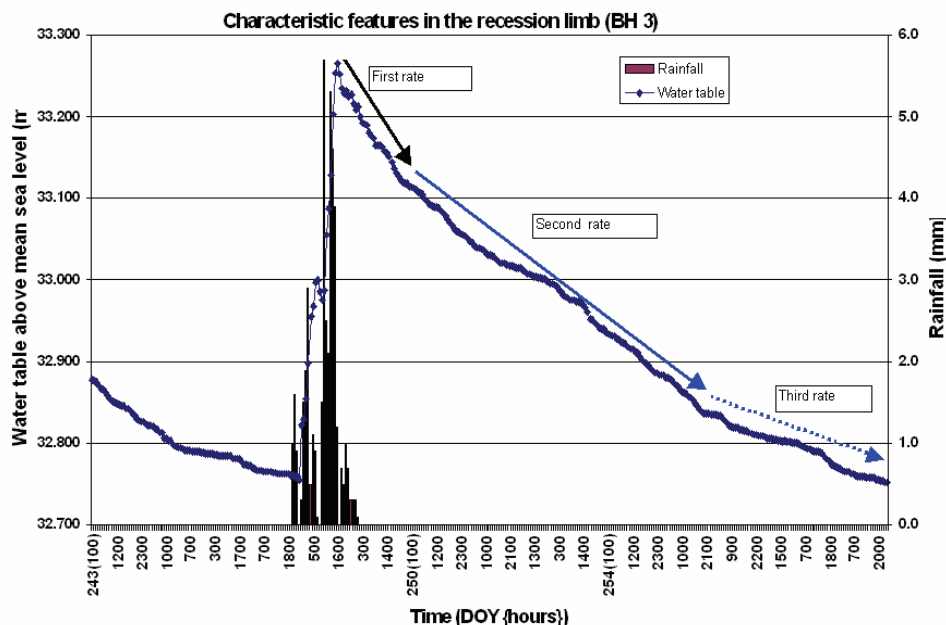


Figure 44: Typical groundwater hydrograph for a shallow phreatic surface in Richards Bay showing the different recession rates

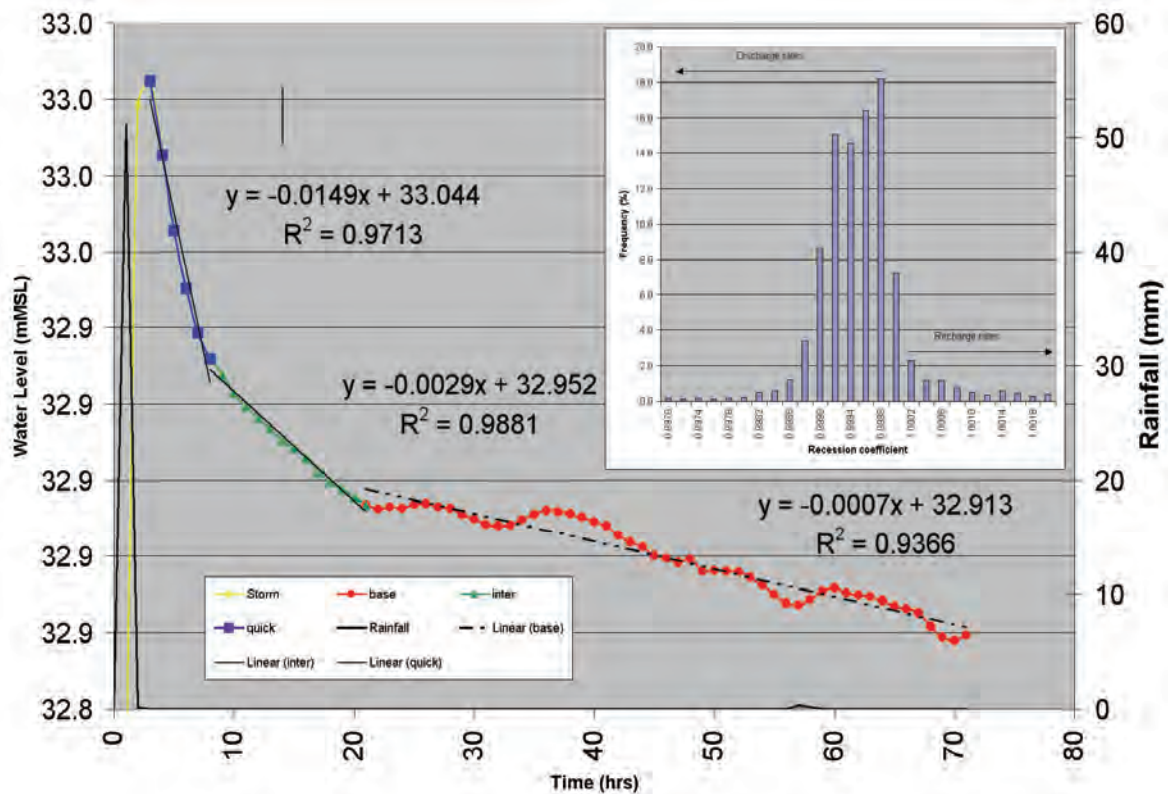


Figure 45: Groundwater hydrograph analysis of head showing similar features to the discharge hydrograph of streams

If the groundwater reservoirs behave in the same or similar manner, then the recession rate of the discharge would be described by a similar constant which is governed by the hydrogeological characteristics of the aquifer. This concept applies to the discharge (flux) which is not measured in aquifers. Generally, it is the piezometric head that is monitored and this has a complex relationship with the discharge rate (q) that depends on the boundary conditions and the aquifer properties. Consequently, a different analysis is required when analysing the recession rates in Fig. 44. However, Fig. 45 shows an example of the recession curve for a groundwater hydrograph that has similar characteristics to the recession curve often observed in rivers and described in the next sections. This example for a shallow primary aquifer shows three recession rates that have been labelled as 1st (quick), 2nd (intermediate) and 3rd (base) recessions. The slope of the linear recession rates (equivalent to K coefficient) varies from 0.015 (m/h) for quick response to 0.0007 (m/h) for the slow response. If an exponential decay of the form above is applicable to this recession rate of head, then these coefficients would be equivalent to 0.046, 0.23 and 0.99 for the quick, intermediate and base rates respectively. However, the relationship between these rates and those observed in river runoff is not well understood.

The frequency distribution for the recession rate for daily time steps for one year of data for the same borehole is also shown. Values greater than 1 indicate recharge while those less than 1 reflect the discharge rate. The recession rate is generally between 0.9998 and 0.9999. If the spatial extent and storativity of the discharging aquifer were known it may be possible to estimate the discharge rate from these recession curves.

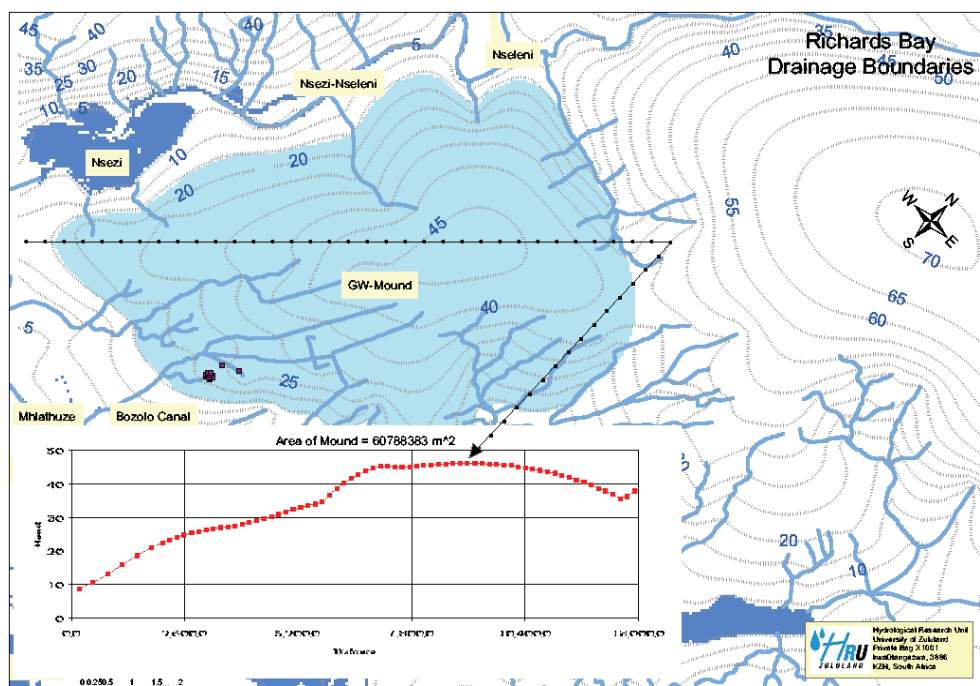


Figure 47: The groundwater mound that contributes to the fluxes into rivers used to estimate the discharge rate in the text

5.1.1 Evaporation

Evaporation and evapotranspiration occur from all the water resources in varying amounts and at different rates. The conceptual model of the evaporative boundary is one where the water vapour is released into the atmosphere at a maximum rate determined by the atmospheric demand. The atmospheric demand is the rate at which the atmosphere is capable of accepting the vapour flux. This rate is determined by the moisture content of the air, its prevailing temperature, the available energy for transformation and the dispersion rates of the atmosphere (Fig. 48).

The evaporation process is considered to utilize water from all available water sources depending on many factors that include the rooting depth and density, the availability of water, the physiology of the plants and length of the pathway from adsorption in the roots to evaporation in the leaves (Fig. 49). The preference for evapotranspiration pathways is unknown. A conceptual model has been presented by Kelbe *et al.* (2002) which is reproduced in Fig. 50. It must be recognized that the atmosphere has a controlling limit on the evaporation rate.

However, the surface water may be evaporated in a different boundary layer to the leaf canopy which draws its supply from the roots.

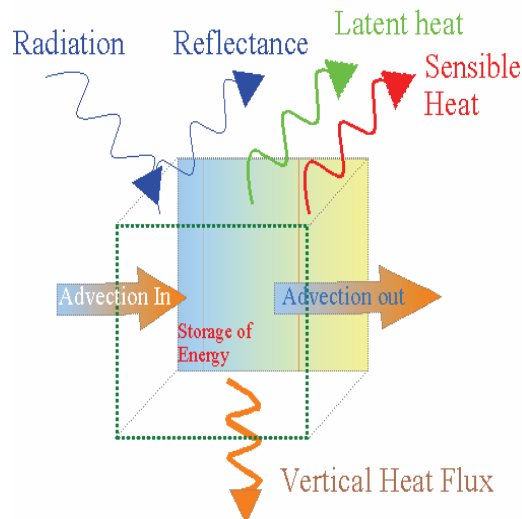


Figure 48: Components of the energy balance driving evaporation

The actual evaporation is also determined by the availability of water and the properties of the evaporating surface. The maximum total evaporation from all resources is determined by the atmospheric demand but the actual evaporation must be partitioned between the various resources to achieve this potential when there is sufficient water for evaporation to occur at the potential rate. There is no known method of determining the preference for partitioning the evaporation rate between these resources.

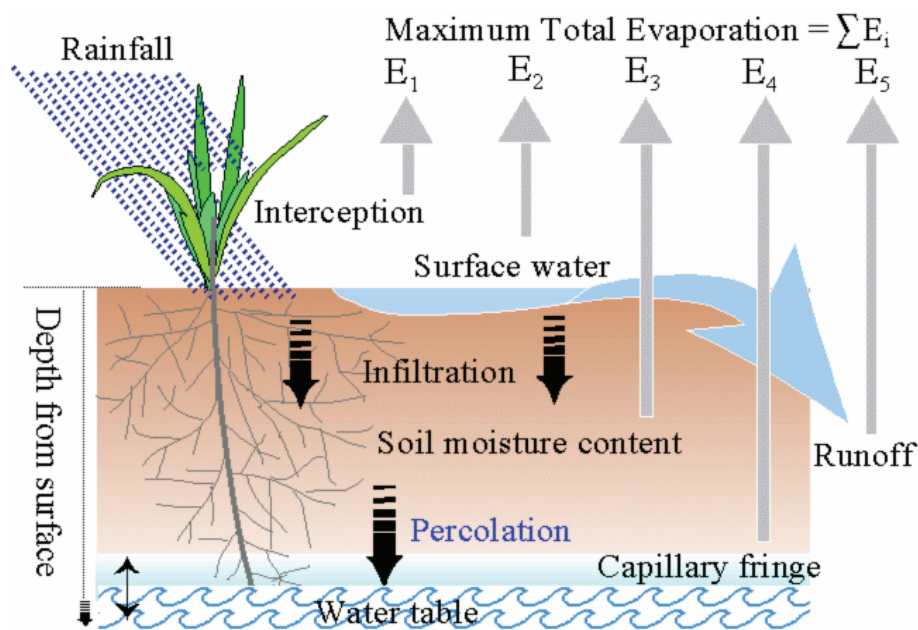


Figure 49: Conceptual model of the evaporation boundaries

5.1.1.1 Observed Impacts

Case Study 3 describes the change in the hydrology of the Siyaya Catchment and Estuary when the land use changed from sugar cane to *Eucalyptus*. This radical change in land use led to a reduction in the baseflow that can be attributed to the degradation of the Siyaya Estuary. The *Eucalyptus* plantations have much deeper rooting systems and higher rates of evapotranspiration that are conceived as the major source of abstraction on the rivers of the Siyaya Estuary leading to the cessation of baseflow. Consequently, any evaluation of the relationship between groundwater and surface water needs to incorporate the necessary evaporative components in the water balance.

5.1.1.2 Evaporation Models

Many models for estimating the actual evaporation have been presented in the literature. These have been summarized by Kelbe *et al.* (2002). However, in groundwater studies, a common method of specifying the recharge is to assign a fixed proportion of the incident rainfall (MAP). Snyman and Kelbe (2006) use effective rainfall which relates to a response rather than to the incident amount. However, the recharge and the evaporation are intrinsically linked and they need to be considered together when evaluating the importance of ground water and its relationship to surface water bodies. Few of these models partition the water loss from the different sources to the total evaporation loads.

The evaporative losses at the groundwater boundary are usually incorporated in water balance studies by adjusting the rainfall to give the NET recharge that incorporates many of the evaporation losses. However, the next section describes methods and models that enable the estimation of the evaporation rate from vegetated surface to be incorporated into groundwater studies. These methods have been used by Kelbe and Rawlins, 1992b) and presented in the different case studies to demonstrate the need to include all the recharge components demonstrated in the previous section to allow evaporation to occur from the groundwater.

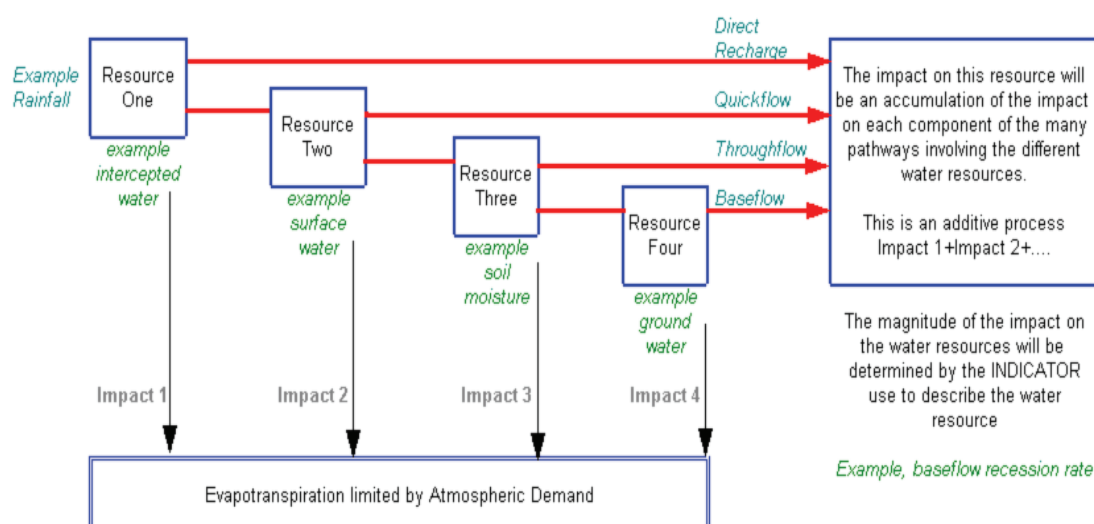


Figure 50: Conceptual sequence of evaporations (Kelbe, Potgieter and Magagula, 2002)

It is difficult to link the evaporation process to any one of the recession profiles observed in the groundwater hydrographs presented. What is vitally important in evaluating the surface water/groundwater interaction is that the recharge and discharge are intrinsically linked. If the surface runoff and evaporation are extracted from the recharge then the conceptual model must exclude these processes from the interaction. Clearly, vegetation can continue to transpire for very long periods of time, in regions with shallow water tables, long after any rainfall has occurred. This is a very complex process and no known studies of the evaporation from the different zones shown have been found in the literature. Conceptual models of the groundwater abstraction through evaporation are presented in the next chapter.

5.1.2 Decomposition of River Runoff and Discharge Estimation

The number of pathway(s) that a volume of water falling on a catchment surface is able to take on its journey to the outlet is almost infinite in number. In the special case where the surface is impermeable, there are a finite number of pathways that are defined by the direction of the steepest slope that directs the parcel to the outlet along easily identifiable routes. This is the concept that Snyman and Kelbe (Case Study 1) have adopted in their model to simulate the flow paths for river runoff components. In most hill slopes, the runoff takes preferential grouping of pathways (see Section 2.9.2 Channel flow) and Fig. 17) that lead to characteristic hydrographs for small headwater catchments. Hydrograph analysis (Fig. 51) suggests that the runoff in a small headwater catchment is made up of several preferential grouping of pathways. The traditional approach to evaluating the groundwater contribution to rivers (Schulz, 1976) has been through hydrograph analysis where the discharge has been equated to the outflow from different storage reservoirs (Fig. 1). This classical approach identified three components of the hydrograph that were traditionally referred to as the baseflow, interflow and quickflow. Many attempts were made to relate these clearly identifiable components in headwater catchments to hydrological processes. Kelbe and Germishuyse (1999c) show the separation of the three components for a small catchment in Zululand. The baseflow component shows the typical oscillations associated with evaporation during the day.

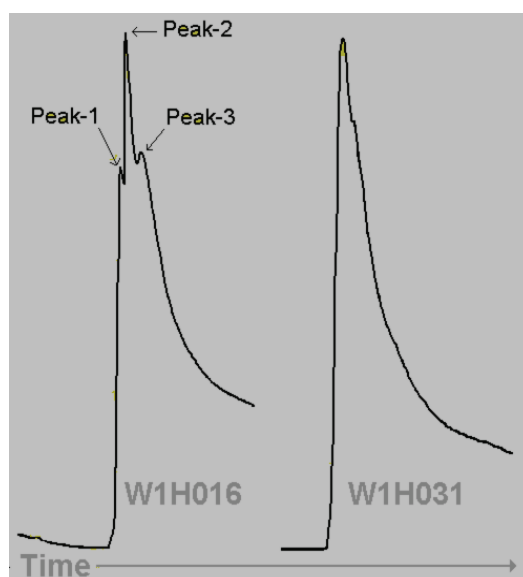


Figure 51: Peak separation in small head water catchments

This hydrograph separation concept is applicable to hill-slope processes where there has been little mixing along the pathways. As the river runoff travels down a channel it gets mixed with flow from other hill slopes and eventually other catchments that may be discharging different flow components into the channel. An examination of the hydrograph from a small catchment in the Ngoye Hills by Mulder and Kelbe (1992a) identified three peaks in flow that were estimated to be derived from three contributing areas along the main river channel. In direct contrast, a similar sized catchment with three converging streams showed no separation in peak flows. The separation of the peaks is conceived to be due to the overlapping or mixing of delayed flow from different contributing sub-catchments. This would imply that the hills-lope mixing becomes evident along stream channels that are in excess of 4 to 5km. Clearly, the mixing process must be complete for large catchments and the application of hydrograph separation techniques would need a different conceptual model. This is discussed in more detail in the model presented in Case Study 1. All reference in this report to this technique implies hill-slope processes where the scale of application is limiting.

Quick-flow (from hill slopes) is that component of the discharge that flows to the catchment outlet the fastest and is generally perceived to be runoff that is associated with (near) surface processes. The interflow is that discharge component that is released much slower than the quickflow but subsides almost as quickly in many areas. Numerous studies have assumed that the baseflow to a stream is equivalent to groundwater discharge and that the discharge can be identified (Hafford and Mayer, 2000). However, they point out that the most basic hydrograph-separation techniques arbitrarily truncate stream-discharge peaks and assume that the remaining discharge is groundwater discharge. Groundwater discharge, on the other hand, is the component of the runoff that persists long after the storm event has subsided. In catchments with large groundwater reserves, the baseflow is known to persist for a long period of time and usually causes the stream to flow for most of the season.

Traditional hydrology considers that perennial streams have large groundwater reserves or the catchment has sufficient rainfall to feed the streams on a regular basis. There are numerous small streams with catchments of a few square kilometres in Maputaland that seldom stop flowing, while there are large river systems, also in Maputaland, that cease to flow on regular occasions.

The baseflow recession rate can be described by various mathematical equations of which the exponential model [$Q_t = Q_0 K r^t$] is the most common (Smakhtin, 2001). However, these techniques are based on the assumptions that periods of exclusively groundwater discharge can be reliably identified and that stream discharge peaks approximate the magnitude and timing of recharge events. Hafford and Mayer (2000) examined these assumptions and estimated that the recession index (K) could not be estimated reliably from stream discharge records alone because the estimates of the recession index were greater than 1 000 d. The ratio of stream discharge during baseflow periods was 2 to 36 times greater than the maximum expected range of groundwater discharge at many sites in the US. They contend that the identification of the groundwater component of stream discharge records was ambiguous because drainage from bank storage, wetlands, surface water bodies, soils and other sources frequently exceeded groundwater and also decreased exponentially during recession periods. Hafford and Mayer (2000) suggest that hydrograph separation techniques, when used alone, are poor tools for estimating groundwater discharge or recharge. However, in the context of this study, where the emphasis is on the interaction between groundwater and surface water components, all saturation zones

are classified as groundwater zones. Consequently, the recession rates (K_r) identified and applied in this project relate to the groundwater seepage.

Automation of hydrograph separation has become a common method of estimating groundwater discharge for pragmatic reasons. These methods generally estimate groundwater discharge and recharge rates that are similar to the manual estimates (Hafford and Mayer, 2000). According to them, the automation process enforces a more consistent approach to identifying groundwater discharge than the manual approaches, but it is still highly subjective and it is affected by the same underlying assumptions.

The impact of abstraction from the hill slope on the river runoff must be considered for each of the flow components. Direct abstraction through pumping from production boreholes will deplete the reservoir and cause sinks in the piezometric surface that will restrict the groundwater flow to the river and hence deplete the baseflow component of the river runoff. It is unlikely to cause a major impact on the quickflow or throughflow components. For other forms of abstraction, such as afforestation and agriculture, the abstraction occurs in unsaturated and saturated (capillary) zones that will deplete the storage and partitioning of the recharge to the groundwater. This will also impact on the baseflow component (see Case Study 2).

5.1.3 Groundwater Seepage into Coastal Lakes

Lakes and wetlands are recharged by rainfall, rivers and groundwater that all vary in response to different driving forces. The discharge from lakes and wetlands can also occur through groundwater, rivers and evaporation. The lakes are often used for water supply and consequently, there is also the need to determine the abstraction for the mass balance. Consequently, the analysis of these resources involves an understanding of the interactions and conditions that prevail during specific conditions.

Lakes have been classified according to their location in a geological setting (Townley *et al.*, 1993). Kelbe and Germishuys (2000b) classified the Maputaland lakes as:

- Coastal lakes with direct connection to the sea through estuaries (Lake St. Lucia, Kosi)
- Coastal lakes with no surface outlet and with a strong groundwater dependency (Lake Sibaya)
- Coastal lakes with outflow and strong groundwater dependency (Mzingazi, Cubhu and Nhlabane)
- Off-channel lakes situated in flood plains (Lake Mangeza, Mpangeni, etc.)
- Inland lakes with strong surface water dependency (Lake Nsezi)

While the level of dependency of these different classes of lakes may differ, the interactive processes remain the same for all classes. Any river runoff into the lake will contain baseflow (groundwater) from the catchment area of the stream/river. Direct seepage to the lake will occur along the shoreline. In all cases, any head difference between the underlying aquifer and stream/lake will conform to Darcian flow through the bed sediments. Consequently, the principal need in evaluating the interaction between lakes/estuaries and the groundwater is in the principal processes and not in their ecological and geological setting. In this study, the lakes can be classified according to the type of interaction with the surrounding aquifer (Townley *et al.*, 1993).

There are many lakes that act as sinks for groundwater drainage. Similarly, there are many lakes that constantly act as source of recharge to aquifers. However, most lakes along the Maputaland coastal plain are likely to act as mixed mode or flow-through lakes where there are areas of seepage into and out of the lake shoreline (Townley *et al.*, 1993). Figure 52 is an example of the shoreline seepage vectors for Lake Cubhu showing the groundwater flow direction. There are significant areas where there is direct seepage into the lake from the northern groundwater mound and the southern coastal dune ridge. However, there is compensation groundwater outflow to the Mhlathuze estuary in the east.

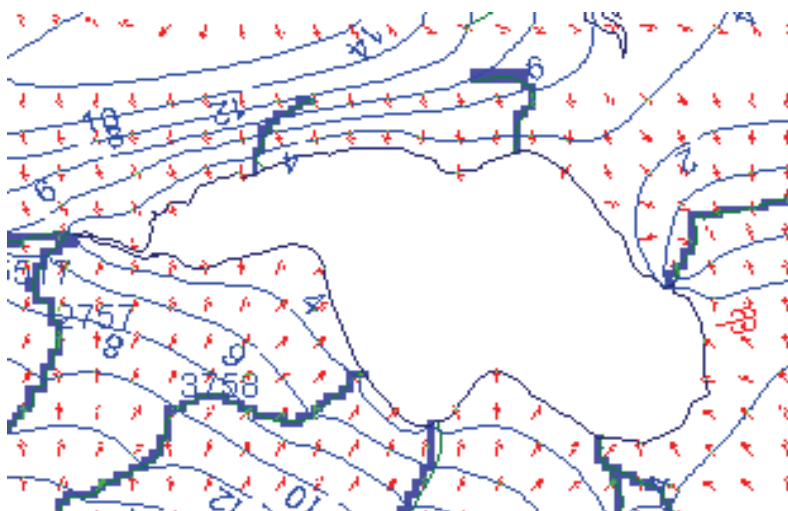


Figure 52: Groundwater velocity vectors showing the zones of inflow and outflow for a coastal lake

The Reserve determination of the coastal lakes in the Mhlathuze catchment is related directly to the variability in the lake level (DWAF, 2000). Similarly, the recent rapid Reserve determination of Lake St. Lucia (Van Niekerk, 2005) is based on the assessment of the major contributors of the water balance that established the Van Niekerk hydrological characteristics conditions that affect the ecological functions of the system. These assessments invariably involve an estimate of the mass balance of the systems. This in turn requires the assessment of the river inflow and outflow, rainfall and evaporation as well as an estimate of the groundwater inflow and outflow. In flow-through lakes, the NET groundwater contribution may be small but the actual inflow and outflow components may be substantial. Consequently, the net mass balance study may not give a true indication of the importance of groundwater in sustaining the water balance of the lake system.

5.1.3.1 Mhlathuze lakes

The mass balance of the coastal lakes in the Mhlathuze system was derived using numerical models by Kelbe and Germishuys (2000a). The contribution to the mass balance by groundwater was derived from numerical models. Figure 53 gives an example of the estimated water balance for Lake Mzingazi (Kelbe and Germishuys, 2000a). In this example the groundwater contribution is the NET balance between inflow and outflow. In order to estimate this it is necessary to develop a comprehensive model of the interaction. This interaction is based on an understanding of the geohydrology and lake bed interface

(sediment) dynamics. However, the net contribution does not take account of the hydrodynamics of the system and the total amount of water flowing in and out of the lake.

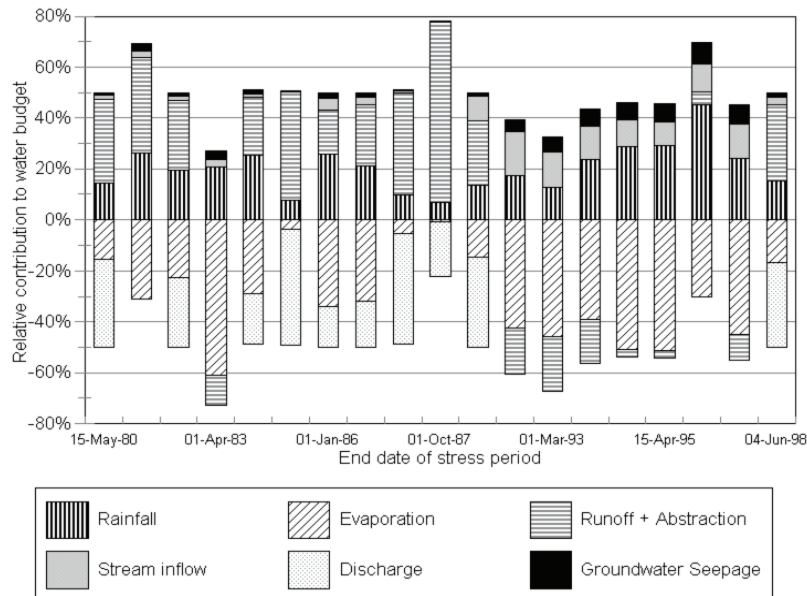


Figure 53: Mass balance estimates for Lake Mzingazi (Kelbe and Germishuyse, 2000)

5.1.3.2 Lake Mzingazi

Studies by Kelbe and Germishuyse (2000b) in Richards Bay have revealed that the groundwater seepage into Lake Mzingazi occurs along the shoreline and through deeper hydrostratigraphic units (Uloa) that account for a large proportion of the lake water budget (Fig. 54).

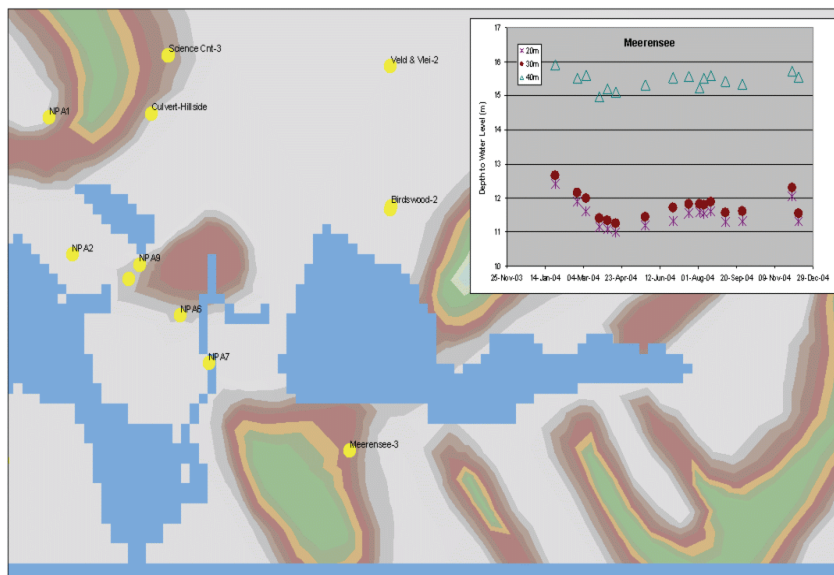


Figure 54: Head variation in the aquifers at Meerensee recharging and draining Lake Mzingazi. The deeper BH is located in the Uloa Formation

A network of monitoring boreholes has been established around Lake Mzingazi. On either side of the Lake (Fig. 61) there are nested boreholes that monitor the different hydrostratigraphic units. There is a difference between the piezometric head for the Uloa and overlying Port Durnford which suggests that the lower HSU is draining faster, creating a lower head (Fig. 54). This would explain the large loss of >20 000 m³/d estimated by Kelbe and Germishuyse (2000b). A similar situation exists at the Science Centre where the head difference is about 1 m. The Uloa formation at the Science Centre also shows a significant increase in TDS (Fig. 55) that is not reflected in the other HSUs suggesting possible links to the surface water resources.

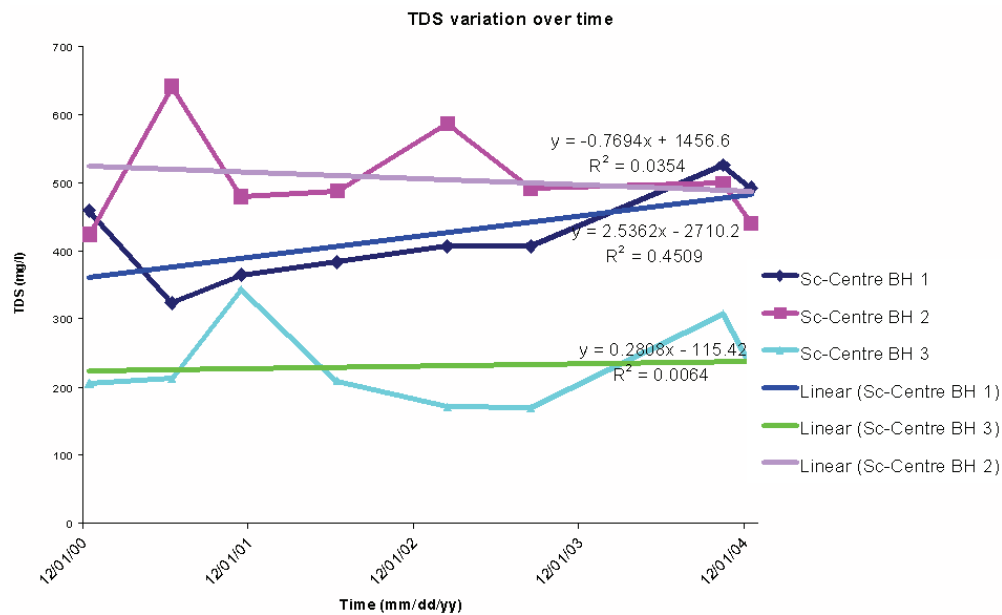


Figure 55: Temporal variation in TDS for three HSUs at the Science Center. The straight lines are linear trends.

The hydraulic and water quality investigation (presented in more detail in the case studies) indicates that the deeper groundwater units in Richards Bay play an important part of the hydrodynamics of the coastal lakes and may be a major source of recharge and discharge for some of the systems.

5.1.3.3 Lake Cubhu

Lake Cubhu is presented as an example of a shallow coastal lake with significant stream inflow flow that has strong groundwater seepage under average conditions. Figure 56 shows the stream network feeding and draining the lake. The inset to Fig. 56 shows the water balance components for the main boundaries. The inflow to the lake from the aquifer from seepage along the shoreline is approximately 14 000 m³/d.

This is only slightly less than the river flow of 20 000 m³/d flowing into the lake (out of the aquifer via the rivers). The simulated evaporation is comparable to the recharge.

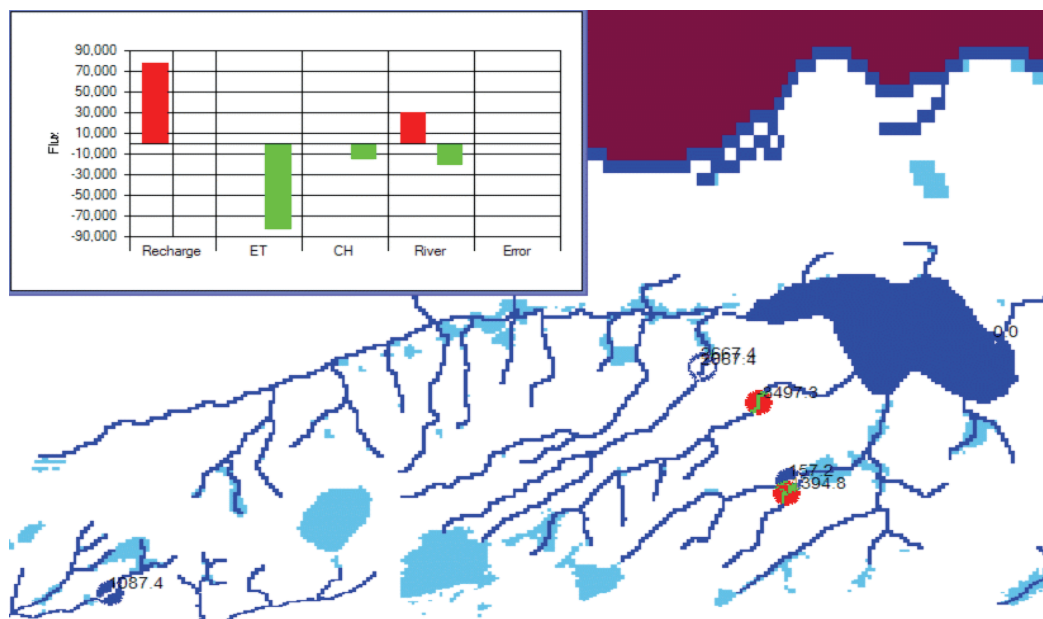


Figure 56: Geohydrological boundaries for Lake Cubhu used to estimate the groundwater contribution to the lake. Inset is the mass balance for the main components of flow in the catchment area of the lake

5.1.3.4 Lake St. Lucia

The importance of considering scale of assessment is presented for Lake St. Lucia by Kelbe *et al.* (2004). This wetland has five major rivers that provide a substantial proportion of the water budget for the lake under normal climatic conditions (Stassen and Mynhart, 1993; Van Niekerk, 2005). Under these conditions, the groundwater seepage to the lake (Kelbe and Rawlins, 1992b) generally provides a much smaller proportion (<10%) of the total water budget (Table 13). However, the rainfall and river runoff show very large variability but the groundwater is very persistent. Consequently, during severed drought conditions, the river and rainfall contributions may be completely depleted, as happened in 2005, and then the groundwater becomes the only source (100%) of freshwater inflow to the wetland. This small but very dependable inflow has been identified as crucial in the ecological resilience of Lake St. Lucia (Kelbe *et al.*, 2005; Taylor *et al.*, 2006).

The storage of water in the lake (water levels) is controlled by the balance between the freshwater inflow from the river runoff, groundwater seepage and direct rainfall, the inflow/outflow of sea water through the estuary and the direct losses through evapotranspiration. The annual average rainfall, runoff and groundwater seepage into the lake are given in Table 12 together with the estimated evapotranspirational losses. The water balance indicates a freshwater outflow to the sea of approximately $256 \times 10^6 \text{ m}^3$ during average climatic conditions. This outward flow of freshwater ceases when the combined rainfall-runoff contribution reaches about $465 \times 10^6 \text{ m}^3/\text{yr}$. The variability in freshwater inflow for the major rivers and groundwater is shown in Fig. 57.

Table 12: Water balance components of Lake St. Lucia (*10⁶ m³/yr)

Contributions to lake water balance	Ave
River runoff (Cornelius, 1992)	362
Groundwater (Kelbe <i>et al.</i> , 1995)	53
Rainfall (1 000mm/yr)	360
Evaporation (4mm/d)	518
Inflow/Outflow to the sea	256

The lake water balance and mouth conditions create a wetland that experiences extreme fluctuations in salinity levels that have a profound impact on the ecological function of the system. During extended drought periods when the estuary is connected to the Indian Ocean, the low inflow of saline water that created very high salinities flux of freshwater and high evaporation lead to a large inflow of saline water that created very high salinities (Fig. 58). During these periods, the freshwater species of fauna and flora are under severe stress. Taylor *et al.* (2006) suggest that there the groundwater seepage zones (see Fig. 65) along the groundwater seepage can be observed along much of the shoreline in support of the refugia theory (Fig. 59). Rainaldi (2005) has surveyed the water table elevation along the shoreline shown in Fig. 60 to estimate the groundwater seepage rate.

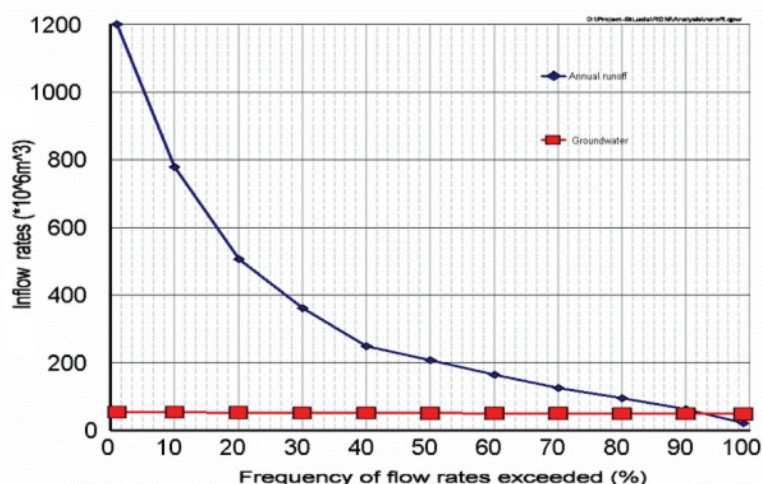


Figure 57: The relative frequency of freshwater inflow to Lake St Lucia from five major rivers (blue line) and the simulated groundwater seepage (red line)

Taylor *et al.* (2006) have identified many of the groundwater diffuse seepage zones and creeks along the eastern shore line of Lake St. Lucia that form the important groundwater-dependent refugia sites (Fig. 61). When the lake is full of freshwater these groundwater seeps may have little importance. However, under the frequent high salinity states (Fig. 62), they become increasingly important as the state persists. The shoreline studies presented are part of a project to map the resistivity and salinity profile of the soils of the exposed beach to examine the relationship between freshwater seepage and plant growth that is in progress.

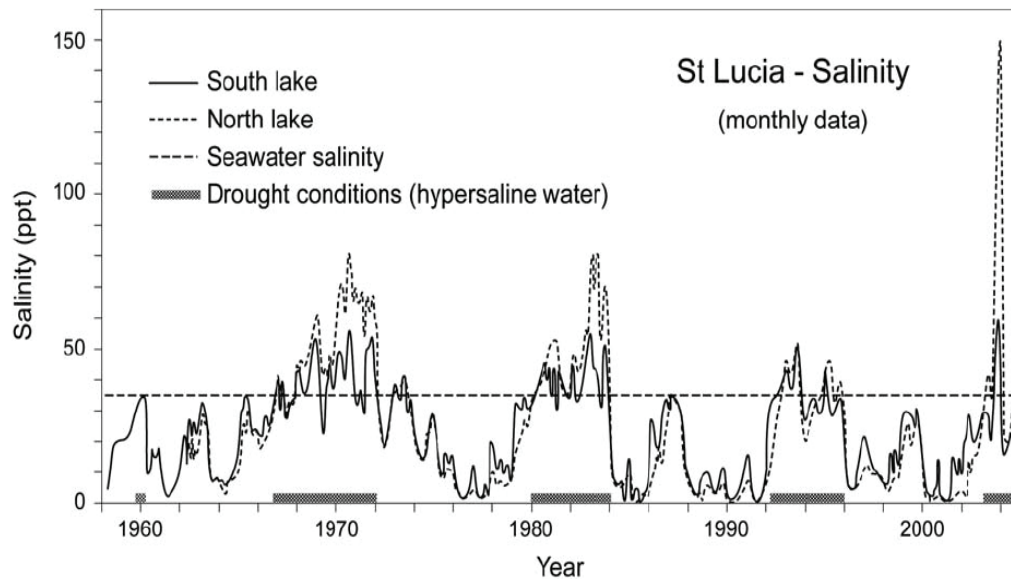


Figure 58: Fluctuation in salinity levels for two locations in Lake St Lucia (from Taylor et al., 2006)

The groundwater seepage into Lake St. Lucia from the eastern shores emanates from a large groundwater mound (Embomveni) or the extended groundwater ridge that provides very consistent flow (see Figs. 63-64). The consistency can be gauged from the observed water level fluctuations on the Embomveni Mound shown. Over an extended observation period stretching back to 1970, the water level in the mound has seldom varied by more than 3 m. A more detailed analysis of these observations has currently been undertaken by Vaeret *et al.* (2009).

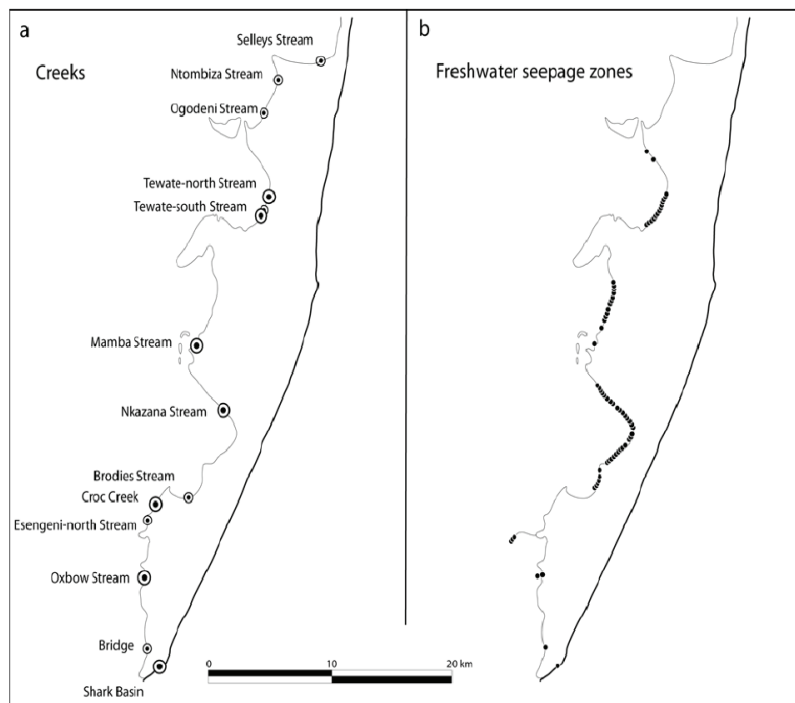


Figure 59: Groundwater dependent refugia sites along the eastern shoreline of Lake St. Lucia (from Taylor et al., 2005)

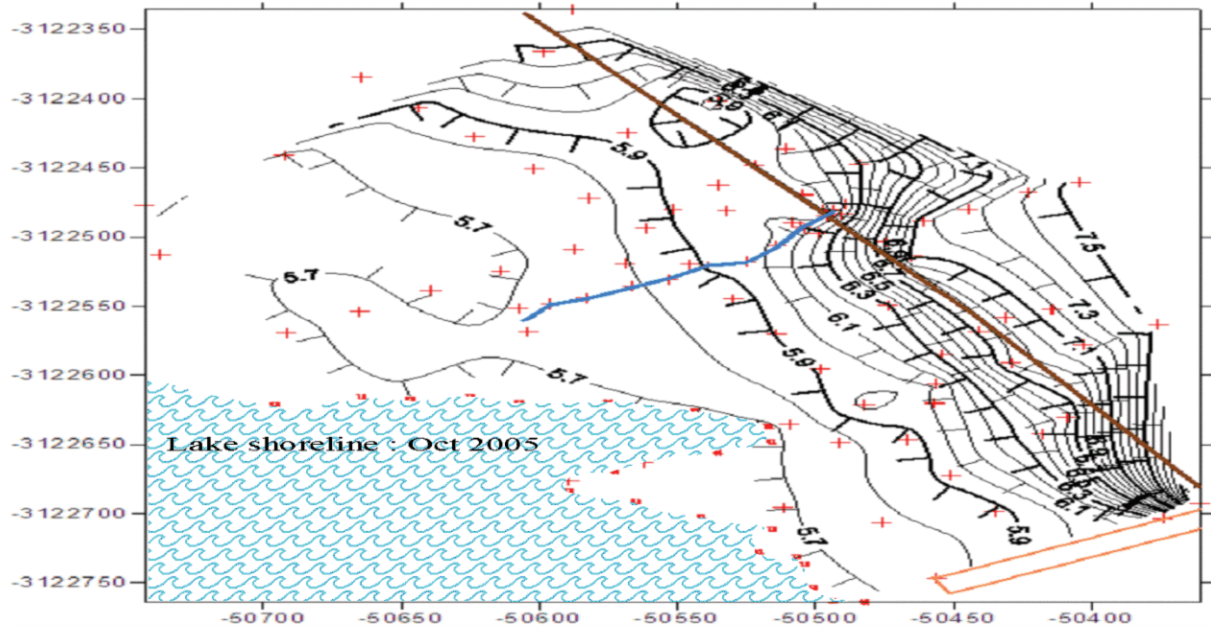


Figure 60: The groundwater elevation contours along the shoreline of Lake St. Lucia (Rainaldi, 2005)

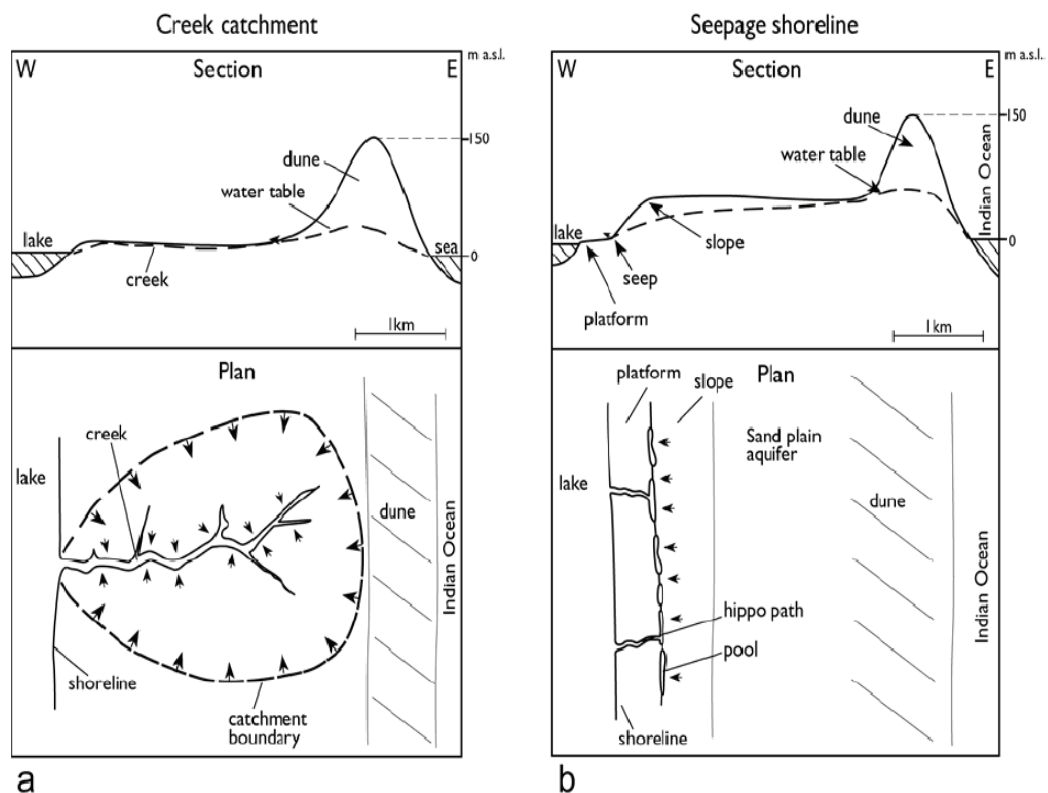


Figure 61: Groundwater seepage zones along the shoreline of Lake St Lucia during the severe drought of 2002-2004



Figure 62: Schematic representation of the groundwater-dependent refugia features located at the sites in Fig. 59 (after Taylor et al., 2005)

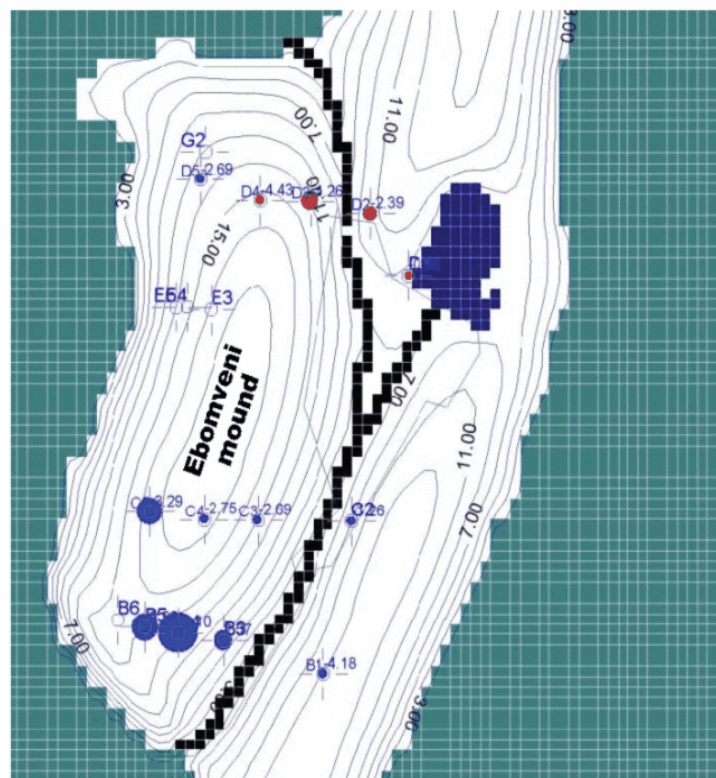


Figure 63: Simulated water table elevations and monitoring boreholes sustaining the groundwater seepage in the refugia sites

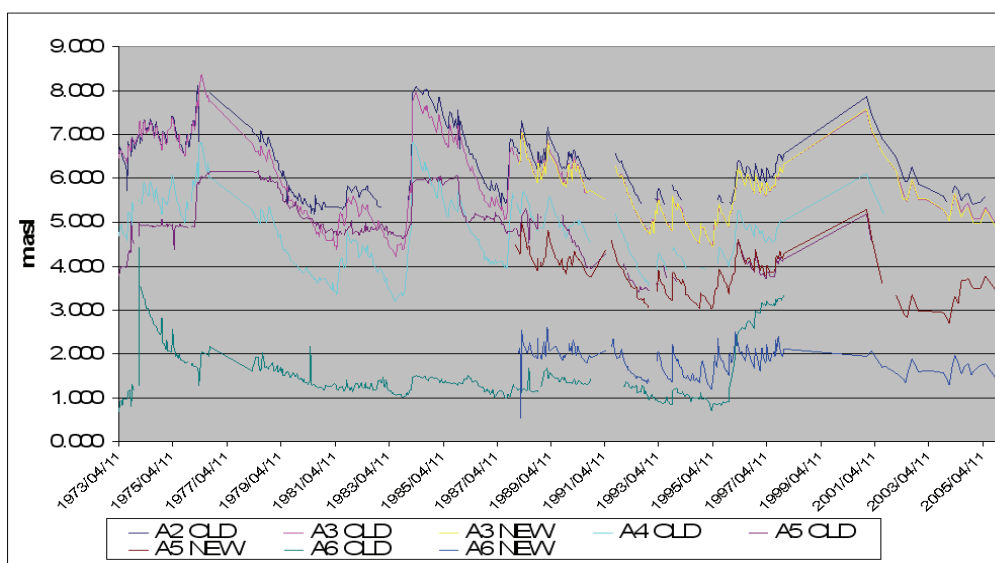


Figure 64: The measured water level elevation for the boreholes shown

5.1.4 Estuaries

The groundwater contribution to estuaries occurs as seepage along the shoreline (Fig. 65) and the baseflow component of river runoff. While the baseflow may be a very large proportion of the inflow into estuaries, the shoreline seepage is unlikely to have a large impact on the mass balance of estuaries, particularly if the estuary has direct and continuous linkage to the sea. However, the groundwater seepage may create refugia sites such as those identified in Lake St. Lucia that may play an important role in the ecological function of the system. If the linkage to the sea is broken by a closed mouth, as often happens in Maputaland, the estuary water level may rise significantly to the point where it causes the estuary to flow into the aquifer.

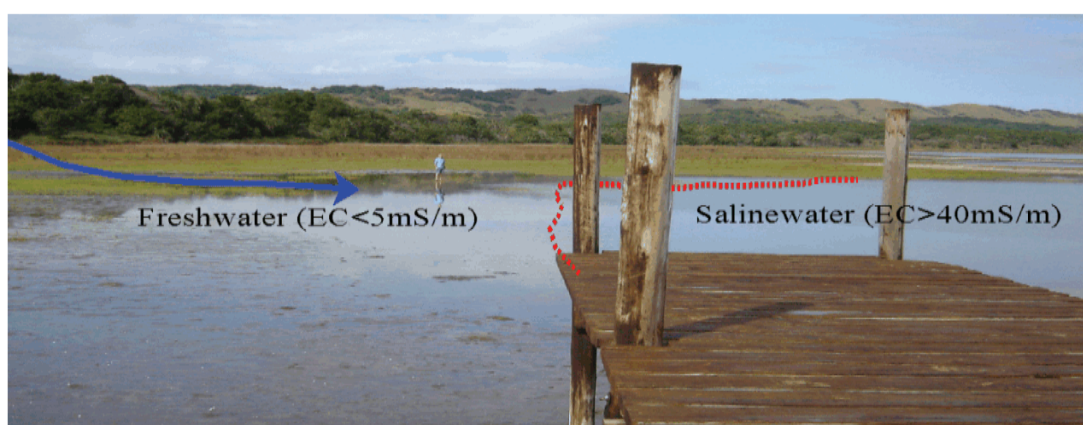


Figure 65: Photograph showing where the interface between the freshwater zone ($EC < 5mS/m$) and the saline lake water zone ($EC > 40mS/m$) was measured

Many of the estuaries along the Maputaland coast close on a regular basis due to climatic conditions and sediment deposition along the coast. Following normal or above normal rainfall, these estuaries often cause flooding in the adjacent wetlands. In many instances, these flood plains have large commercial farming enterprises that are affected by the flooding and request that the mouth be artificially opened to protect their crops. However, it is estimated that the Lake St. Lucia system would rise to an elevation in excess of 3 m before the system would breach naturally. This would induce substantial recharge to large areas of the surrounding aquifers.

5.1.4.1 Umlalazi Estuary

The Umlalazi Estuary (Fig. 66) is prone to frequent mouth closure and subsequent flooding of the adjacent flood plain. It has been necessary on occasions for the authorities (KZN Wildlife) to open the mouth to restrict the flooding of commercial cane lands on the low-lying flood plains. The flooding is also known to kill mangrove if it persists for too long. These flooding events will recharge the aquifers. This flooding is a natural phenomenon due to mouth closure when there is a build-up of river runoff and not from saline water. However, the estuary is frequently in a saline state when the mouth is open and there is little or no river runoff.



Figure 66: Photograph of the flooding of the Umlalazi Estuary flood plain.

5.1.4.2 Siyaya Estuary

The Siyaya Estuary links the twin streams of Manzamnyama and Siyaya to the Indian Ocean. This estuary, similar to many other small, narrow and shallow outlets to the sea (Nyoni on the Matikulu Estuary), experience frequent mouth closure common to the east coast of Africa because of the small catchment area of the feeder streams. However, excessive sediment production from sugar farming in the feeder catchments caused excessive sediment loads that have clogged the estuary. A subsequent change in land use to *Eucalyptus* plantation has completely dried up the rivers, including the baseflow, so streamflow necessary to scour the estuary no longer occurs, except during extreme storm events. With

little or no river runoff from the streams, the estuary is maintained entirely by direct rainfall and groundwater seepage from the dunes on either side of the system. The silted estuary has now developed a complete cover of papyrus reeds that is capturing any further stormflow and Aeolian sediments and is expected to rise slowly to above sea level.

5.1.4.3 St. Lucia Estuary

The St. Lucia Estuary links the largest wetland in Africa to the Indian Ocean (Fig. 67). It has been ranked with the highest priority in terms of importance (Turpie *et al.*, 2002). A recent Reserve determination identified the linkage to the Umfolozi River and the mouth closure as the two most important factors affecting the estuary. However, it was also recognized that the groundwater contribution to the system was a small but extremely important component to the ecological function of the system (Van Niekerk, 2005 and Kelbe *et al.*, 2005). Previous efforts to connect the Umfolozi to the estuary via link canal (see Fig. 67) failed when extreme flooding caused damage to the system and the flooding remove the immediate threat to the lake. Efforts to revive the canal are being pursued during the current drought conditions.

The mass balance model of the St. Lucia Estuary shows that the groundwater seepage into the system during the current drought moderates the change in salinity by up to 70% for extreme conditions. The simulated salinities are slightly higher than the observed values although the water levels are correct following the large influx from the Umfolozi River (Figs. 68 and 69). This still needs further investigation. The impact of forestry on the groundwater has been well established (Lindlet and Scott, 1987 and Rawlins, 1991). Deforestation of selected catchments around the estuary is being planned to improve the groundwater flow into the estuary (Taylor, 2005). This is unlikely to have any significant impact under normal climatic conditions but it would have alleviated the stress during the current severe drought conditions with mouth closure.

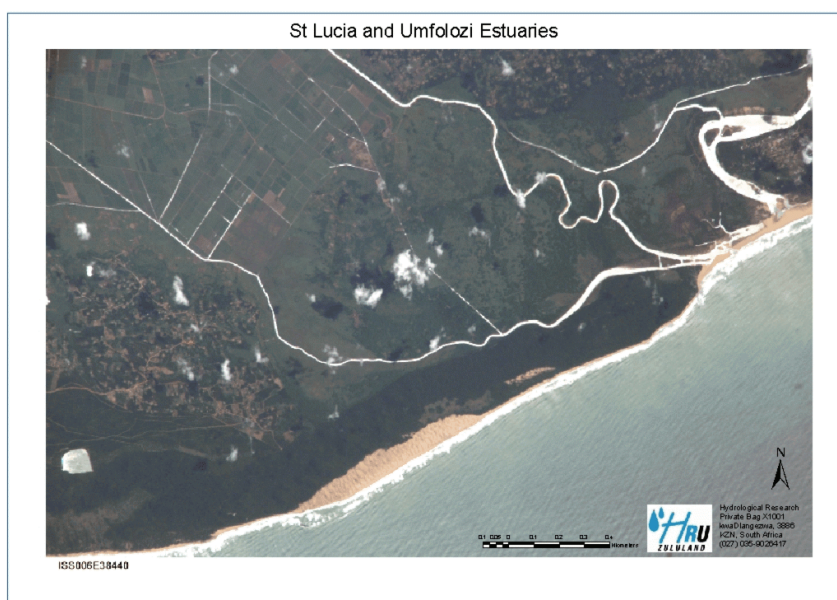


Figure 67: Satellite image (from NASA) of the St Lucia and Umfolozi Estuaries and in relation to commercial farming.

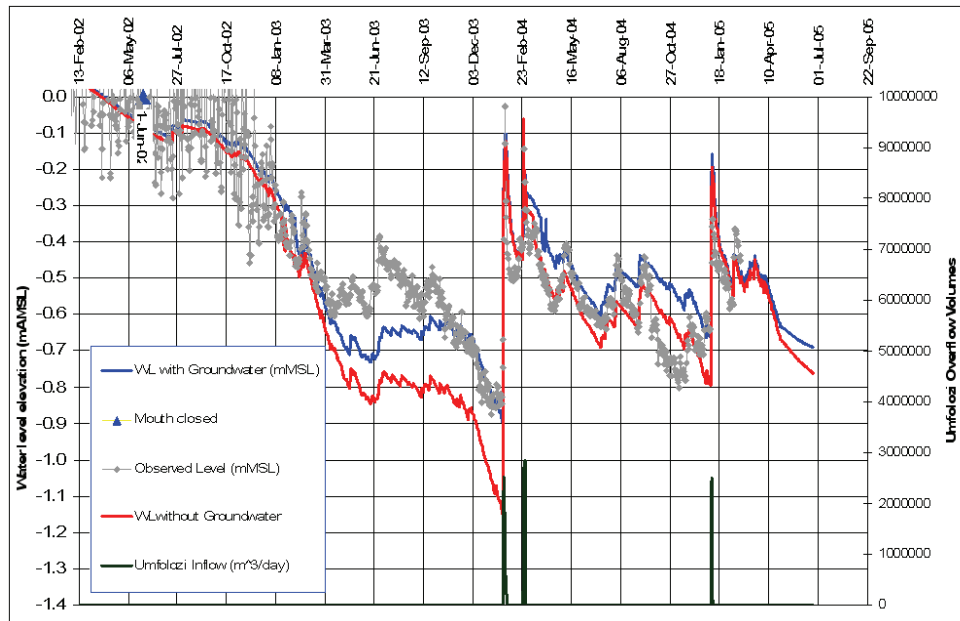


Figure 68: Simulated water levels for the estuary (Narrows) following mouth closure with (blue line) and without (red line) groundwater contribution.

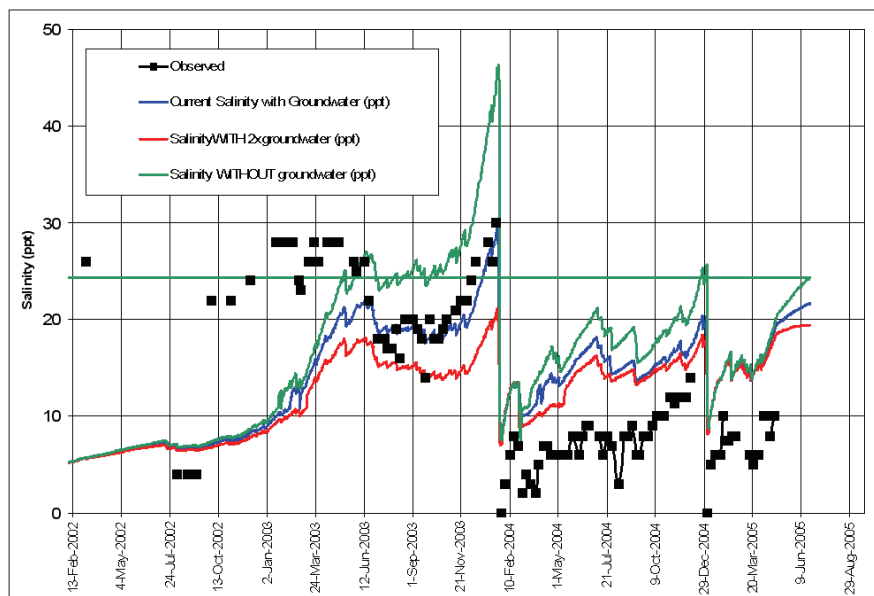


Figure 69: The simulated impact of groundwater on the salinity profile of the St Lucia Narrows. The blue line represents the conditions prevailing since mouth closure in Jun 2002. The red line represents the impact of doubling in the groundwater discharge while the black line indicates what would have happened without any groundwater.

5.1.5 Wetlands

The definition of wetlands is related to the purpose of use. The Ramsar Classification System recognises all forms of water bodies. It is intended to provide only a very broad framework to aid rapid

identification of the main wetland habitats represented at each site for assessing their global importance for the conservation and protection of aquatic birds. However, this categorization is too broad and includes many features that are not related (directly or indirectly) to the groundwater. According to Cowardin *et al.* (2003) the term wetlands is used to describe areas which are neither fully terrestrial nor fully aquatic. These areas range in character from the vast Mkuze swamps in Maputaland to shallow unimpressive depressions which have water for a small period of the year. All these wetlands have important roles, providing a diversity of critical habitats for residential and migratory waterfowl and other animals, controlling floods and sediment transport, and water and nourishment for subsistence living by local impoverished communities (Fig. 70). The Mkuze swamps are essential for protecting Lake St. Lucia from sediment during flooding while they and other wetlands around the periphery of the lake provide essential groundwater-dependent habitats during the extreme droughts when the lake is either too saline or too dry to support the animals and birds.

For the purpose of this study, the wetlands have been separated into categories that describe their dependence on groundwater, i.e. - marine and other large saline resources with little or no concentrated groundwater seepage zones; wetlands with direct connectivity to the aquifer and no overriding surface water influx; and wetlands with significant surface inflow and groundwater contributions along the margins of the wetland. The Maputaland coastal plain is home to all these forms of wetlands.

No studies were conducted on the groundwater contributions to the marine wetlands. However, Kelbe *et al.* (1995) have estimated that the groundwater flux into the marine environment along the Indian Ocean shoreline for the eastern shores of Lake St. Lucia (Fig. 71) is of the order of $2\,500\text{ m}^3/\text{d}\cdot\text{km}$. This is substantially greater than the flux into the Lake from the same source ($600\text{ m}^3/\text{d}\cdot\text{km}$). It is unclear what the impact is of the freshwater on the marine environment – the authors recall hearing tales of groundwater springs near the coral reefs of Sodwana but these have not been confirmed.



Figure 70: Typical inter-dunal wetlands on the coastal plain. Inset: A fish trap used by local communities for subsistence

All the commercial forests on the eastern shores of St. Lucia have been removed and the land use has reverted to nature conservation. After the trees had been removed between 2003 and 2005, many remnant tree stubs were completely covered by up to 0.5m of water in remerging wetlands, highlighting the significant impact of the forestry on the groundwater and associated wetlands. As a consequence of this, KZN Wildlife has adopted a management plan to limit the extent of natural forests and bush-lands in

an attempt to protect the groundwater and the wetlands (Taylor, 2005). See Fig. 72 for photo comparisons of wetland areas under different vegetative covers.

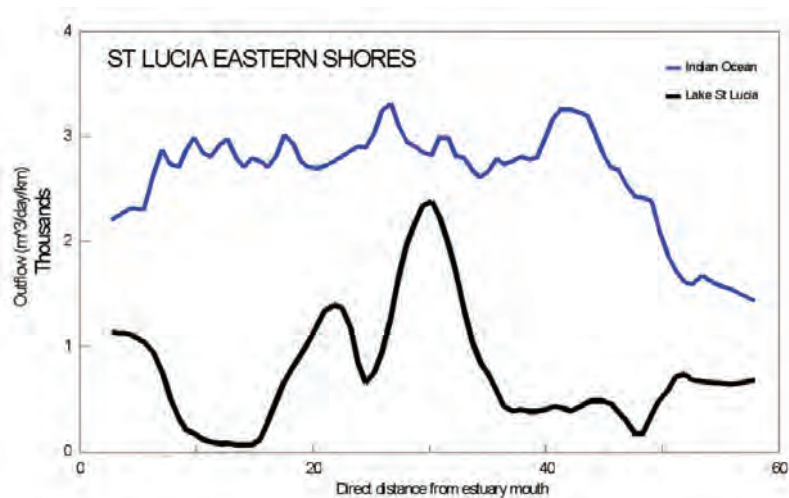


Figure 71: Simulated groundwater flow from the eastern shores

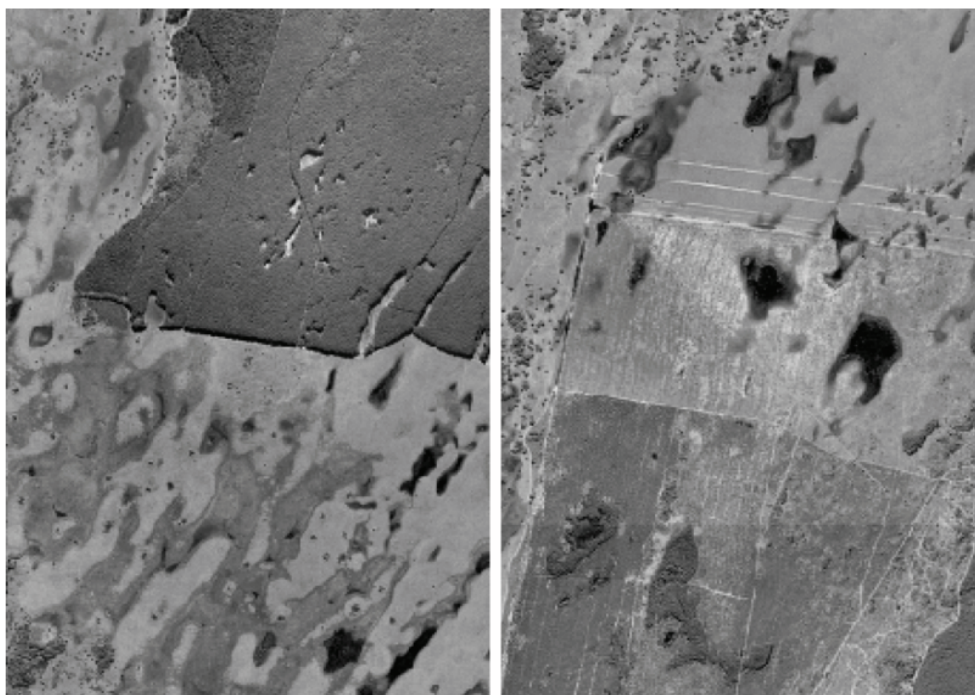


Figure 72: Left photo gives a comparison between the wetlands under natural veld conditions (bottom of photo) and under established afforestation (top of photo). Right photo shows the little difference in wetlands for natural (top) and under recently planted forests (bottom) (Photos from DS&M)

The large Mfabeni swamp wetland on the eastern shores of Lake St. Lucia has no feeder rivers or streams (Fig. 73). It drains south into Lake St. Lucia via the Nkazana stream which has never been known to stop flowing. The swamp either feeds OR drains Lake Bangazi depending on the elevation of Lake

Bangazi. This wetland comprises a very dense land cover of reeds and swamp forest species that restrict the drainage rate and thereby maintains the slowly draining Nkazana Stream.

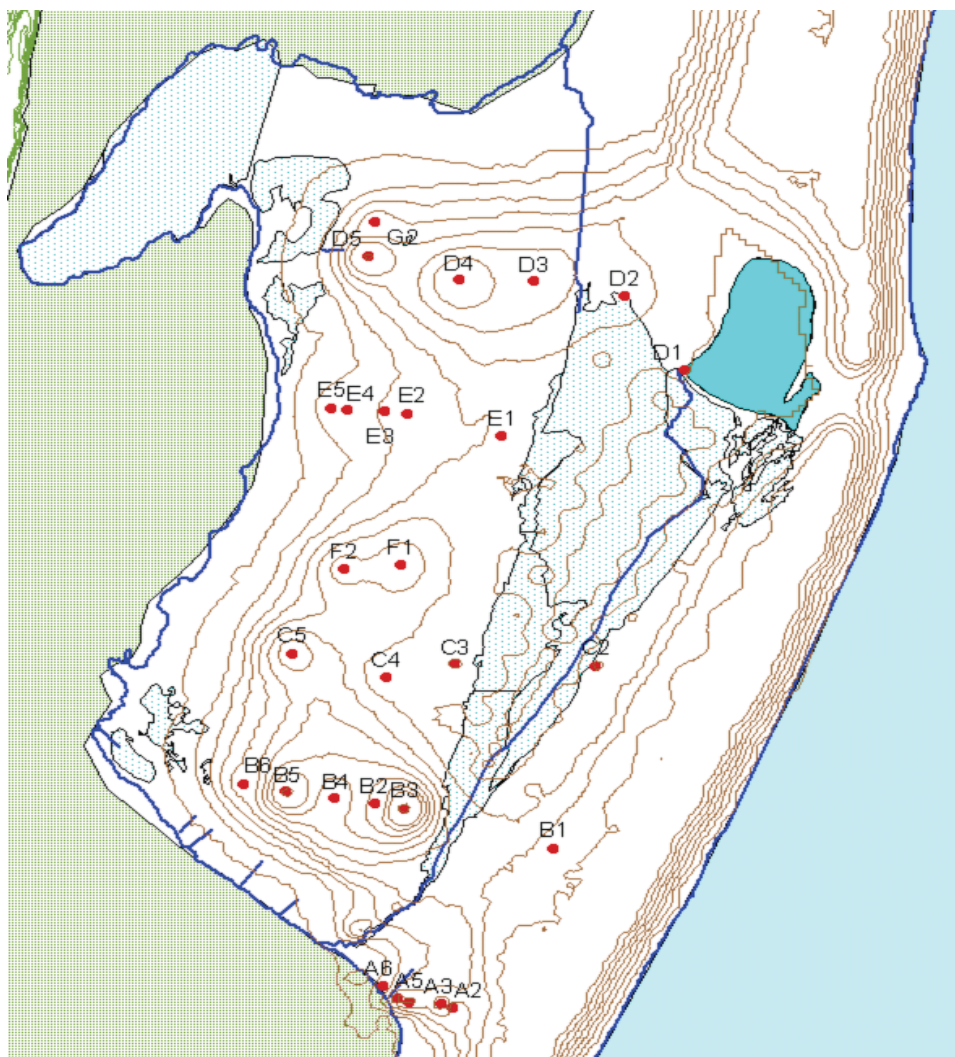


Figure 73: Map of the Mfabeni swamp, Lake Bangazi and the Nkazana stream that form the main depression in the groundwater between the Ocean and Lake St. Lucia (Wedjen, 2000)

5.1.6 Comparative Studies

The water-table elevation response to storm events and river runoff shows similar hydrograph recession characteristics. However, they differ in their time to peak response. Figure 74 shows the comparison between the groundwater and river runoff hydrographs following a short duration, high intensity storm in Alton, Richards Bay. The stream shows an initial peak that is associated with overland flow from a small paved industrial section Richards Bay which is channelled directly to the monitoring station. After a short lull of ~4 h the river hydrograph (red line in Fig. 74) shows the more traditional storm hydrograph with three recession rates for quickflow, throughflow and baseflow. Water quality trends in the streamflow show supporting evidence of these components (black line in Fig. 74).

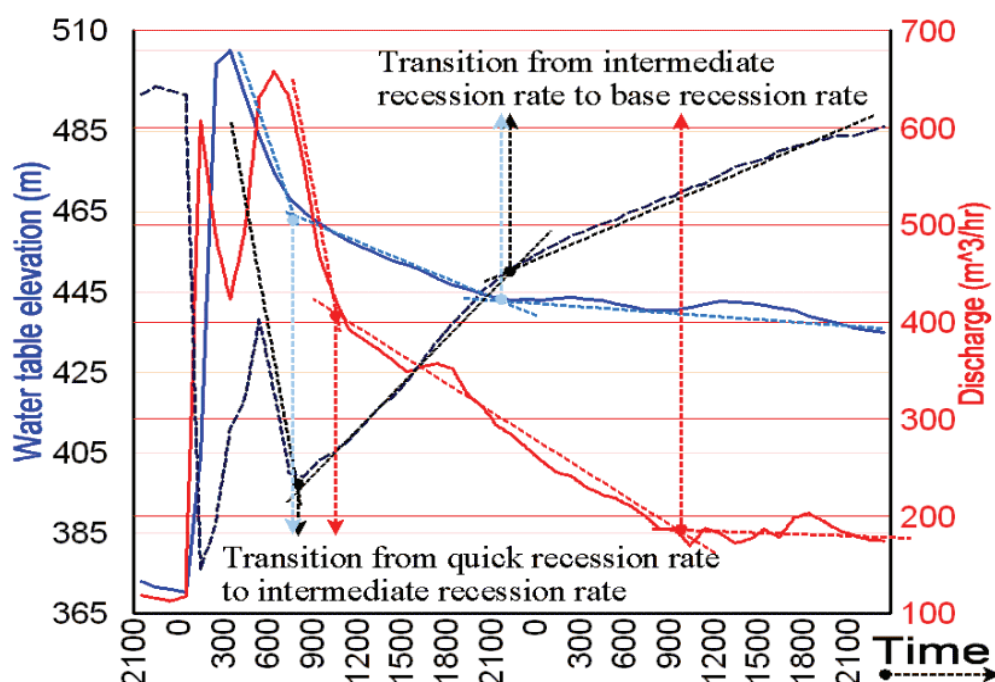


Figure 74: Groundwater elevation (blue line), streamflow (red line) and electrical conductivity (black line) hydrographs in Richards Bay showing comparative transitions between hydrological states. See case study 5 for more details

The groundwater hydrograph (blue line in Fig. 74) for a borehole about 500 m from the stream also shows evidence of similar trends where the peak head is midway between the overland flow and quick flow peaks in the river runoff. The groundwater recession curve also shows three recession rates that have inflection points that are close in time to those observed in the river hydrographs (Fig. 74).

The hydrographs in Fig. 74 indicate that the streamflow TDS (represented by electrical conductivity) is the inverse of the stream discharge rate during the overland flow conditions indicating new water. However, the second minimum in TDS occurs 2 to 3 h after the minimum in stream discharge. The first rise in TDS occurs during the time over which the water table reaches its peak at about 4.3 h indicating an increase in old water discharge. The TDS shows a second minimum at almost the same time as the second discharge peak at 09:00 when quickflow passes through the monitoring point. This inflection point corresponds closely with the change in recession rate of the groundwater hydrograph at about 09:00 when quickflow peaked. The change in the recession rate for the river runoff at 12:00 does not correspond with any significant feature in the other series. However, it does occur about 2 to 3 h after the change in water table recession rate and may be associated with delayed response to a groundwater discharge component. The inflection points in the TDS correspond closely with the groundwater recession change at 24:00, well before the river runoff changes entirely to groundwater.

The peak in groundwater elevation lies midway between the overland flow and quickflow components. These series suggest that the groundwater peak elevation in the shallow primary aquifer in

Richards Bay is approximately 1 to 2 h before the quickflow reaches its peak. Similarly, the stormflow in the river reverts completely to baseflow about 2 to 3 h before the groundwater reaches its long-term recession rate.

It is difficult to associate the peak in groundwater with the peak in runoff using traditional hydrological principles. However, the concept of a pressure wave (wave velocity) may provide a mechanism for pushing the 'old' water (with high TDS) into the stream during quickflow conditions.

5.1.7 Recession Limb (Discharge)

Using the same conceptual drainage model as river runoff described by Schulz (1976) and described in previous sections, the groundwater hydrograph recession limb for the storm event in Fig. 74 shows three characteristic features that start with a high recession rate for a short period of time followed immediately after the peak in piezometric head (Fig. 74). This is followed by a slower rate that eventually changes to the more persistent slow 'base' rate.

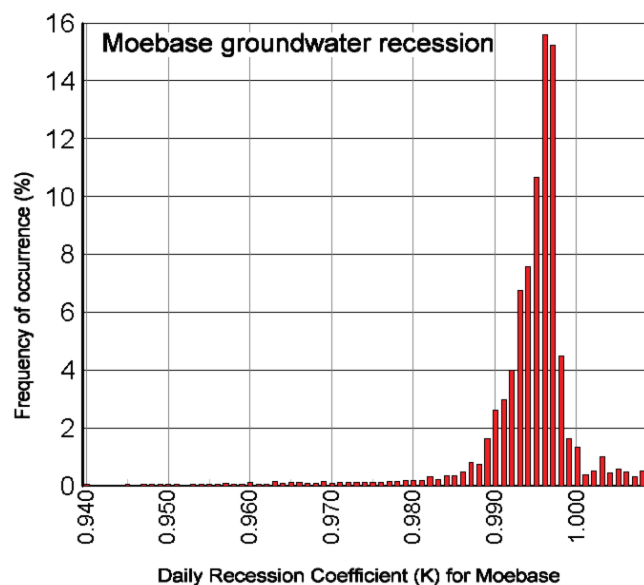


Figure 75: The frequency distribution of daily recession coefficients for a shallow aquifer

The constant recession rate model of Schulz (1976) assumes a draining reservoir where the flow can be estimated on a daily series by the ratio: $\text{Head}_i / \text{Head}_{(i-1)}$. This relationship is a function of the reservoir (aquifer) properties in a similar manner to the stream discharge hydrograph in a stream. The relationship between the change in head and discharge rate needs to be established for each aquifer. An example of the frequency distribution of the hourly recession rate ($\text{Head}_i / \text{Head}_{(i-1)}$) is shown in Fig. 75. The mode in the frequency represents the 'base' recession rate which is approximately 0.996 for this example. This is comparable to river runoff baseflow recession rates presented by Snyman and Kelbe in Case Study 1.

The recession rate for the three boreholes in Richards Bay for a period with little rainfall is presented in Fig. 76. There is a difference between these aquifers that reflects the lack of connectivity between the units.

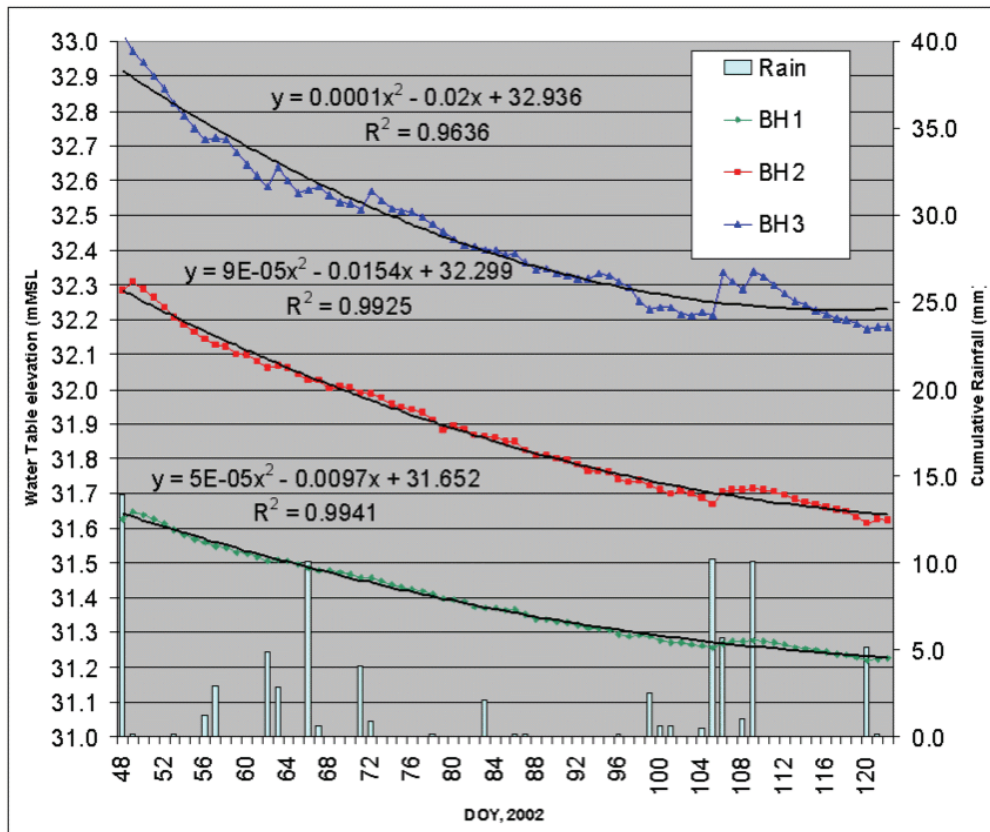


Figure 76: The recession rates for three piezometers at Alton, Richards Bay

CHAPTER 6: MEASUREMENT AND MODELLING OF GROUNDWATER/ SURFACE WATER INTERACTION

The concepts used to evaluate the groundwater interaction with the surface water bodies was described in the previous chapter. This section will look at the details of the measurement and modelling of the system for quantifying the level of interaction.

In the assessment of the interaction between groundwater and surface water bodies, the direct measurement of fluxes is an ideal approach but is difficult to conduct in most setting because of the many pathways that water can follow in the interaction and the nature of the interface/boundary. Consequently, it is often more pragmatic to follow a modelling approach. However, the application of models requires sufficient measurement for calibration of the many parameters used to describe the system and its interactions.

Measurements of aquifer properties are invariably restricted to only a small portion of the system and are inherently prone to representation errors. An isolated borehole may strike an insignificant geological feature or miss a very important one when used to map hydrogeological units. Rain gauges are usually unrepresentative of convective rainfall (Kelbe, 1987) while water quality measurements may be critically dependent on the sampling strategy (Kelbe *et al.*, 1991). However, some measurements that involve integration over space and time may be more representative of the process. For example, river runoff measurements represent integration of the rainfall over a much larger area than a rain gauge but suffer from unknown/unspecified losses in the process (evaporation and percolation). On the other hand, models are representations of the system that are designed to specifically mimic selected features. They are all parametric and require some level of calibration. They are created on the basis of a conceptual understanding of the system (conceptual model) and can take many forms. These can be divided into specific categories:

- Physical models; these are representations of a system using different physical forms that include:
 - *Maps*
 - *Tanks*
 - *Electrical circuits*
- Mathematical models; these have two basic forms that are in common use:
 - *Stochastic models where the outcome always varies but conforms to some specific feature;*
 - *Bayesian models*
 - *Monte Carlo models*
- Deterministic models where the outcome is repeatable for specific conditions:
 - *Analytical models*
 - *Numerical solutions*
 - *1, 2 & 3 Dimensional models*
 - *Finite difference representations*
 - *Finite element representations*
 - *Steady state models represent equilibrium conditions for the system*

- *Transient models include temporal variability*
- In support of these models there are also many methods, techniques and tools that have been used successfully to investigate, quantify and describe the interaction between groundwater and surface water bodies. These include:
 - *Geographic information systems (GISs) with their associated extensions. These systems have a lot in common with the newer versions of the numerical models and there is a strong move toward their merger.*
 - *Spreadsheets. Anderson and Woessner (1992) describe the application of the groundwater hydraulics (Boussinesq equation) in finite difference form using spreadsheets developed by Olsthoorn (1985) and Ousey (1986).*

This section describes some of the techniques and models that have been tested and applied by participants of this project in the Maputaland region. It is by no means a comprehensive review of all models.

6.1 Direct Measurements of (Mass Balance) Seepage Fluxes

The fluxes that are involved in the interaction between groundwater and other aquatic systems include rainfall (or the recharge component), evaporation, and seepage fluxes within and between resources.

- **Measurement of rainfall** is conducted on a regular basis in most areas of South Africa although the density of gauging stations does vary considerably between regions. Rainfall is the main source of recharge and consequently it has been used as a direct index of recharge (Meyer and Godfrey (1995), Worthington (1978), Meyer and Kruger, 1987, and Bredenkamp et al. (1995). Rain-gauge networks were the only source of measurements in the past but radar and satellite systems have become common methods of rainfall assessment, particularly in remote areas (ESA, NASA web sites). In remote parts of Mozambique, NASA rainfall generated series were used to estimate the recharge by Kelbe (2002) when there were no available rain-gauge records except for a short period when AWS data was available. The correlation was not good when compared to the point source measurement but it was assumed to be better than no data. With increasing spatial and temporal resolution in remote sensing, the satellite estimates are improving rapidly (Terblanche, 2007).
- **Measurement of recharge** to an aquifer from any source is seldom done in any region. It is generally more pragmatic to develop a model to link the recharge to the rainfall. Bredenkamp et al. (1995) and Meyer and Godfrey (1995) used chloride measurements of rainfall and groundwater to estimate recharge. Rawlins and Kelbe and (1998) developed a simple model that was based on the direct measurement of rainfall events and the groundwater response to these events for the Eastern-shores of Lake St. Lucia (Fig. 77). Several examples from other location have been presented in previous chapters. References to other models have been presented earlier in the report. The methods that were used in these projects were based on the direct relationship between rainfall events and the groundwater response that has been described in previous

sections. However, in all cases where models were used for groundwater predictions and assessment, the recharge was estimated using the rainfall and then calibrated against streamflow as described in the next section on modelling. This project has established and used several monitoring stations in Southern Africa to measure the direct recharge to primary aquifers from rainfall events. Some of the results have been presented in previous sections. One such station has been deployed on the eastern shores of Lake St. Lucia (Fig. 78) where the main atmospheric variables for estimating the potential evaporation (Penman Model) are measured together with rainfall and groundwater level at hourly intervals. Rawlins and Kelbe (1991b) and Kelbe and Rawlins (1992c) have used these data to estimate the direct recharge rate to the primary aquifers (Fig. 85). Kelbe and Shezi (Case Study 5) have used several similar stations in Maputaland and Mozambique to examine the influence of depth to water table on this model which has been presented in a previous section of this report.

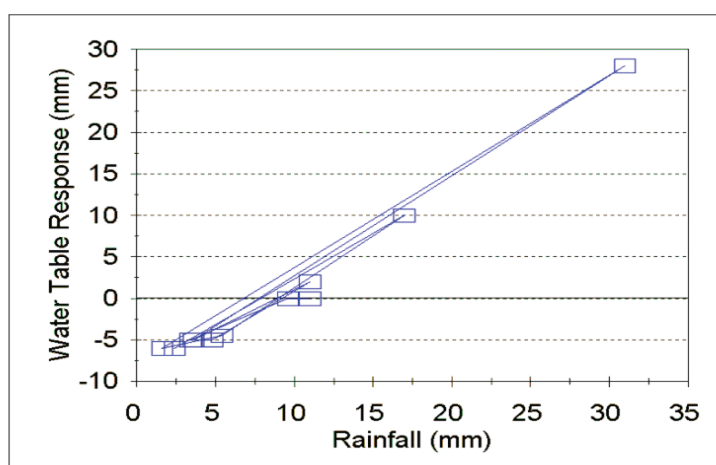


Figure 77: Typical groundwater response to rainfall events for shallow primary aquifers (Kelbe and Rawlins, 1991)

- **Measurement of evaporation.** Evaporation into the atmosphere can be measured in several different ways with varying levels of accuracy (Kelbe *et al.*, 2002). However, only in specific cases, such as open water surfaces, is the source of the evaporating water well known. Previous sections discussed the difficulties in estimating the evaporation component from the groundwater. This is problematic when the impact of land use, particularly afforestation, on the groundwater resources is required. In this project, the evaporation was derived from atmospheric monitoring of radiation, temperature, humidity and wind speed using the Penman model for potential evaporation (E_p). The rooting depth of the vegetation was estimated from the literature. The gross evaporation can be measured with expensive and elaborate instrumentation such as the eddy correlation or Bowen ratio which are not routinely available. The actual evaporation is generally estimated from models such as the Penman-Monteith. It is not possible to measure the component of the evaporation that occurs from the different zones (such as the groundwater). In

this study, the groundwater contribution to evaporation was derived from calibration of numerical models (see the next section).

- **Measurement of spring-flow.** Like springs, the flow in some streams and rivers can be measured under specific conditions using streamflow when the other components of flow can be eliminated. For the Maputaland streams, a simple rectangular section of different dimensions has frequently been used in this study to measure the baseflow together with a velocity meter. Similar approaches using small flumes and weirs are also suitable devices routinely available (Boiten, 2000).



Figure 78: Automatic weather station used for the continuous measurement of rainfall, evaporation parameters and groundwater levels at St. Lucia

- **Measurement of seepage flow.** Outflow (seepage) from the aquifer can be measured in specific cases with varying levels of accuracy.
- **Seepage meters.** Similarly, devices are available for measuring the flux out (into) of aquifers in many aquatic settings. Seepage meters have been used for many years using different principles. Swarzeski *et al.* (2005), Lee (1977), Rosenberry and Morin (2004) and Taniguchi and Fukuo (1993) have described the use of such instrument for measuring the velocity and direction of flow at the interface between groundwater and surface water features such as lakes, wetlands and estuaries. Scolkovitz and Morin (2004) have described a seepage meter used to measure the temporal variability in lake seepage. John and Lock (1977) describe a cylindrical container, closed at one end that is pressed a few centimetres into the sediments ensuring that it is completely immersed. During this process, water is vented through a valve and any trapped air released. After allowing a short period for settling a measuring head is attached, comprising a valve connecting the container to a preformed, readily deformable rubber sack in a collapsed state. The sack is buffered from external water movement by a perforated clear plastic cover. Measurement is effected by opening the main pipe and closing the venting system. The water is displaced by that entering through the sediments during a measured interval of time is collected in the sack. The volume of gas entering the container from the sediments is measured and subtracted from the water volume. A device is being developed by the project team but has not yet been tested.
- **Tracer measurements.** Sholkovitz *et al.* (2003) and Swarzenski *et al.* (2001) have described the use of multiple geochemical tracers to characterize the hydrogeology of the submarine spring. The project team used food dyes to determine the rate of movement of overland sheet flow in St.

Lucia seeping from the exposed shoreline of Lake St. Lucia after the lake had subsided during the recent drought (Fig. 79). This was then used to derive an estimate of the seepage rate through the shoreline. Seepage from the shoreline that resulted in sheet flow down the beach at St. Lucia was estimated from a measure of the flow velocity, and spatial dimensions using a marker dye (Fig. 79). In some cases where the sheet flow converged it was possible to measure the discharge using a small cross-section.



Figure 79: The use of tracers to determine the rate of movement of overland sheet

- **Measurement of subsurface flow.** The research team (Dindi, Rainaldi, Korsman and Kelbe) developed and tested a simple device to measure the flow rate of groundwater in the near surface zone using EC principles and injected NaCl tracer. Six EC probes were inserted into the water table in a suitable array and a salt slug was added to the groundwater and its rate of movement monitored using a data logger (Fig. 80) The data logger provided time series traces of each electrode and identified the time the tracer reach the monitoring point. It was important to add the tracer slug in a manner that did not induce subsurface flow. This system is still being developed for further use in St. Lucia.



Figure 80: Photograph showing the EC probes and array of monitoring points used to measure flow rate of groundwater in St Lucia (Korsman, 2005)

6.2 Mass Balance Approaches

While the physical dimension of aquifers and other water bodies can be derived using the techniques described previously, these do not provide estimates of the interaction with the surface water bodies unless they are incorporated into other models. Most models that quantify the interaction use mass balance principles coupled to other models that describe the fluxes. The conceptual models of the boundaries have been described in the previous sections. These can be described by mathematical models and linked to the mass balance models. Some of the methods tested in this project are described in the next sections.

6.2.1 *Steady State = Equilibrium models*

The first step in most investigations is to start by evaluating the average conditions which are intended to represent the most probable state of the system. In the case of aquifers, the average state generally refers to the most likely aquifer recharge conditions in equilibrium with the discharge states of the various boundaries. This state is an indicator of the sustainable conditions for management purposes. If the recharge is normally distributed, then the fluctuations in recharge are equally distributed about the mean/average. If the recharge is closely related to rainfall, and annual rainfall is normally distributed, then the mean or average rainfall is a good indicator of the sustainable levels of the aquifer.

This state is often referred to as a steady state condition in the assessment of aquifer conditions as it implies no change in storage where the recharge is equal to the discharge and an equilibrium state exists. This is an important state for evaluating the surface water/groundwater interaction as many features at the interface with the aquifer have evolved around the average conditions. For example, swamp forests and many wetland species are good indicators of consistent (average) access to groundwater. The diversity in species may then be an indicator of the variability about the mean.

One important issue in the assessment of the equilibrium states for primary aquifers with well-defined boundaries is that the sum of the discharges is equal to the recharge. In a simple case where the aquifer is completely drained by a network of well defined streams, the total baseflow in these streams is a good indicator of the recharge after taking account of the evaporation from the groundwater. This principle has been used to estimate the recharge for the eShikawini Aquifer to the south of Richards Bay (Fig. 81) by measuring the baseflow in a section of the stream network (red triangles shown in Fig. 81). The groundwater contours were used to identify the catchment area of the streams (Fig. 81) which was then used to establish a mass balance for the catchment. The recharge to the catchment zone was adjusted until the simulated streamflow matched the measured streamflow. Figure 81 shows the catchment area used to calculate the mass balance components of the aquifer that were compared to the measured flow rates. The total streamflow at the sites shown was 5 890 m³/d which compared very favourably with the calculated value of 5 900 m³/d (difference between stream inflow and outflow fluxes). The evaporation component of the mass balance is for losses from the aquifer and must be subtracted from the recharge component to get the net recharge rate (~32% MAP). This is then subtracted from the actual rainfall to get an estimate of the overall evaporation losses (Ea). In the example of Fig. 81, the actual evaporation loss was estimated to be about 68% of MAP (i.e. recharge of 32% MAP). This is slightly higher than the recharge reported by Meyer and Godfrey (1995) for Zululand (18 to -24%).

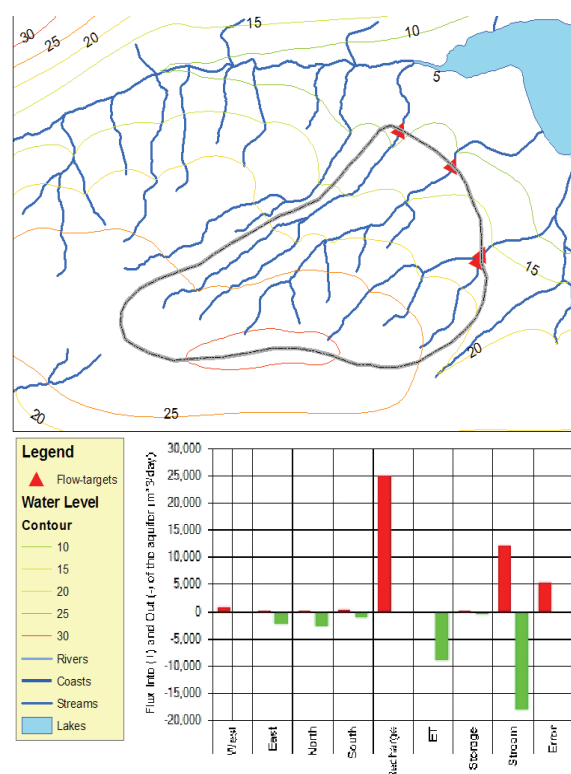


Figure 81: Example of the drainage network and mass balance components used to calibrate the recharge for the coastal dune aquifer in eShikawini, Richards Bay

The same principles can be applied to lakes, wetlands and other water bodies but the recharge is derived from multiple sources such as rainfall, groundwater, rivers, etc., if the recharge zone can be identified. However, this principle cannot be used in a case like the eastern shores of Lake St. Lucia where the entire aquifer is surrounded by specified head boundaries where the discharge cannot be measured. Nor can it be used for the St. Lucia Estuary where seepage along the shoreline constitutes the main groundwater component that cannot be measured easily. An alternate approach was adopted by Kelbe and Taylor (2005) to investigate the groundwater contribution to an estuary where the stream inflow was a minor component and seepage was considered to be the main groundwater source.

A simple mass balance of the St. Lucia Estuary and lakes during periods of fluctuation when the mouth was closed provided a measure of the groundwater interaction with the system (Kelbe and Taylor, 2005). A detailed bathymetry of the system components was used to derive a set of rating curves (Fig. 82) described in the next section. These were incorporated into a mass balance model shown schematically in Figs. 83-84. This model calculated the fluctuation in the volumes that were recharged by direct rainfall, overflow from the Umfolozi (measured), and groundwater seepage (calibrated) which were balanced against the evaporation losses and the transfers between compartments. The model was calibrated against lake levels and salinity. The effect of groundwater on the water quality has been described in other sections. This model will be compared to 3D finite difference models in the near future.

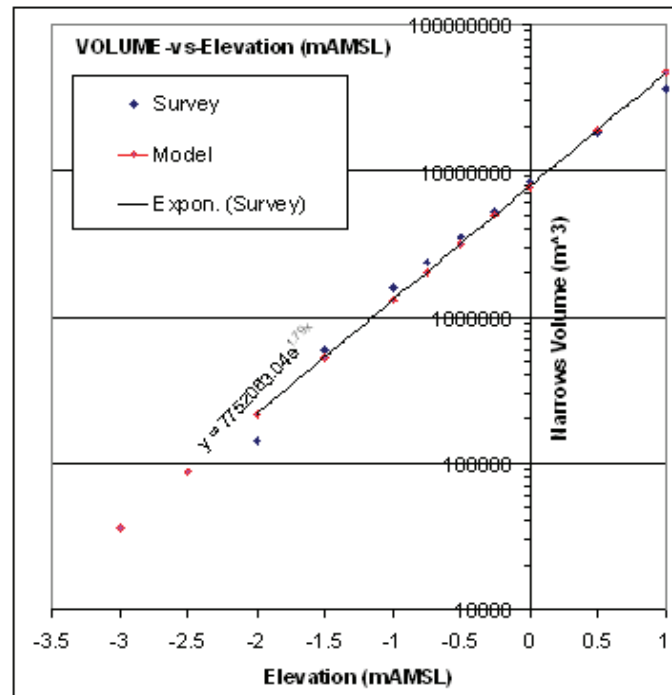


Figure 82: Derived rating curve for the St Lucia Narrows using surface contouring methods

6.2.2 Measurement of Change in Volumes

Several methods and systems can provide an estimate of the physical dimensions of aquifers that have been described in previous sections. The creation of surfaces for the boundaries of aquifer or aquifer units (HSU) can be used to determine the volume defined by another surface or the planar area or wetted surface. This technique was used to determine the rating curve for the lakes compartments in St. Lucia to develop a mass balance model of the system. Figure 82 shows the rating curve for the Narrows volume (m³) in relation to the depth (m a.m.s.l.). Similar relationships for volume, area and perimeter were derived for the other compartments at St. Lucia.

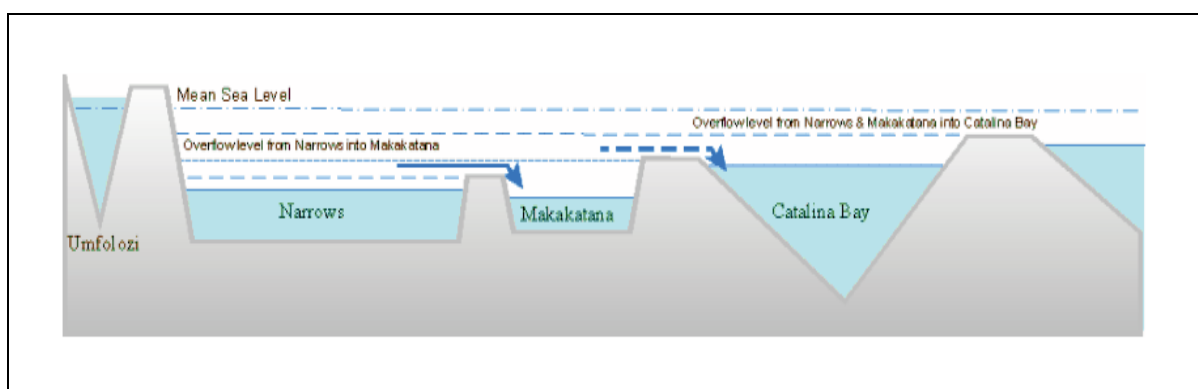


Figure 83: Schematic of the St Lucia Narrows water balance model (Kelbe and Taylor, 2005)

These relationships were then used to calculate the change in the evaporating surface and seepage area.

The derivation of surface boundaries using the techniques described above can equally be used to determine the volume of aquifers. Figure 23 explained the technique used to determine the volumetric extent of the Uloa formation in Richards Bay. If the storage coefficient is known, the volume of water in the aquifer can then be determined. Several computer programs exist that can determine the surface boundaries and derive the physical dimensions of the enclosed bodies. The two programs that were tested and applied in this study are:

- SURFER
- 3D Analyst in ArcView/ArcGIS

Groundwater monitoring is expensive and consequently, simulation models are providing a quick and pragmatic method for assessment of groundwater systems. However, the measurements of aquifer boundaries are invaluable for calibrating the groundwater models. From all the studies on the Maputaland coastal plain, it was abundantly clear that there were never enough measurements of any type to completely satisfy the calibration and application of groundwater models. These groundwater models have been used extensively for many purposes such as environmental impact assessments (Kelbe and Rawlins, 1992a; 1996; 1998), Lake Water Requirement (Kelbe and Germishuyse, 2000a) studies and waste site investigations (Kelbe *et al.*, 1995, Kelbe and Germishuyse, 1997).

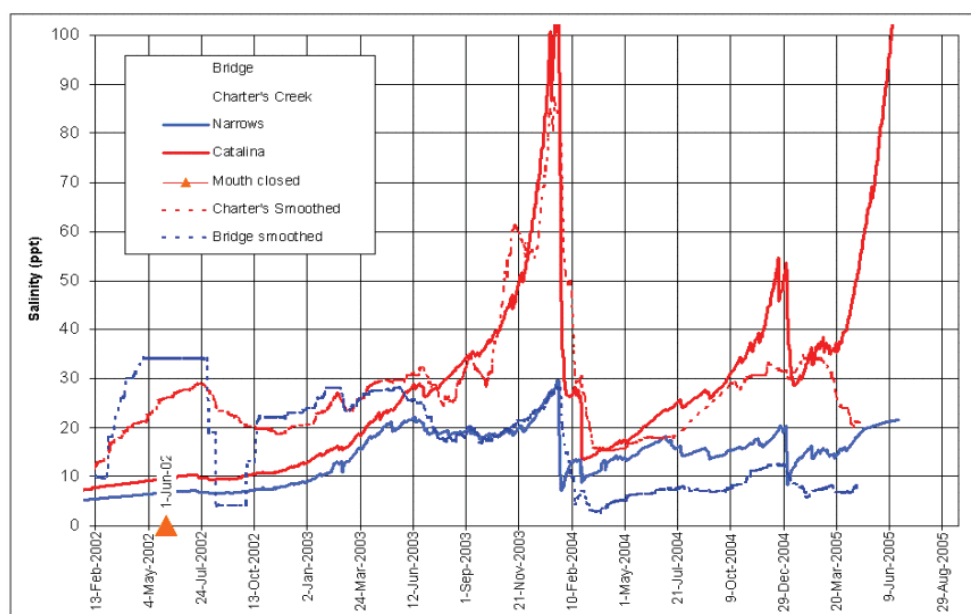


Figure 84: Simulated and observed salinities in the estuary of Lake St. Lucia based on a simple mass balance model incorporating groundwater influx during extreme drought conditions

6.3 Numerical Models

Several models have been used to quantify the interactions between water bodies and aquifers. These can be classified into several categories:

- **Finite element.** In fluid mechanics and associated engineering disciplines, the most common models are based on finite element techniques. Finite element groundwater models have become common.

- **Stochastic models.** A Bayesian model has not been sufficiently well investigated to be covered in this report
- **Finite difference.** The more common models in geohydrology are based on finite difference techniques. Arguably the most common is the USGS MODFLOW code (McDonald and Harbough, 1983). It is a well-known groundwater model which has a central kernel that is supported by many linked packages that model the boundary conditions and provide I/O and analytical functions.

6.3.1 Finite Element Models

The common numerical models described in this report are developed to describing the distribution of heads, hydraulic conductivities and storage properties everywhere in the system using Eqs. (1) and (2) (Boussinesq Equation for unconfined aquifers - Anderson and Woessner, 1992). When gross simplification is applied to the non-linear Boussinesq Equation (Eq. (2)) it can be solved analytically. Except for applications to well hydraulics, these analytical solutions are not widely used for practical applications and numerical solutions are generally preferred (Anderson and Woessner, 1992). Anderson and Woessner (1992) indicate that there are five common numerical methods used in groundwater flow modelling, namely finite element (FE), finite difference (FD), integrated finite difference (IFD), the boundary integral equation method and analytical elements. They go on to indicate that FE and FD are the more commonly used methods and IFD is similar to FE. Anderson and Woessner (1992) describe AQUIFEM-1 which is a finite element model developed by Townley and Wilson (1980).

The finite element models are growing in popularity. However, they were not tested in this project. One that is being considered for future use in the project is the Finite Element Subsurface Flow & Transport Simulation System (FEFLOW) developed by the Institute for Water Resources Planning and System Research (Diersch, 2002). However, this model has not been investigated in this project. For a summary of finite element modelling, the reader is referred to Anderson and Woessner (1992) or Kinzelbach (1989).

According to Anderson and Woessner (1992), finite difference models have gained wider acceptance than finite element models, as they generally require fewer data inputs, but the selection of choice is generally based on user preference. They go on to demonstrate that the finite difference model is a special case of the finite element method. However, this report is restricted to an evaluation of the finite difference methods described in the next section.

6.3.2 Finite Difference Models

(A) The complexity and variances in describing the interaction between SW and GW have been presented in preceding sections. It suffices to say that a single model that can describe the interactive processes involving SW and GW must be a gross simplification of the real world. An international attempt to develop a comprehensive numerical model that incorporates many of the interactive processes involving surface water and groundwater features was developed in 1981 by a consortium of scientists from three national organizations in Britain (Institute of Hydrology), Denmark (Danish Hydraulic Institute (DHI)), and France (Societe Grenoblois d'Etude et d'Applications Hydrauliques). This model, called the Systeme Hydrologique Europeen (SHE) has subsequently been greatly simplified and merged with GIS to

provide an integrated model of the complete modelling environment designed for modelling all the major processes in the land phase of the hydrological cycle. Subsequently, DHI have continued with the project to create MIKE SHE, a more simplified system that includes precipitation, vegetation-based evaporation, overland sheet flow, river and channel flow, unsaturated subsurface flow and saturated groundwater flow, thereby allowing greater flexibility in application. They have also extended the suite of MIKE models that include the MIKE BASIN, Mike FLOOD and MIKE 11. These are expensive models that have not been tested in this study.

(B) The finite difference solution to the nonlinear Boussinesq Equation is set by defining the basic equation in finite difference form and then solving the resultant matrix using iteration techniques. A good text on finite difference modelling is presented by Anderson and Woessner (1992) and Kinzelbach (1989). The most common form of finite difference model is the USGS code called MODFLOW. MODFLOW (McDonald and Harbaugh, 1983) is a block-centred finite difference code that configures a three-dimensional system as a sequence of layers of porous material. The layers are defined by the upper and lower boundaries and may be configured as confined or unconfined. The vertical transmission (leakance) between layers needs to be specified and is often calculated from the (harmonic) mean of the hydraulic properties of the upper and lower layers. For further details see McDonald and Harbaugh (1983) or Harbaugh (1990).

The MODFLOW configuration requires a vast array of parameter values that specify the spatial and temporal variability of hydraulic properties and boundary conditions. These include K_x , K_y , K_z (hydraulic conductivity in the x , y , z axes, S (in various forms) and W (see Eq. (3)) for each node in the 3D space. Most of these parameters are unknown and need to be derived for each cell from the calibration techniques using observation data in the context of a conceptual model. The specification of the parameters is usually done by assuming varying levels of homogeneity in the identifiable hydrostratigraphic units or zones. The general approach is to start with a lumped model where all nodes in the model conform to flow under conditions of homogeneity. As HSU with anisotropic and heterogeneous features are identified, it is necessary to specify zones and layers in a model with different hydraulic parameters. The spatial extent (zones) of identifiable soils is a possible basis for defining zones with common hydraulic parameters. An example is shown in Fig. 86 for a 7-zone model. This is equivalent to lumped catchment parameters in runoff models. It is unlikely that a model can be suitably calibrated if the conceptual basis is not sufficiently well defined.

The boundary conditions are specific to each cell and need to be specified individually. However, it is possible to specify each finite difference cell or boundary individually if the information is available. An example of a transect and plan view of the identified and calibrated HSU (zones) are shown in Fig. 85. In this example the units were specified on the basis of extensive geological explorations and pump tests. Normally these levels of detail are unknown and the entire layer may be classified as one homogeneous unit.

The MODFLOW code has been written to support the addition of interactive packages that expand the capabilities of the main program (kernel). The original MODFLOW code was upgraded to MODFLOW 2000 to support additional features in addition to the application of inverse modelling and solute transport (Harbaugh *et al.*, 2000). Modifications to MODFLOW have been presented over time. In 1992 MODFLOP (Hill

et al., 2000) was followed by MOC3D (Konikow *et al.*, 1996) which extended the capabilities of MODFLOW by solving calibration and solute transport simulations.

This project has specifically excluded solute transport models and they are not discussed any further. However, the application of solute transport modelling in Richards Bay has been presented by Kelbe and Germishuys (1998). MODFLOW was developed as a modular system with the main kernel managing the subroutines comprising packages that have various functions. Because MODFLOW has become the generally accepted groundwater model in the industry and has made the source code open to the public (Open Source), it has attracted many additional packages that provide added functionality. It is beyond the scope of this report to describe all the additional packages that have been presented in the literature. Only those of direct relevance to the surface water/groundwater interaction and specifically those that have been evaluated and applied to the groundwater studies in Maputaland are presented here.

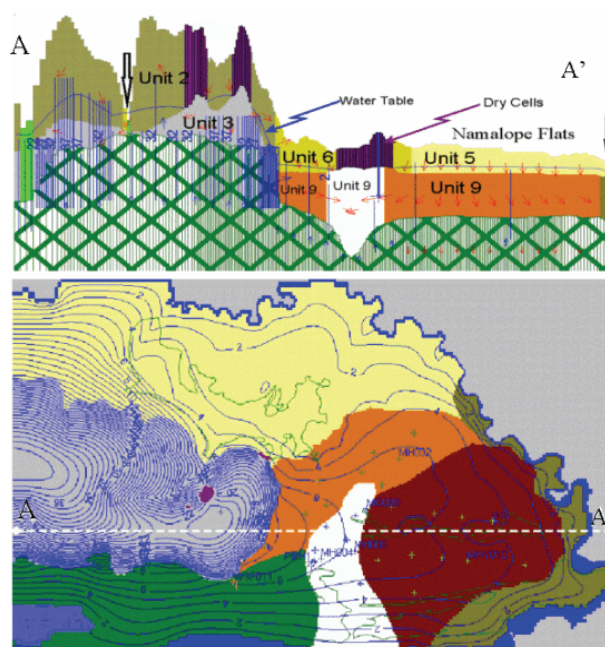


Figure 85: Example of a transect and plan view showing the HSU zones specified to limit heterogeneity

6.4 Boundary Types

In the Maputaland region, all models are configured as unconfined aquifers in the upper layer that interacts directly with the boundaries. The lower layers have been configured according to their level of interaction with the upper layer. In many instances where the upper layer becomes dewatered, it was considered appropriate to configure the second layer as an unconfined aquifer with varying storage. In other aquifers, it was necessary to either configure the second and subsequent layers as confined with a specified vertical leakance. In some of these it was necessary to extend the boundary conditions, such as the harbour and Lake Mzingazi, into the lower layers. In all cases, the rivers and streams were restricted

to the upper layer while the recharge was limited to the upper active layer. Evaporation occurred in any layer where the water table was within the rooting zone of the vegetation.

6.4.1 Specified Head Boundaries

Specified head boundaries are assumed for all large water bodies that have a volumetric capacity that is substantially greater than any groundwater contribution. This includes the ocean, harbours and large estuaries. The fluctuations in elevation of these large water bodies were varied during specific phases (stress periods) of the simulation period if the levels were known. In certain instances, lakes were configured as constant head boundaries if the purpose was to model the groundwater and not the lake water balance. Where the level of interaction was required, it was necessary to simulate the lake as a head dependent boundary described below.

6.4.2 Specified Flow Boundaries

Specific flux boundaries include the recharge, evaporation and abstraction/injection wells. The wells are specified for specific points whereas the recharge and evaporation need to be specified for the model domain.

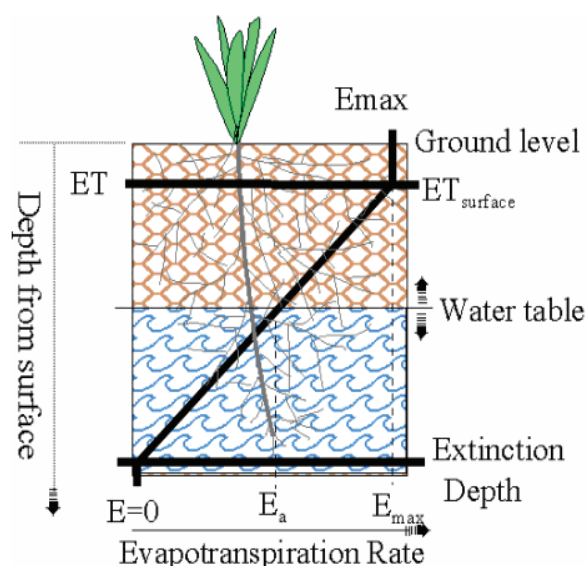


Figure 86: Conceptual model of groundwater evaporation used in the Maputaland studies and employed in the MODFLOW model

The recharge was applied to the upper active layer and varied according to a preconceived conceptual model of the controlling factors such as interception, infiltration and runoff that were external to the MODFLOW model configuration. The average recharge was calibrated against the 'average' river baseflow, assuming all other parameters were constant. The baseflow was derived from ad hoc measurements that were assumed to represent average conditions for the sake of the calibration as no regular monitoring exists in any of the study areas.

The evaporation was assumed to be a proportion of the potential rate for each land use type and the rooting depth derived from the literature. The MODFLOW evaporation package assumed that the rate of

discharge was zero below the rooting depth and increased linearly to the specified rate at the surface (Fig. 86). Other models such as Intersat (Kelbe and Rawlins, 1992a) specify an upper level where the maximum rate is reached.

6.4.3 Head-Dependent Boundaries

Rivers, lakes, and wetland volumes that interact with aquifers have sufficient capacity to be influenced by the rate of seepage. These boundaries were treated as head-dependent boundaries (Fig. 87) where the seepage rate is described by Darcy's law. The rate of flow between the aquifer and these boundaries is determined by the hydraulic gradient between the two features and the conductance of the intervening layer. The flux of water into or out of the model node is then computed by the term:

$$Q = C(H_b - H_m)$$

where:

Q = flux into or out of boundary cell (L^3/T)

H_b = boundary head (L)

H_m = aquifer head (L)

C = boundary conductance (L^2/T)

The conductance term is a coefficient that is usually computed using an equation similar to the following:

$$C = K_b A / B$$

where:

K_b = hydraulic conductivity of the boundary material (L/T)

A = area of the boundary (L^2)

B = thickness or width of boundary (L)

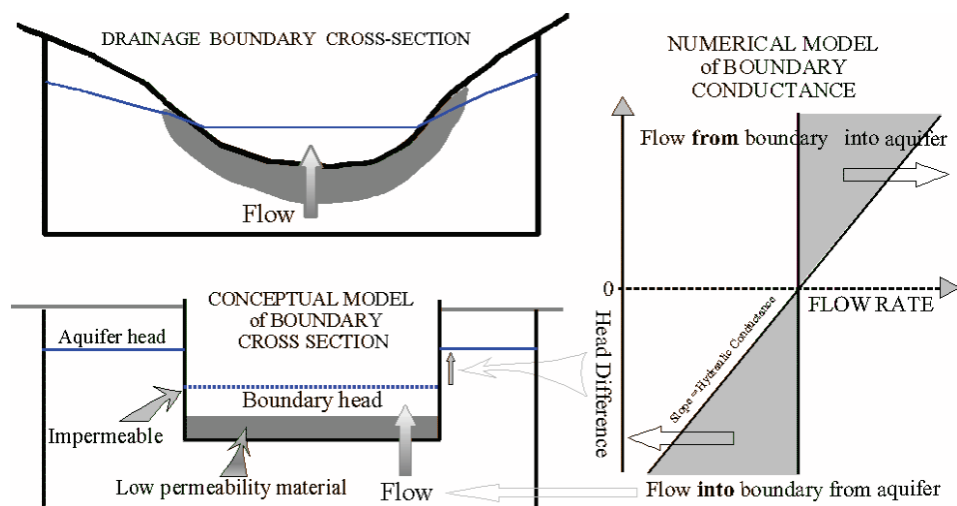


Figure 87: The numerical form of the conceptualized head-dependent boundary

Specialized packages have been presented for the following types of boundaries:

6.4.3.1 River

The conceptual model of the river boundary using the above numerical format conforms to Fig. 87. These packages are incorporated in MODFLOW (98). The RIV package is described in detail by McDonald and Harbaugh (1983) and the STR1 package by (Prudic, 1989). The hydraulic head gradient is between the river stage and the aquifer head in the cell occupied by the river section. The conductance is through the river bed sediments of thickness B and specified hydraulic conductivity (K_b). This boundary package was evaluated in Case Study 2 for a small section of a river in Richards Bay.

If K_b is set greater than the aquifer equivalent parameter, then the river flow is dependent on the aquifer properties. This package is able to simulate both aquifer discharge into the river and transmission losses back into the aquifer from the river if the appropriate conditions exist. The stage of the river is determined from the inflow (Q) at the start of each reach and the channel dimensions (width).

6.4.3.2 Drains

Drains are a special case of the model depicted in Fig. 87 where there is no reverse flow from the boundary into the aquifer. If the head in the model cell drops below the drain elevation, the drain will not inject water into the model. Under these conditions, the drain becomes inactive.

6.4.3.3 Streams

The increasing need to quantify the interaction between the aquifer and streams led Prudic *et al.* (2004) to develop a new streamflow routing (SFR1) package to simulate the stream-aquifer interaction with MODFLOW 2000. This package has been developed to link with the Lake Package (LAK3) described in the next section. These packages do not work with MODFLOW 98.

This package allows for the addition and subtraction of water from runoff, precipitation and evaporation with each node. It routes flow to specified channel sections and can divert water to other boundaries such as constant flux/head and as direct abstraction.

The width of the stream can be specified or calculated on the basis of the streamflow using channel characteristics. The SFR1 package provides 5 options for computing stream depth on the basis of streamflow (Q) in each reach and streambed elevation (**Note:** this can lead to inconsistencies if the streambed elevation is not specified properly):

- Same as RIV but calculated at the midpoint of the reach
- Manning's equation used to route water down the channel and (optional) the flood plain where one of several options is available
- Assuming a rectangular channel with width $\gg$ depth
- Specified channel geometry using 8 point cross-section
- Based on depth-discharge and width-discharge relationships for Q . Specify the values of depth and width for given stream-flows

This model has been used in conjunction with the lake package to route river flow into and out of the Lake Cubhu described in subsequent sections.

6.4.4 Other Boundaries

It is desirable to include only natural hydrologic boundaries as boundary conditions in the model. Most numerical models, however, employ a grid that must end somewhere. Thus, it is often unavoidable to specify artificial boundaries at the edges of the model. When these grid boundaries are sufficiently remote from the area of interest, the artificial conditions on the grid boundary do not significantly impact the predictive capabilities of the model. However, the impact of artificial boundaries should always be tested and thoroughly documented in the model report. The model grid should be expanded to include far more field conditions until the effect of these boundaries on the domain of interest is insignificant. This manual is not intended as a tutorial on the proper use of boundary conditions. If you are interested in more information on boundary conditions, Anderson and Woessner (1992), Franke *et al.* (1987) and Franke and Reilly (1987) provide excellent explanations of boundary condition usage.

Grid-independent boundary conditions are those defined by spatial coordinates rather than row column-layer coordinates. These boundary conditions are assigned to model nodes (row, column, layer) when model data sets are created. The advantage to these types of boundary conditions is that they do not change when you insert or delete row and column grid lines. Grid-independent boundaries are added to the model using the Add menu and are referred to as 'Analytic Elements' in many of the GV menus. This term was adopted to be consistent with other ESI software such as WinFlow. The analytical elements that are supported by GV include wells, line boundaries, and circular boundaries. Wells are always defined as constant flux boundary conditions or may be defined as 'Fracture Wells' for use in MODFLOW-SURFACT. Line and circle boundaries may be constant head, constant flux, or any of the head-dependent boundaries except for streams.

6.4.5 Lake Package (LAK 1 and LAK 3)

The seepage into, and presumably out of, the lake from the aquifer has been identified as an exponential decay function (Pfannkuch and Winter, 1984; McBride and Pfannkuch, 1975; Winter, 1976; Lee, 1977). However, the numerical models are based on the vertical transfer of mass through a conducting layer (lake bed) similar to the previous model shown in Fig. 87. The model of lake-aquifer interaction (LAK1), based on the same principles as those shown in Fig. 87, was developed by Council (1997). This was upgraded by Cheng and Anderson (1993) and subsequently by Merritt and Konikow (2000).

- **LAK1** (Council, 1997) was used to determine the water budget for Lake Mzingazi described in detail in Case Study 6. However, LAK3 in conjunction with the new streamflow routing package (STR1) is being used to determine the water balance components and lake/stream/groundwater interaction for Lake Cubhu in ongoing studies by the project team.
- **LAK3** (Merritt and Konikow, 2000) package is linked to MODFLOW by representing the lake 'as a volume of space within the model grid which consists of inactive cells extending downward from the upper surface of the grid. Active model grid cells bordering this space, representing the adjacent aquifer, exchange water with the lake at a rate determined by the relative heads and by conductances that are based on grid cell dimensions, hydraulic conductivities of the aquifer material, and user specified leakance distributions that represent the resistance to flow through

the material of the lakebed' (Merritt and Konikow, 2000). 'The variations of lake stage are determined by independent water budgets computed for each lake in the model grid. This lake budget process makes the package a simulator of the response of lake stage to hydraulic stresses applied to the aquifer. Implementation of a lake water budget requires input of parameters including those representing the rate of lake atmospheric recharge and evaporation, overland runoff, and the rate of any direct withdrawal from, or augmentation of, the lake volume' (Merritt and Konikow, 2000).

6.5 Other Models

Several other models have been encountered in this project but have not been evaluated. However, they do provide estimates of the surface water/groundwater interaction and are listed here for reference. They include:

- MODPATH, a three-dimensional particle-tracking model that works with MODFLOW. MODPATH was developed by the USGS to simulate particle tracking and includes both steady state and transient simulations. An example of particle tracking in Richards Bay is shown in Fig.88.
- The FLOW-THRU model has been used by Townley *et al.* (1993) to examine groundwater flow in a vertical section near surface water bodies. A large number of flow regimes are identified, with their characteristics controlled by regional water table gradients, recharge to the aquifer, water body length, aquifer anisotropy, and the hydraulic resistance of the bottom sediments.
- The DAFLOW model routes flows through a system of inter-connected one-dimensional channels and subdivides the system into a series of branches, with each branch divided into a number of subreaches.
- MODFLOW simulates groundwater flow through a three-dimensional grid of cells. The models are coupled by adding an exchange between each sub-reach and a specified groundwater cell, with the amount of flow from the stream to groundwater being limited to the available streamflow. The water exchange for each sub-reach is computed on the basis of the stream-aquifer head difference, the streambed thickness, stream width, and streambed hydraulic conductivity.
- MODBRNCH (Swain and Wexler, 1996; Swain, 1992). Surface water and groundwater interactions can be simulated by the coupled BRANCH and USGS modular, three-dimensional, finite-difference groundwater flow (MODFLOW-96) models, referred to as MODBRNCH. MODFLOW-96 simulates steady and non-steady flow in an irregularly shaped flow system in which aquifer layers can be confined, unconfined, or a combination of confined and unconfined. BRANCH simulates steady or unsteady flow in a single open-channel reach (branch) or throughout a system of branches (network) connected in a dendritic or looped pattern by solving the one-dimensional equations of continuity and momentum for the river flow. Channel-aquifer flows are leakage through a confining layer or riverbed. Computation of this leakage in the groundwater and surface-water systems allows these processes to be coupled for simulation purposes.

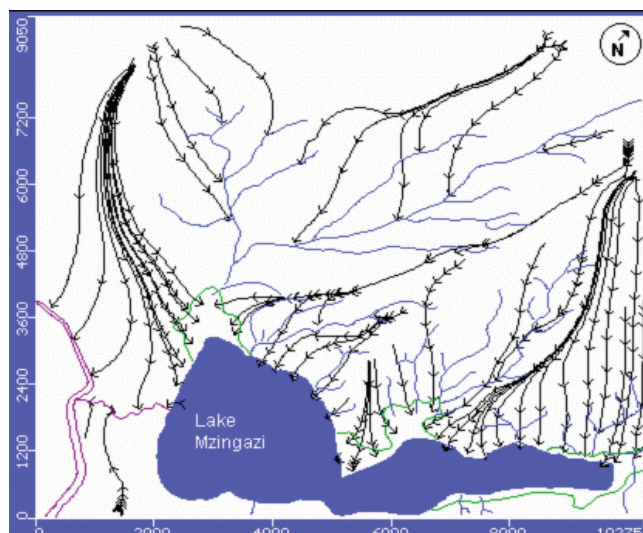


Figure 88: Example of MODPATH simulations linking pathways between sections of the Richards Bay aquifer and Lake Mzingazi

6.6 Snyman Model

This model has been described in Case Study 1 in conjunction with some of the underlying concepts described in previous sections of this report. The basic concepts and model configuration are summarized in Fig. 89. It has to be verified/calibrated on the Ntuze research catchments and applied to the Lake Pobane catchment in the Mhlathuze. The simulated storm events of runoff into Lake Pobane catchment are shown in Fig. 90 together with the simulated series of the three flow components. This model holds potential for estimating the groundwater component of river runoff required for management purposes such as the Reserve determination.

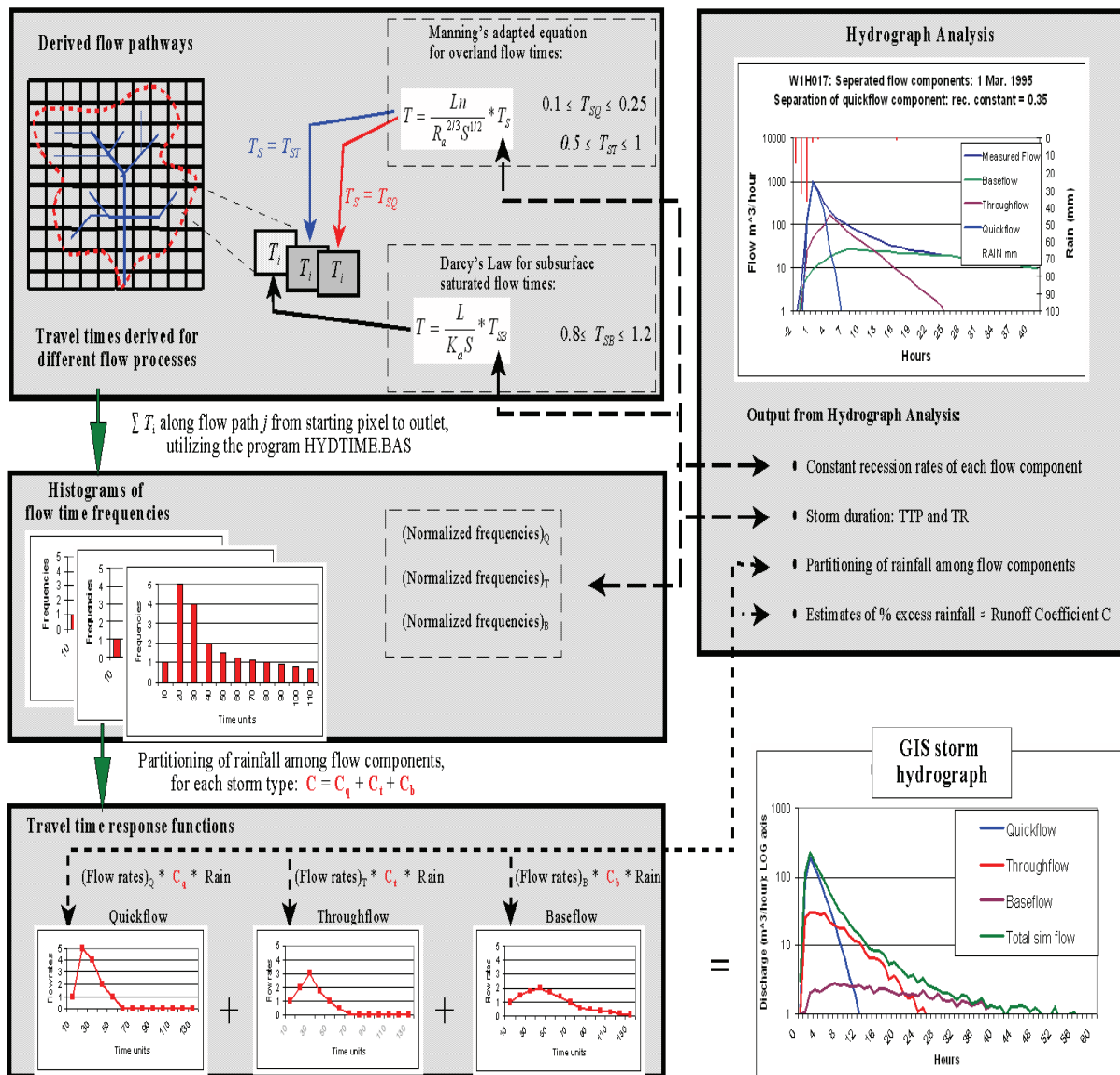


Figure 89: Schematic of the Snyman model for predicting river runoff components of baseflow and throughflow

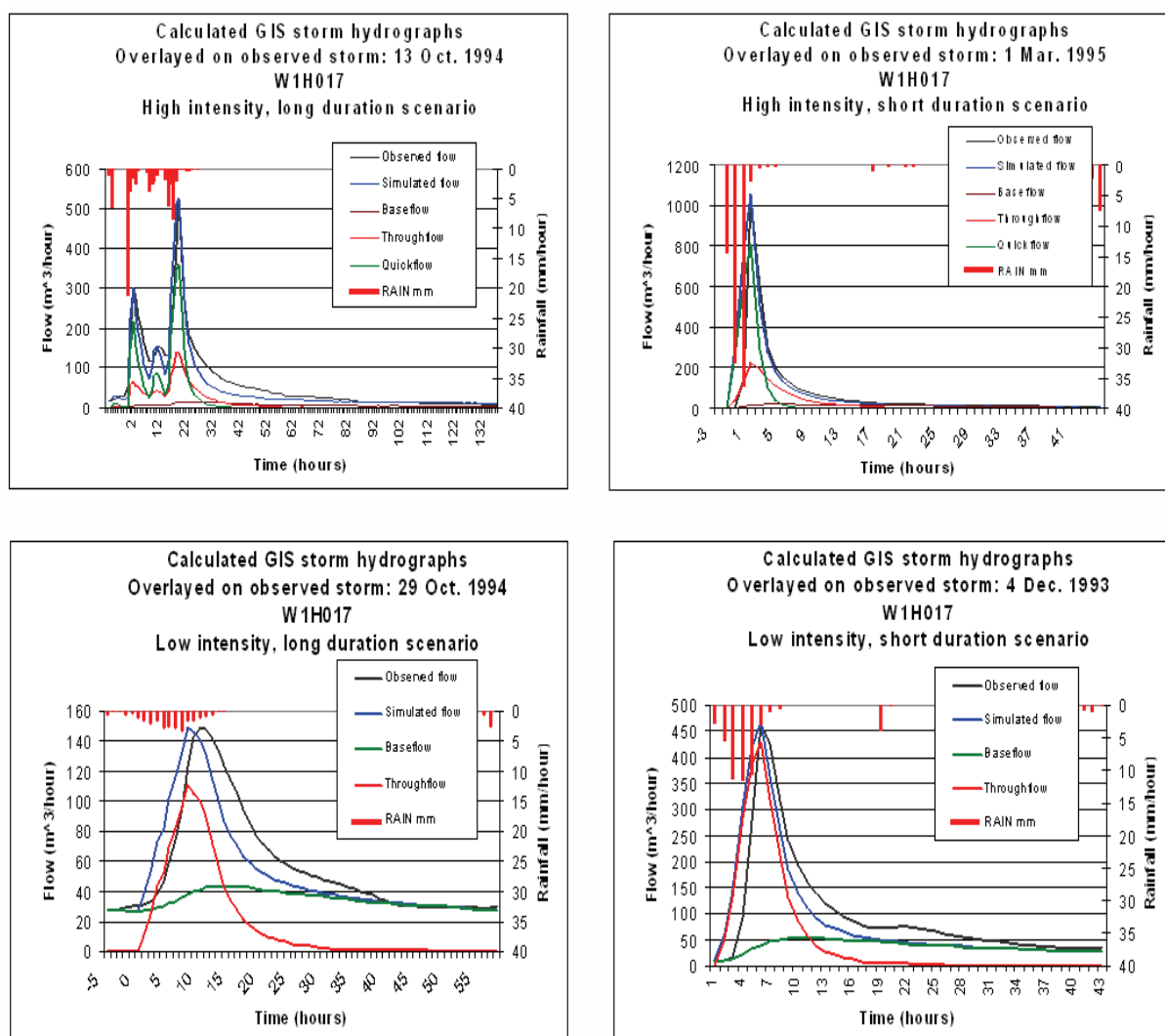


Figure 90: Simulated and observed storm hydrographs of individual storms in the catchment of Lake Pobane, Mhlathuze Catchment (Snyman, 2006)

CHAPTER 7: SUMMARY AND CONCLUSIONS

This report has attempted to describe the main concepts in the surface water-groundwater interaction with specific reference to the shallow primary aquifers found along the south-east coast of Africa. The initial description of the groundwater system and the conceptual models relating to specific surface water resources (interactive boundaries) have been related to experience in the evaluation and quantification of the Maputaland system. It is not an attempt to review generic concepts and models but to document the observations and methods used to describe the process that have been the main focus of the report.

It has been difficult to define a clear and distinct boundary between the definition of surface water and groundwater except through the specification of a physical boundary between zones through which water from the saturated porous medium (called groundwater) passes to an exposed zone of water (called surface water). The many pathways through which water can pass on its journey through the hydrological process make it difficult to create a clear distinct between these resources. A physically based model by Snyman (2005) has been presented that attempts to simulate the processes that differentiate the flow patterns in river runoff to assist with quantifying the level and magnitude of interactions for management purposes.

Groundwater models examined in this project could not simulate the individual flow processes but were generally restricted to only the groundwater component. However, the new stream-routing packages (SFR1) by Prudic *et al.* (2004) model the groundwater contribution to the river runoff but need the other components to be derived from different sources (models).

The finite difference numerical models such as MODFLOW, linked to other boundary packages such as SFR1 and the lake model of Merritt and Konikow (2000) provide the means to simulate the level of groundwater interaction with surface water bodies such as lakes and wetlands where the form and physical features of the streams feeding and draining the resources are specified. Kelbe and Germishuys (2000 a) used similar packages (LAK1 and STR) to evaluate the groundwater interaction with Lake Mzingazi that identified the flow through nature of the system leading to large loss of subterranean water. Similar studies are being pursued to investigate the role of the groundwater for Lake Cubhu and St. Lucia based on previous modelling of the groundwater lake interaction (Wejden, 2004; Kelbe *et al.*, 2005; Taylor *et al.*, 2006).

This study has highlighted several aspects of the investigation into the shallow Maputaland aquifers and their role in sustaining the functions of surface water bodies. The work of Taylor *et al.* (2006) has shown that the groundwater is an essential part of the Lake St. Lucia resilience to ecological functions during climatically induced changes that make it a unique system that has attained world heritage status. Because the lake undergoes frequent and persistent fluctuation in salinity associated with the marine linkages, the groundwater provides refugia sites during hyper-saline epochs that sustain ecological functions. Lake Sibaya, on the other hand is an entirely freshwater system that is an extension of the regional water table under natural conditions. Unlike St. Lucia, groundwater is the main source of inflow, other than rainfall, to maintain the lake levels, so groundwater plays a very different role. It also appears that many other coastal lakes are critically dependent on groundwater in various forms and for

different reasons. Kosi Bay has several lakes with varying levels of marine and groundwater flow. It is anticipated that the role of groundwater would also be unique to the Kosi system.

The study has focused primarily on surface water resources, such as lakes and rivers, which have been identified as important management requirements, principally water allocations. However, the concepts and methods described in this report can be (have been) applied to all exposed water bodies to determine their level of interaction that is required as a precursor to the determination of the importance of the groundwater in sustaining the functions (ecological or otherwise) of the systems.

The study has used, and described, the many different methods and techniques that were evaluated in an attempt to understand and quantify the interaction between the surface water and groundwater of Maputaland. The numerical models were found to play a very pragmatic role in an assessment of groundwater systems. They were built on the basis of a conceptual model of the hydrogeology, calibrated against various types of measurements and used to examine the conceptual models, explore the system dynamics and quantify the level of the interaction. Linking the conceptual model to the observation data through the numerical model provides a good method of examining the conceptual model if the simulations cannot be made to match the observations. Similarly, the surface fitting of observations using numerical models like MODFLOW, is done on the basis of a physically based model, as opposed to a statistical model (such as Kriging), that has much greater interpolation and extrapolation ability. On the basis of a few streamflow measurements and a scattering of water level measurements, including exposed water surface like wetlands, the numerical model is better able to predict the water surface profile than subjective and statistical methods. However, numerical models are critically reliant on adequate data sets to evaluate conceptual models and make predictions on hydrodynamic processes. The numerical model can only be as good as the data used for calibration. One of the modelling studies in this project was used to highlight the severe inconsistencies in otherwise good observation data.

During the course of this study, the principal author has developed a strong belief that the water resources of the Maputaland coastal plain are critically dependent on the groundwater for their form or function. Most of the people of Maputaland, particularly the poor and impoverished people who live off the land, are directly or indirectly dependent on the groundwater for their survival or income. The worldwide recognition of the importance of the wetlands of Maputaland by the declaration of the World Heritage Status for the Greater St. Lucia Wetland Park bears testimony to its ecological importance on a worldwide basis. The ecotourism industry developed to support the sustainability of the Park and its peripheral areas is the main source of income for most of the local inhabitants, one of the most densely populated areas of South Africa.

GLOSSARY OF TERMS

Actual evaporation [E_a or AET] [Braune et al. \(2000\)](#): The amount of water evaporated under a given set of circumstances where water supply limits evaporation (as opposed to potential evaporation which is evaporation under unlimited circumstances, as from an evaporation pan).

Alluvial aquifer [Braune et al. \(2000\)](#): An aquifer formed of unconsolidated sediments deposited by a river or stream; typically occurring beneath or alongside a current channel, or in a buried old or paleo-channel of the river.

Aquatic ecosystem [Braune et al. \(2000\)](#): Not defined in the Water Act. They are defined in the water quality guidelines (DWAF, 1996) as the abiotic (physical and chemical) and biotic components, habitats and ecological processes contained within rivers and their riparian zones, reservoirs lakes and wetlands and their fringing vegetation. Terrestrial biotas, other than humans dependent on aquatic systems for their survival, are included in this definition.

Aquifer [Bowen \(1986\)](#): A permeable deposit which can yield useful quantities of water when tapped by a well. [Driscoll \(1986\)](#): A formation, group of formations, or part of a formation that contains sufficient saturated permeable material to yield economical quantities of water to wells/boreholes and springs. Water Act (1998): A geological formation which has structures or textures that hold water or permit appreciable water movement through them. [Parsons \(1995\)](#): Strata of a group of interconnected strata comprising of saturated earth material capable of conducting groundwater and of yielding usable quantities of groundwater to boreholes and /or springs. A supply rate of 0.1 ℓ/s was considered to be a usable quantity. It is implied that the quality of the water should be fit-for-use by the end user. [Todd \(1980\)](#): A formation that contains sufficient saturated permeable material to yield significant quantities of water to wells and springs. [Vegter \(1995\)](#): A stratum which contains intergranular interstices, or fissures/fractures (as such) of a system of interconnected fissures/fractures (as such) capable of transmitting groundwater rapidly enough to directly supply a borehole or spring. The fissures/fractures are generally bound by either aquiclude/aquitard or aquiclude.

Baseflow [Hatton et al. \(1998\)](#): Baseflow is a purely conceptual quantity resulting from statistical or graphical methods of breaking up a stream hydrograph into components of quick and slow responses to rainfall. These are not directly equivalent to the amount of surface and subsurface (groundwater discharge) flow. Baseflow is made up of several components, such as groundwater discharge, bank storage and lateral unsaturated flow. However, baseflow is generally considered to arise to some significant degree from groundwater discharge to streams. [Todd \(1980\)](#): Streamflow originating from groundwater discharge is referred to as groundwater runoff or baseflow. During periods of precipitation streamflow is derived primarily from surface runoff, whereas during extended dry periods all streamflow may be contributed by baseflow. Typically, baseflow is not subject to wide fluctuation and is indicative of aquifer characteristics within a basin. [Vegter \(1995\)](#): That part of streamflow which is contributed by effluent groundwater.

Biodiversity [Braune et al. \(2000\)](#): A term which encompasses the diversity of living organisms in a system, their interactions and their roles maintaining ecological processes and functions. Formally, there are three components: (a) composition: what is there and how abundant it is; (b) structure: how the

units are organised (structured) in space and time; and (c) function: the roles the different units play in maintaining processes and dynamics. These three components are each represented at four different levels or scales of organisations: (i) genes, (ii) species and/or populations, (iii) communities (habitats) or ecosystems and (iv) landscapes (Noss, 1990).

Bog Kotze (2000): Bog: a mire (i.e. a peat accumulating wetland) that is hydrologically isolated, meaning that it is only fed by water falling directly on it as rain or snow and does not receive any water from a surrounding catchment. Bogs generally have acidic waters and are often (but certainly not always) dominated by mosses. Mitsch and Gosselink (1986): The term bog is frequently used much more broadly (and incorrectly) in South Africa to refer to high altitude wetlands that have organic-rich soils. Most of these wetlands would not be bogs in the correct sense. The lack of true bogs in South Africa is probably due mainly to the fact that evaporation is generally high in relation to precipitation. Winter *et al.* (1998): Bogs are wetlands that occupy uplands or extensive flat areas, and they receive much of their water and chemical constituents from precipitation.

Borehole Braune *et al.* (2000): Includes a well, excavation or any artificially constructed or improved underground cavity which can be used for the purpose of (a) intercepting, collecting or storing water in or removing water from an aquifer; (b) observing and collecting data and information on water in an aquifer; or (c) recharging an aquifer; Cape Water Programme (1999): Generic term used for any drilled or hand-dug hole used to abstract or monitor groundwater, irrespective of diameter or construction.

Capillary zone/fringe Bowen (1986): The belt of ground immediately above the water table, i.e. at the bottom of the zone of aeration and containing capillary water. Braune *et al.* (2000): The lower division of the zone of aeration (vadose zone) that overlies the zone of saturation and in which the pressure of water in the interstices is less than atmospheric (McGraw-Hill, 1978). Todd (1980): The capillary zone (or capillary fringe) extends from the water table up to the limit of capillary rise of water. Braune *et al.* (2000): In relation to a watercourse or watercourses or part of a watercourse, means the area from which any rainfall will drain into the watercourse or watercourses or part of a watercourse, through surface flow to a common point or common points. Cape Water Programme (1999): The area of land which drains all runoff from rainfall to the sea (or a lake) via one or more watercourses. A catchment has a definable physical boundary which acts as a barrier to many of the physical and biological processes within it.

Coastal forest Braune *et al.* (2000): Medium to tall forest vegetation dominated by evergreen trees, climbers and shrubs often associated with coastal lakes and swamps. Distributed from Alexandria to the Mozambique border. In the sense used here it includes the sand forest of the northern KwaZulu-Natal coastal belt which is dominated by deciduous trees (see Low and Rebelo, 1996).

Confined aquifer Braune *et al.* (2000): An aquifer that is confined between two less permeable confining beds. The confining bed has a significantly lower hydraulic conductivity than the aquifer.

Confined groundwater Vegter (1995): Groundwater under pressure significantly greater than that of the atmosphere and whose upper surface is the bottom of a layer of distinctly lower permeability than the material in which the water occurs.

Discharge Braune *et al.* (2000) citing *Water Words Dictionary* (1999): In its simplest concept, discharge means outflow and is used as a measure of the rate at which a volume of water passes a given point.

Therefore, the use of this term is not restricted as to course or location, and it can be used to describe the flow of water from a pipe or a drainage basin. With reference to groundwater, the process by which groundwater leaves the Zone of Saturation via Evaporation, Evapotranspiration, or by flow to the surface through springs and seeps.

Ecoregion Braune *et al.* (2000) citing DWAF (1998): The breakdown of a catchment into water resource units for the purpose of determining the reserve for rivers. It is done primarily on a biophysical basis, according to the occurrence of different ecological regions within the catchment.

Ecosystem Braune *et al.* (2000): An interconnected and interacting system comprising living organisms – including animals, plants, fungi, and microorganisms and their non-living environment.

Effective recharge Bredenkamp *et al.* (1995): Is the total recharge minus losses that occur subsequent to infiltration to the groundwater.

Ephemeral Braune *et al.* (2000): rivers, streams or pans which are ephemeral are fed by inflows of surface water after rains and are seasonal. There is no groundwater contribution (baseflow) therefore these are influent features which, in permeable areas, recharge groundwater.

Estuary Braune *et al.* (2000): A partially or fully enclosed body of water which is open to the sea permanently or periodically; and within which the sea water can be diluted, to an extent that is measurable, with fresh water drained from land

Evaporation Braune *et al.* (2000): The total loss of water in vapour form from all sources – open water, from the plant surface (interception), through plants (transpiration) and from the soil surface: it involves the transition of water from the liquid phase to the vapour phase, and during this process energy (termed latent heat) is absorbed. Carageeg *et al.* (1987): Evaporation occurs when water is converted from the liquid to the vapour phase and returned to the atmosphere. It occurs from open water surfaces, such as water intercepted by vegetation, ponded in depressions, flowing as surface runoff, or stored in wetlands or lakes. It also removes water directly from the soil. Its rate depends on atmospheric conditions (solar radiation, wind speed, temperature and humidity) and on conditions within the body of water (temperature).

Evapotranspiration Braune *et al.* (2000): This refers to that component of the evaporated water that comes from transpiration by, or interception losses, from plants. In modern usage it is often replaced by the term evaporation, as defined above. Carageeg *et al.* (1987): Evapotranspiration is a collective term used to describe the combination of evaporation and transpiration, both of which depend on atmospheric conditions and on the availability of water. Vegter (1995): Loss of water from a land area through transpiration of plants and evaporation from the soils.

Exfiltration Carageeg *et al.* (1987): Exfiltration is a term used to describe evaporation of soil moisture, or evaporation directly from the water table when the latter is close to but below the land surface. It occurs due to atmospheric demand for water (i.e. in dry or windy conditions) when soil is moist or the water table is shallow.

Fault Driscoll (1986): A fracture or zone of fractures along which there has been displacement of the sides relative to one another, parallel to the fracture.

Minerotrophic mire usually has a wider range of species than bogs. [Kotze D \(2000\)](#): Fen: a mire (i.e. a peat accumulating wetland) that receives some drainage from soil in the surrounding catchment, i.e. it is not hydrologically isolated. According to the *Oxford Compendium*: fen1 // n. (1) a low marshy or flooded area of land. [Winter et al. \(1998\)](#): Fens are wetlands that commonly receive groundwater discharge; therefore, they receive a continuous supply of chemical constituents dissolved in the groundwater. Fens can be present on slopes.

Finite difference model [Braune et al. \(2000\)](#): A particular kind of digital computer model based upon a rectangular grid that sets the boundaries of the model and nodes, where the model will be solved.

Finite element model [Braune et al. \(2000\)](#): A digital groundwater- flow computer in which the aquifer is divided into a mesh formed of a number of polygonal cells.

Geohydrology [Vegter \(1995\)](#): The branch of hydrology dealing with subsurface water, i.e. water in both the saturated and unsaturated zones. Also used interchangeably with hydrogeology.

Geological formation [Braune et al. \(2000\)](#): The fundamental lithostratigraphic unit and may consist of consolidated or unconsolidated material (Bates and Jackson 1980). This definition could arguably include formations such as the Kalahari sands and the Quaternary sands of the dune fields in the Western Cape.

Geology [Braune et al. \(2000\)](#) citing *Water Words Dictionary* (1999): The science that studies the physical nature and history of the earth.

Geomorphology [Braune et al. \(2000\)](#): The study of the configuration of the earth's surface, including classification, description, nature, origin and development and their relationships to the underlying geological structures and the history of geological changes seen in these structures. Includes and has largely replaced the concept and study of physiography (Bates and Jackson, 1980).

Groundwater [Bowen \(1986\)](#): Water in the zone of saturation. [Parsons \(1995\)](#): All subsurface water occupying voids within a geological stratum. [Driscoll \(1986\)](#): Underground water that can be removed by wells. All other water in the ground is termed subsurface water and is not available for man's use directly. [Fetter \(1994\)](#): Water stored in the zone of saturation. [Todd \(1980\)](#): Water occupying all the voids within a geologic stratum. This saturated zone is to be distinguished from an unsaturated, or aeration, zone where voids are filled with water and air. [Vegter \(1995\)](#): Water in the zone of saturation. It flows into boreholes/wells or debouches as springs. Synonym: underground or subterranean water.

Groundwater basins [Todd \(1980\)](#): A hydrogeological unit containing one large aquifer or several connected and interrelated aquifers. It implies an area containing a groundwater reservoir capable of furnishing a substantial water supply.

Groundwater body [Braune et al. \(2000\)](#): In this document the term is used to include both the aquifer in its conventional sense (see definition) and significant volumes of water stored in (deep) colluvial or alluvial deposits or deeply weathered rock (regolith).

Groundwater level [Vegter \(1995\)](#): The water level in a borehole/well penetrating the zone of saturation and from which no water is being withdrawn. Synonym: static water level.

Groundwater modeller Braune *et al.* (2000): The person to carry out the simulation of groundwater flow and transport, with numerical methods.

Groundwater quality Braune *et al.* (2000) citing *Water Words Dictionary* (1999): A term used to describe the chemical, physical, and biological characteristics of groundwater, usually in respect to its suitability for a particular purpose.

Hydraulic conductivity Braune *et al.* (2000): A coefficient of proportionality describing the rate at which water can move through a permeable medium. The density and kinematic viscosity of the water must be considered in determining hydraulic conductivity. Vegter (1995): The rate of flow of water through a unit cross section of soil or rock, under unit hydraulic gradient.

Hydrology Vegter (1995): The science that deals with continental water on and under the earth's surface.

Hyporhoeic zone Braune *et al.* (2000): The saturated and biologically active zone in and alongside an alluvial river bed. The hyporhoeic zone is important in river system nutrient budgets as it acts as a nutrient storage system (Valett *et al.*, 1994). It also provides a habitat and refuge for aquatic organisms, thus also serving a buffering function which promotes rapid recovery of aquatic ecosystems after floods or droughts. Winter *et al.* (1998): The subsurface zone where stream water flows through short segments of its adjacent bed and banks.

Infiltration Bowen (1986): Slow movement of water through or into the interstices of a soil. Besbes and De Marsily (1984): Infiltration is defined as that fraction of rainfall which will eventually reach the aquifer. This flux is considered at the moment when it starts its downward migration through the unsaturated zone. Using the Stanford model (Crawford and Linsley, 1966) infiltration is defined as the flux leaving the superficial soil reservoir. [At the same point in space recharge is defined as the flux arriving at the water table, i.e. after percolation through the whole unsaturated zone.] Fetter (1994): The process by which rain will seep into the ground. Venetis (1971): The quantity of water reaching the groundwater table, known also as effective infiltration or percolation.

Instream habitat Braune *et al.* (2000): Includes the physical structure of a watercourse and the associated vegetation in relation to the bed of the watercourse.

Inverse modelling Braune *et al.* (2000): A modelling technique, which uses measurement data as input, for estimating unknown aquifer parameter values within a modelling area.

Interception Carageeg *et al.* (1987): Interception occurs when roads, roofs or vegetation canopies catch rainfall and prevent its passage to the earth's surface. Intercepted water is usually returned directly to the atmosphere by evaporation. The amount of interception depends on vegetation type, canopy cover and rainfall characteristics. Vegter (1995): The process by which water from precipitation is caught and stored on plant surfaces and eventually returned to the atmosphere without having reached the ground.

Land use Braune *et al.* (2000) citing *Water Words Dictionary* (1999): The primary or primary and secondary uses of land, such as cropland, woodland, pastureland, etc. The description of a particular

land use should convey the dominant character of a geographic area, and thereby establish the types of activities which are most appropriate and compatible with primary uses.

Leaf area index Braune *et al.* (2000): This is the ratio of the total area of the leaves on the plant divided by the area of the canopy when projected vertically onto the ground (as though the sun was directly above). The leaf area is measured as the area of the one (upper) side only unless explicitly indicated otherwise.

Leakage Anderson and Woessner (1992): Leakage refers to the movement of water through a layer of material that has a vertical hydraulic conductivity lower than that of the aquifer. Rushton and Tomlinson (1979): Total leakage = baseflow. Carageeg *et al.* (1987): Leakage is the vertical flow of water from one aquifer to another. Its rate depends on resistance to flow in the aquitard separating the two aquifers and on the difference between piezometric heads in the two aquifers.

Leakance Cherkauer and Nader (1989): Leakance = hydraulic conductivity / sediment thickness. Rushton and Tomlinson (1979): Leakance: The flow rate between the aquifer and river is a direct function of permeability and head difference. Walton (1970): Total leakage = baseflow

Lithostratigraphy Braune *et al.* (2000): The science of classifying or organising rock strata according to the properties of the constituent material (Bates and Jackson, 1980).

Management class Braune *et al.* (2000): The present and historical condition of a water resource, its sensitivity and importance, risk of degradation and its potential for restoration are all factors which determine the management class and the resource quality objectives (RQOs).

Management unit Braune *et al.* (2000) citing DWAF (1998): A management unit is the breakdown of ecoregions or hydrological regions according to management requirements.

Marsh Gore (1983): Marshes are non-peat forming; seasonally waterlogged areas.

Mire Gore (1983): The term mire includes all those ecosystems described in English by such words as swamp, bog, fen, moor, muskeg and peat land. According to the *Oxford Compendium*: marsh / / n.(1) low land flooded in wet weather and usually watery at all times; and (2) (attrib.) of or inhabiting marshland.

MODFLOW McDonald and Harbaugh (1983): A modular, three-dimensional, finite difference groundwater flow model, developed by the U.S. Geological Survey.

Numerical model Braune *et al.* (2000): A model of groundwater flow and/or solute transport in which the aquifer is described by numerical equations, with specified values for boundary conditions, that are solved on a digital computer.

Palustrine wetlands Braune *et al.* (2000): These are typical wetlands with emergent vegetation which can range from grasses to forest. Lacustrine wetlands are the natural, permanent or seasonal, freshwater lakes. Riverine floodplain wetlands occur on the floodplains of rivers but do not include the riparian systems along the flowing water channels. Endorheic wetlands form in closed drainage areas and include brackish, saline or alkaline lakes, flats, pans and marshes; they can be seasonal or perennial (Cowan and Van Riet, 1998).

Phreatic surface Braune *et al.* (2000): The upper limit of the saturated zone, i.e. the water table.

Protection Braune *et al.* (2000): In relation to a water resource, means maintenance of the quality of the water resource to the extent that the water resource may be used in an ecologically sustainable way; prevention of the degradation of the water resource; and the rehabilitation of the water resource

Rainfall Braune *et al.* (2000) citing *Water Words Dictionary* (1999): (1) A shower or fall of rain. (2) The quantity of water that falls as rain in a specified area and time interval. Not strictly synonymous with precipitation.

Recharge Bowen (1986): The difference between precipitation and runoff plus other losses, i.e. that part of precipitation which resides as groundwater, surface storage and soil moisture. Rushton and Ward (1979): Groundwater recharge is that amount of surface water which reaches the permanent water table either by direct contact in the riparian zone or by downward percolation through the overlying zone of aeration. It is this quantity which may in the long term be available for abstraction and which is therefore of prime importance in the assessment of any groundwater resource. Besbes and De Marsily (1984): Recharge is defined as the flux arriving at the water table, i.e. after percolation through the whole unsaturated zone. In this definition we assume that the transfer of the 'effective rainfall' through the unsaturated zone is conservative. Average infiltration and average recharge are thus identical over a long period ... Implicit in this definition is the assumption that there is no upward migration from the water table to the soil surface.... This assumption is probably valid in temperate or dry climates, where the water table is deep and the delay between infiltration and recharge therefore significant. Anderson and Woessner (1992): The volume of infiltrated water that crosses the water table and becomes part of the groundwater flow system (Discharge refers to groundwater that moves upwards across the water table and discharges directly to the surface or unsaturated zone). Dinçer (1982): It is necessary to study processes other than infiltration of the precipitation. A recharge possibility through wadi beds which absorb large quantities of flood waters must also be considered. Vegter (1995): The processes involved in the absorption and addition of water to the zone of saturation.

Reference conditions Braune *et al.* (2000): Describe the natural un-impacted characteristics of a water resource. Reference conditions include: Water levels, the seasonal and annual variations thereof. Drought scenarios are also included; Water quality _ the concentration of the key water quality constituents, including their seasonal and inter-annual variability; Groundwater/surface water interaction (including recharge/discharge); rainfall; aquifer structure and integrity.

Regolith Braune *et al.* (2000): Technically a layer of fragmented or unconsolidated rock material, whether residual or transported, of variable character and overlying bedrock (Bates and Jackson, 1980). The authors use regolith in the sense of non-transported material, i.e. from *in situ* weathering.

Remote sensing Braune *et al.* (2000) citing *Water Words Dictionary* (1999): The measurement or acquisition of information of some property of an object or phenomenon by a recording device that is not in physical or intimate contact with the object or phenomenon under study. Also, the utilisation at a distance (as from aircraft, spacecraft, satellites, or ships) of any device and its attendant display for gathering information pertinent to the environment, such as measurements of force fields, electromagnetic radiation, infrared sensing, land use, water bodies, etc. Such systems typically employ

devices such as cameras, lasers, radio frequency receivers, radar systems, infrared detectors, sonar seismographs, gravimeters, magnetometers, and scintillation counters.

Reserve Braune *et al.* (2000): National Water Act (1998): The quantity and quality of water required: to satisfy basic human needs by securing a basic water supply, as prescribed under the Water Services Act, 1997 (Act no. 108, 1997), for people who are now or who will, in the reasonably near future, be relying on, taking water from, or being supplied from the relevant water resource, and to protect aquatic ecosystems in order to secure ecologically sustainable development and use of the relevant water resource.

Riparian Braune *et al.* (2000): Growing by rivers or streams. The Water Act contains the following definition: 'riparian habitat includes the physical structure and associated vegetation of the areas associated with a watercourse which are commonly characterised by alluvial soils, and which are inundated or flooded to an extent and with a frequency sufficient to support vegetation of species with a composition and physical structure distinct from those of adjacent land areas'.

Risk Braune *et al.* (2000): A combination of the probability, or frequency, of occurrence of a defined hazard and the magnitude of the consequences of the occurrence.

Root constant Rushton and Ward (1979): The root constant is the criterion which determines whether or not evaporation will take place at the potential rate. A reduction in the value assigned to the root constant leads to a significant reduction in the actual evaporation estimate during summer months and a consequent increase in the recharge. Shaw (1988): The root constant defines the amount of soil moisture (mm depth) that can be extracted from a soil without difficulty by any given vegetation.

Runoff Bowen (1986): That part of precipitation flowing from a catchment area and finding its way into streams, lakes, etc. Includes direct runoff and groundwater runoff. Total runoff includes both surface runoff and shallow percolation. Carageeg *et al.* (1987): Surface runoff occurs when water is unable to infiltrate and when the ground surface is sloping. Its rate depends on surface slope and roughness, soil moisture content at the surface, as well as on the rates at which additional water is supplied by rainfall and extracted by infiltration or evaporation.

Seepage Todd (1980): Seepage areas indicate a slower movement of groundwater to the ground surface. Seepage is a general term describing the movement of water through the ground or other porous media to the ground surface or surface water bodies. Townley *et al.* (1993): ...the distribution of lake sediments plays a dominant role in controlling seepage through the lakebed...

Specific storage Braune *et al.* (2000): The amount of water released from or taken into storage per unit volume of a porous medium, per unit change in hydraulic head.

Specified yield Braune *et al.* (2000): The ratio of the volume of water a rock or soil will yield by gravity drainage, to the volume of the rock or soil. Gravity drainage may take many months to occur.

Storage coefficient Braune *et al.* (2000) : The volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in hydraulic head. It is equal to the product

of specific storage and aquifer thickness. In an unconfined aquifer, the storage coefficient is equivalent to the specific yield.

Stagnation point Winter (1978): The continuity of the local flow system boundary beneath the lake is the factor that controls the interaction of lakes and groundwater. Further, the point of minimum hydraulic potential below that divide, the stagnation point, is the key to determining the continuity of the boundary. If the stagnation point exists, the boundary is continuous and the lake cannot lose water through its bed because the head at the stagnation point is always greater than the head represented by the lake level. Thus the local flow system associated with the lake is essentially isolated from the rest of the groundwater system. If the stagnation point does not exist, the boundary is not continuous and the lake can lose water through all or part of its bed.

Sustainable yield Caro and Eagleson (1981): The long term rate at which water can be withdrawn without producing undesirable effects such as reservoir exhaustion, excessive pumping costs or water quality deterioration. (Because of the enormous time constants of most groundwater reservoirs, sustainable yield is usually limited by local aquifer transmissivity and storativity, and it bears little relation to the theoretical limit given by the current rate of reservoir recharge.)

Swamp Gore (1983): Swamp has a very wide range of meanings in international usage...it can include marsh and wetland as well. In North America, for example, swamps are usually wooded or forested mires (Heinselman, 1963) *Oxford Compendium*: swamp // n. & v. n. a piece of waterlogged ground; a bog or marsh.

Topography Braune *et al.* (2000) citing *Water Words Dictionary*, 1999): The general configuration of the land surface including relief and position of natural and man-made features.

Transmissivity Braune *et al.* (2000): The rate at which water of a prevailing density and viscosity is transmitted through a unit width of an aquifer or confining bed under a unit hydraulic gradient. It is a function of properties of the liquid, the porous media, and the thickness of the porous media. Driscoll (1986): The rate at which water is transmitted through a unit width of aquifer under a unit hydraulic head.

Transpiration Bowen (1986): The emission of water vapour by living plants into the atmosphere, almost always in daylight. Braune *et al.* (2000): The loss of water vapour from the living cells in the plant through pores (stomata) in the leaves in vapour form. Roberts (1982): The transpiration component of evaporation is influenced by many factors; climate, forest age, species, structure and soil moisture conditions may well, *a priori*, produce large variations in forest transpiration. There is now accumulating evidence that soil moisture levels rarely limit transpiration in field-grown trees. Balek *et al.* (1983): The availability of water in the saturated zone increased significantly the amount of water transpired during the dry periods and after, while the transpiration of the plant of the soil profile without a saturated zone was much lower. Carageeg *et al.* (1987): Transpiration is the process by which plants take up water from the soil and return it to the atmosphere. Its rate depends on vegetation characteristics (species, root depth and density etc), atmospheric conditions and conditions within the soil (especially the distribution of soil moisture near the roots). Some plants, which are known as phreatophytes, have deep roots which intersect the water table and are capable of withdrawing directly from the saturated zone; the transpiration rates of phreatophytes still depend on atmospheric conditions.

Unconfined aquifer Braune *et al.* (2000): An aquifer in which there are no less permeable confining beds between the saturated zone and the groundwater surface.

Water balance Braune *et al.* (2000) citing *Water Words Dictionary* (1999): (1) A measure of the amount of water entering and the amount of water leaving a system. Also referred to as Hydrologic Budget; (2) the ratio between the water assimilated into the body and that lost from the body; also, the condition of the body when this ratio approximates unity; (3) An evaluation of all the sources of supply and the corresponding discharges with respect to an aquifer or a drainage basin.

Water management area Braune *et al.* (2000): An area established as a management unit in the national water resource strategy within which a catchment management agency will conduct the protection, use, development, conservation, management and control of water resources; water resource, which includes a watercourse, surface water, estuary, or aquifer

Wetland Braune *et al.* (2000): 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. Gore (1983): Very widely used to include both mires and marshes. Kotze D (2000): The difference between a marsh and a swamp depends very much on whether you are defining it in American or European terms. Both refer to wetlands subject to extended periods of inundation. In America, such habitats dominated by trees are referred to as swamps and those dominated by emergent herbaceous plants would be referred to as marshes. In European terms swamps may also be herbaceous, particularly tall herbaceous such as papyrus. Depending on the definition peat accumulating wetlands are usually excluded from both 'swamp' and 'marsh'. I prefer the American way of distinguishing between marshes and swamps. *Oxford Compendium*: wetland // n. (usu. in pl.) a swamp, marsh, or other usually saturated area of land. Cowen and Van Riet (1998): (They use the Ramsar definition) areas of marsh, fen, peat-land 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 tides does not exceed 6 m.

Zone of aeration Bowen (1986): Subsurface between the surface and the water table divisible into a belt of soil water, and intermediate region and a lowermost capillary fringe.

Zone of saturation Bowen (1986): The mass of water-bearing ground below the main water table comprising solid rocks and incoherent materials. Vegter (1995): The zone below the water table in which all interstices are filled with groundwater.

Part 2: Case Studies

INTRODUCTION

The conceptual model of groundwater and its relationship to surface water bodies has been presented in a generic sense but has been illustrated very much on the work done in the north-eastern coastal regions of South Africa (Fig. (a)) and parts of Mozambique.

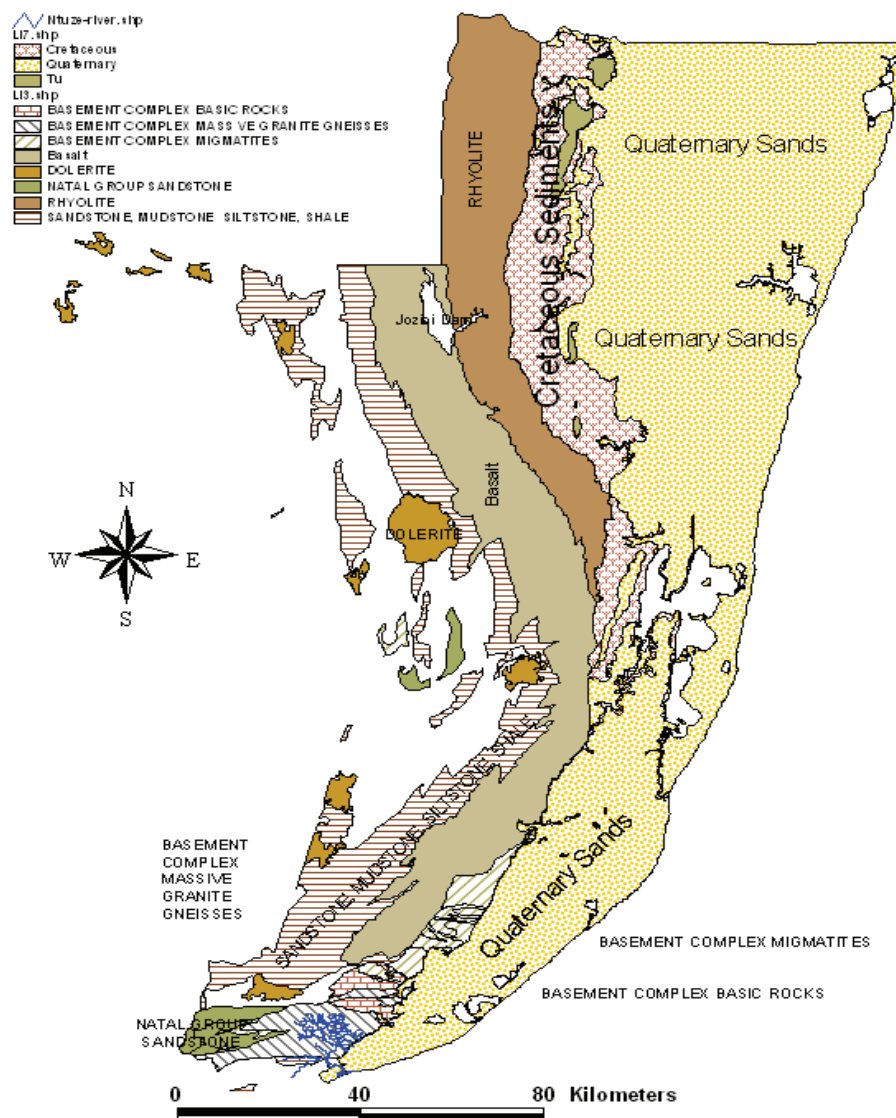


Figure (a): The main geological units of the study area (DWAF, 2000)

Much of the data and derived information in this study have been obtained from primary aquifers in unconsolidated sediments in the low-lying Maputaland Coastal Plain, stretching from Mtunzini in the south to the northern sections of Mozambique. These studies, however, have also been extended to

surface water/groundwater interaction in the Ngoye Hills that are part of the basement complex (Fig. (a)). This part of the report presents a series of case studies that describe specific features of the surface water /groundwater interaction highlighting the role and importance of groundwater to surface water features and methods of analysis. The case studies cover concepts and methods for understanding and estimating the importance of groundwater in rivers, streams, lakes and estuaries in the Maputaland region. The case study sequence starts with a description of a new method of estimating the groundwater component in river runoff using spatial attributes of catchment characteristics. This is followed by a study of the application of groundwater models in deriving the streamflow along a segment of a small stream in Richards Bay that is dominated by groundwater. The next case study looks at the influence of land use on the role and importance of groundwater (baseflow) in coastal rivers and estuaries. These are followed by case studies of the recharge in primary aquifers and a look at the role of tracers in understanding the groundwater/surface water interaction. The next two case studies look at the role and importance of groundwater in estuaries and coastal lakes.

To understand the context of these studies, the first part of this section will describe the main features of the region and their role in the ensuing discussions in the case studies.

Geology of Maputaland/Zululand

The geology of the Maputaland study region is portrayed in Fig. (a). The first case study is located in the basement complex massive granite Gneisses on the Ngoye Hills and the upper reaches of the Umlathuze catchments. The case studies have been done in the Quaternary Sands of the Maputaland complex. A vertical transect through these systems is shown in Figs. (b) and (c).

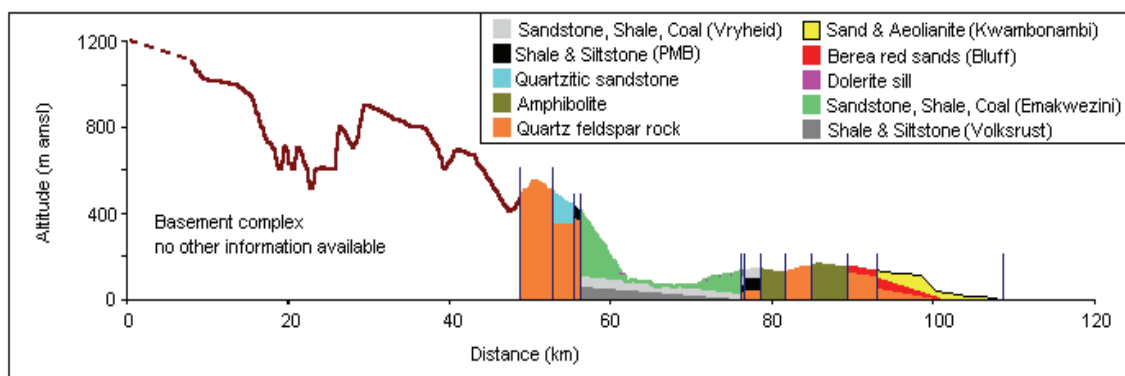


Figure (b): Transect through the Mhlathuze catchment derived from various sources

The hydrogeology of the region has been described by Meyer and Godfrey (1995) and Kelbe *et al.* (1995). Detailed investigations have also been conducted in various locations in the region by many other investigators.

The hydrogeological units of relevance to the Maputaland studies are the marine, alluvial and Aeolian sediments overlying the Cretaceous. All the interactions between the groundwater and surface water are linked to this primary aquifer. The role of groundwater is determined by the physical boundaries with the surface water. These include the recharge and discharge boundaries as well as the hydraulic boundaries within the hydrogeological units.

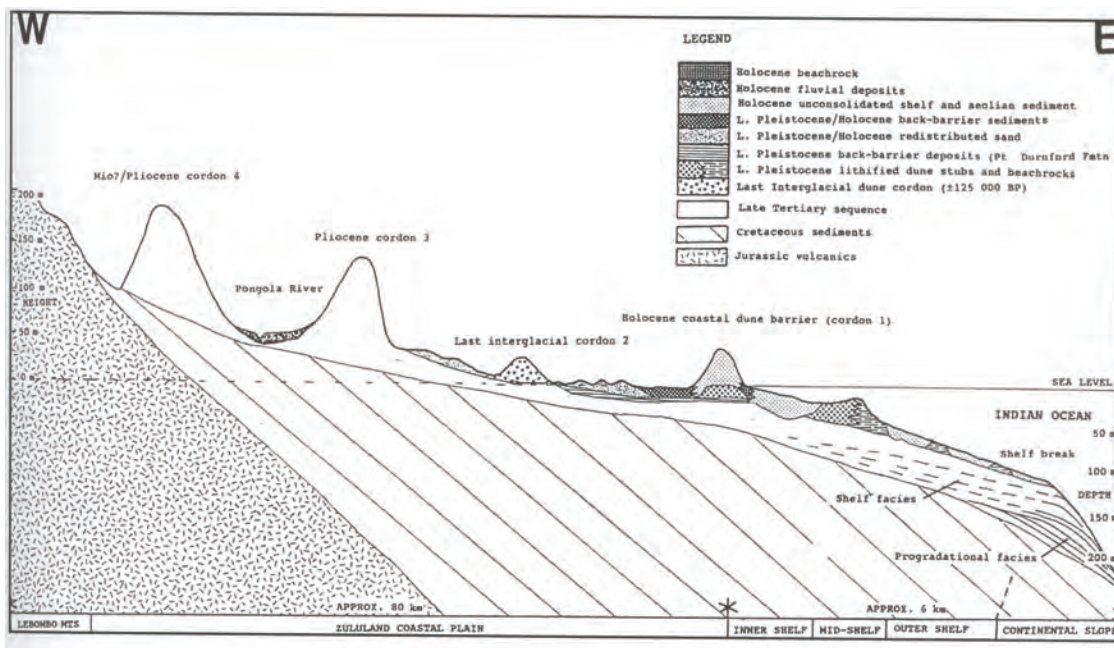


Figure (c): An idealised geological cross-section through the northern KwaZulu-Natal coastal plain taken from Wright (1995) and Ramsay (1996) Geological Survey reports

Water Resource Features of the Study Area

The surface water resources form principal recharge and discharge boundaries for the primary aquifer. The Zululand region has a relatively high rainfall for South Africa and consequently it is generally well endowed with water resources. Several major rivers flow from the interior escarpment and Lubombo Mountains toward the Indian Ocean through many of the important wetlands and estuaries. Nearly all the rivers feed into estuaries that are important breeding grounds for marine fisheries (Maree *et al.*, 2003). However, many small streams feed into wetlands that are scattered throughout the Maputaland Coastal Plain. The largest of these wetlands with no outlet is Lake Sibayi, the largest inland lake in South Africa.

The rivers, lakes and wetlands along the Maputaland coastal plain are predominantly groundwater driven. However, the general picture is one of large interior rivers draining through the coastal plain into the estuaries although there are many small streams that drain directly into the sea (see Case Study 2). These rivers and wetlands on the coastal plain are almost entirely dominated by groundwater. This section describes the various resources and attempts to show the linkages between them.

Rivers

The main rivers of the Maputaland coastal plain are the Mhlathuze, Umfolozi, Mkuze, and Pongola. The Mhlathuze flows into the Richards Bay harbour/sanctuary. The Mkuze River, together with three smaller rivers (Mzinene, Hluhluwe, and Nylazi) flows into Lake St. Lucia while the Pongola flows northward into Maputo Bay, Mozambique. The Umfolozi was once part of the St. Lucia wetland but has been separated by channelization of the lower section to protect farming enterprises. An Ecological Reserve determination has been conducted on the Mhlathuze River below the Goedertrouw Dam and the

management class is presently under review. A desktop RDM for Lake St. Lucia was done in early 2004 and a GWRD for the same region is currently underway.

No flow separation studies have been conducted on the Maputaland rivers with the exception of the recent attempts by Snyman and Kelbe (see Case Study 1). This study suggests that as much as 80% of the average river flow is derived from groundwater.

The first case study applies a methodology derived by Snyman (2005) to estimate the groundwater contribution to the river runoff in the Mhlathuze above the Phobane Lake (Goedertrouw Dam). The second case study illustrates the use of a groundwater model in estimating the runoff from a segment of a small stream feeding directly in to Richards Bay Harbour.

There are numerous small streams and rivers scattered along the Maputaland coastline and shoreline of all lakes and wetlands. These streams are almost entirely groundwater driven and many of them play a fundamental role in sustaining features and functions of the wetlands and lakes. The case study on Lake St. Lucia highlights the role of these streams in sustaining the ecological resilience of the lake. Another case study describes the loss of estuarine function because of the extreme impact of deforestation on a coastal river.

Estuaries

Six of the Maputaland estuaries (Fig. (d)) consistently lie in the top seven rankings for estuaries in terms fish importance (Maree *et al.*, 2003). The Maputaland estuaries make the number one ranking for fish, botanical and waterbird prioritization (Maree *et al.*, 2003).



Figure (d): The Maputaland estuaries

The main Maputaland estuaries have been ranked consistently high in terms of their importance, relative to all the 248 estuaries in SA, in relation to seven criteria (Table 1). This highlights the importance of these systems for the country. Unfortunately, the groundwater contribution to these systems is only recently being identified as an important feature in the sustainability of the system. While it was recognised that groundwater played a crucial role in the Ecological Reserve for Lake St. Lucia at the recent RDM (Van Niekerk, 2005), no formal scoring was afforded to the groundwater contribution. Case Study 3 describes the role of groundwater in sustaining the ecological resilience of Lake St. Lucia. The estuaries of the Maputaland coastal plain form a vital ecological link between the rivers and the sea. This link is part of temporary open/closed systems that creates large diversity in the physical conditions of the estuaries is not well understood. The case study for St. Lucia reveals the very important role that groundwater plays in sustaining the ecological resilience of the system. However, St. Lucia may not be representative of other important estuaries in South Africa and there is desperate need to understand the role of groundwater in these other dynamic systems. St. Lucia groundwater

dependency is described in more detail in the third case study. The Kosi System has no large rivers draining into the lakes and it is dependent entirely on groundwater and direct rainfall for its freshwater inflow.

Table 1.1: The importance score for Maputaland estuaries, ranked out of 248 of the South African estuaries (after Maree *et al.*, 2003).

Estuary	Total score	Species richness	Endemic richness	Exploitable richness	Estuarine type	Estuarine condition	Estuarine size	Estuarine isolation
Matigulu	3	9	53	9	1	75	26	29
Siyaya	176	9	53	9	60	154	211	56
Mlalazi	4	7	53	7	1	75	33	29
Mhlathuze	7	7	110	7	1	154	9	36
Richards Bay	6	4	110	1	1	154	8	49
Nhlabane	53	4	110	1	1	154	169	17
Mfolozi	75	4	110	1	238	154	133	24
St. Lucia	1	1	8	1	1	75	1	7
Mgobezeleni	16	3	8	1	1	75	211	5
Kosi	2	1	8	1	1	75	5	14

Wetlands

In the NWA, the wetlands are considered separate resources from the rivers and estuaries. They are the transitional zone between the terrestrial and groundwater aquatic systems where the water table is at or near the surface. The US Fish and Wildlife Service (UAFWS) classify wetlands as either marine, estuarine, riverine, lacustrine or palustrine. Alternatively, they can be defined as any ecosystem whose soils show evidence of at least periodic water logging (Davies and Day, 1998). The riverine systems have been discussed and a case study is presented. The lakes and estuaries are considered separate systems and are discussed in the next sections.



Figure (e): The main wetlands and lakes of Maputaland

The Maputaland coastal plain is covered by many wetlands that are either fed by the various rivers and streams or they are an extension of the water table or governed by the groundwater. Consequently, wetlands are considered to be critically dependent on groundwater in the Maputaland region.

The groundwater is the principal source of water for most of the wetlands in Maputaland. On the eastern shores of St. Lucia, many wetlands within extensive plantations of pines were slowly destroyed

by the slow drawdown of the water table (Rawlins, 1991). These plantations have recently been removed and there were many places where the residual tree stumps were found lying in half a metre of water once the water table had recovered. Long term monitoring of the region is being analysed for confirmation of the groundwater recovery following the deforestation on the eastern shores.

Lake St. Lucia mouth has been closed since 2001 and the lake has receded to the point where it has become compartmentalized. This has exposed large beaches along the shoreline that have enabled studies of the groundwater seepage. A section of shoreline in 2003 was studied and the groundwater seepage mapped. Figure 1.5 shows the large freshwater sheet flow draining from the old shoreline. The small flow through a breach in the foreshore dunes was measured at $10 \text{ m}^3/\text{d}$ during this period of severe drought.

The small stream (breach in the shoreline ridge) believed to have been created by a hippo trail, has continued to flow during the drought. An ad hoc measurement of flow from the small seepage zone in the shoreline shown in Fig. (f) showed it to be about $10 \text{ m}^3/\text{d}$. This was repeated in May 2005 and the flow was measured at over $4\,000 \text{ m}^3/\text{d}$ after a day of rainfall during average rainfall conditions over the past few months.



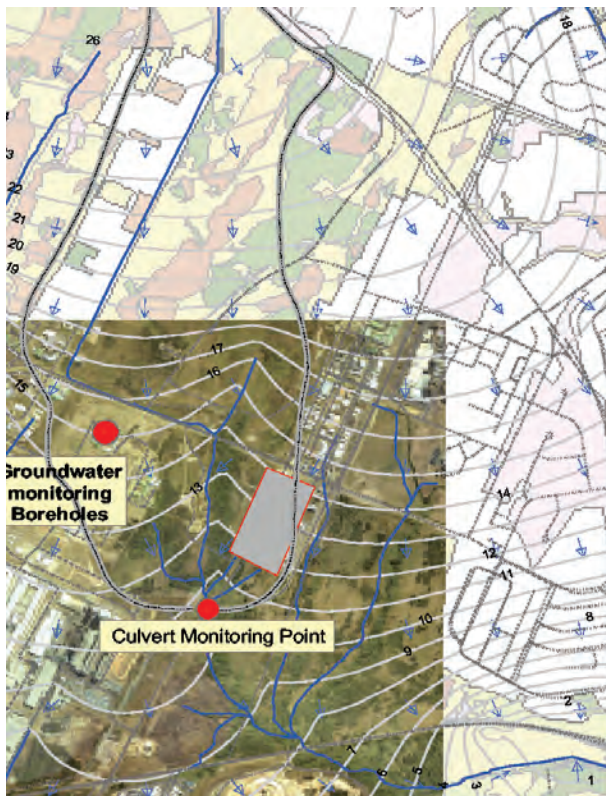
Figure (f): Groundwater seepage mapping on the shoreline of Lake St Lucia when the lake was receding during a severe drought and the estuary mouth was closed

A small stream draining a wetland in the middle of Richards Bay (Alton) was monitored over several years during a period of below normal rainfall. The flow in this stream, with an estimated channel length of 4 km and a recharge area of approximately 4 km^2 , never ceased to flow. The hydrograph for selected storms in this stream are compared to the groundwater hydrograph at monitoring borehole with the catchment. Figure (g) shows the location of the stream and monitoring points. The link between groundwater and streamflow are discussed in Case Study 4 through hydrograph analysis of streamflow and water table fluctuations.

Coastal Lakes

The lakes of Maputaland are considered separately to the other wetlands in this report.

The lakes of the region can be classified in many different ways depending on their linkage with the sea, the form and magnitude of the surface outflow (river), the form and size of the river inflow and the extent of the groundwater seepage. For this study, the lakes have been classed into four broad categories comprising the coastal inland lakes with river outflow (Mzingazi, Cubhu) and without river outflow (Sibayi), Lakes with an estuary (St. Lucia, Kosi, Nhlabane), and the off channel lakes found on alluvial flood plains of the Mfolozi and Mhlathuze rivers. Kelbe and Germishuyse (2000a) also separated out Nsezi as a lake on the boundary between the coastal plain and fed by substantial inland catchment (Nseleni River). This lake is not considered in this report.



Lake Sibayi has been studied in detail by Miller (2001). Preliminary efforts to map the surrounding aquifer that forms part of the lake have been presented by Meyer and Godfrey (1995). This lake formed many thousands of years ago during the regression and transgression of the sea. The mouth closed permanently and the lake became completely isolated and elevated above the sea. Meyer and Godfrey (1995) estimated that the seepage from the lake into the ocean was of the order of $1 \times 10^6 \text{ m}^3/\text{km}$.

Figure (g): The stream and groundwater monitoring points in Alton, Richards Bay

The Lake St. Lucia water balance under normal climatic conditions is dominated by direct rainfall and river runoff from the four major rivers (Mkuze, Mzinene, Hluhluwe, and Nylazi) feeding into the northern parts of the lake (see Case Study 4). Several smaller groundwater dependent streams feed into the lake and estuary at various points around the lake (Mpate, Nkazana). Under average climatic conditions, the groundwater contributes less than 10% of the freshwater inflow to the lake water balance. Under severe and prolonged drought conditions, the lake oscillates between a freshwater and a highly saline system. Under these drought conditions, the groundwater seepage may be the only source of freshwater and is considered essential for the ecological resilience of the lake in the frequent transitions from saline to freshwater states (Kelbe *et al.*, 2005). Case Study 4 describes in more detail the important role that the groundwater plays in sustaining the ecological resilience of the lake to respond to these large fluctuations in salinity.

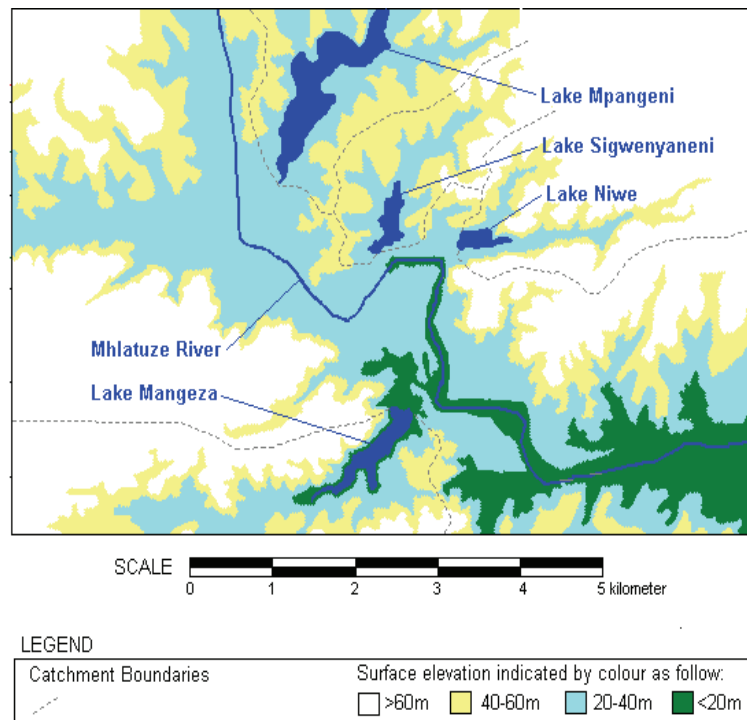


Figure (h): Off-channel lakes that formed on the small tributaries of the Mhlathuze River

Lake Mzingazi in Richards Bay was formed when the link to the Richards Bay Estuary was broken several thousand years ago. The lake level has been artificially raised to increase the yield as part of the Richards Bay water supply. Case Study 5 examines the level of groundwater interactions and highlights some of the complexities in ascertaining the level of dependence on groundwater.

There are numerous examples of off-channel lakes in the Maputaland rivers. The main ones on the Mhlathuze River are shown in Fig. (h).

Lake Mangeza was used for municipal water supply until about three years ago for the peri-urban community in kwaDlangezwa and Vulindelela just south of Empangeni.

For a short period, Lake Mpangeni was used by Empangeni to supplement its water supply. Similar lakes occur in the low-lying sections of the Umfolozi River which are also used for community water supply schemes. A postgraduate study is underway to evaluate the role of groundwater in one of the off-channel lakes (Lake Mangeza) in the Mhlathuze system.

These off-channel lakes are created by the deposition of sediments in the main river channel during high floods which then block the tributaries that feed the river in the lower sections of the catchment. These tributaries have insufficient flow to erode these flood plains and they eventually establish stable bars between them and the main river channel. There is a small overflow when the lakes are sufficiently full to discharge. During high flood events, the main river channel is normally sufficiently full to cause back-flooding and hence reduce the potential for breaching the sand bar that forms the lake.

There are many small pans scattered throughout the Maputaland coastal lake. These pans are usually shallow systems and associated with large flood plains. Because they are shallow and often perched above the surrounding aquifers, they do not have a strong groundwater dependency. However, several of

the deeper ones may be fed by substantial groundwater seepage. No studies of the groundwater regime of these pans have been found in the literature.

Recharge

The groundwater recharge rate for the Maputaland region has been described in various studies by Bredenkamp *et al.* (1995), Meyer and Godfrey (1995), Kelbe and Rawlins (1992a) and Rawlins (1991). The processes involved in groundwater recharge have been described in previous sections. The approach preferred for the simulation of the groundwater system in the primary aquifers of Maputaland is not well received in the general applications of groundwater studies. The preferred program for evaluating groundwater systems has been the USGS finite difference MODFLOW code. This and similar programs used in previous studies by Kelbe and Rawlins (2000a) called Intersat incorporated the loss of water from the saturated zone through the evaporating surface using a simple linear response function. The evaporation rate was scaled over the rooting depth of the vegetation. To incorporate this land use effectively, it is necessary to also include the full percolation rate. Kelbe and Rawlins (2000a) observed a response in the groundwater table in the St. Lucia aquifer when the rainfall events exceeded 10 mm over a 5 d interval. They equated this to a soil moisture depletion rate of 2 mm/d due to evaporation from the unsaturated zone that was about 2 m to 5 m deep. It was further assumed that all percolating water below the rooting depth of the vegetation would pass through the underlying unsaturated zone and into the groundwater because there was no evaporation below the rooting zone, only free drainage.

The Kelbe and Rawlins (2000a) model used the total rainfall for recharge after accounting for the 10 mm initial loss taken over 5 d. They also assumed an upper level of infiltration that was dependent on the surface conditions and the depth to the groundwater. Using this model and applying an adjusted potential evaporation rate for groundwater, they were able to show that the net recharge was similar to the more conventional estimate of other researchers.

Germishuyse and Kelbe (1999) used a binomial filter on the recharge series to create stress periods with relatively uniform recharge rates in the case study for Lake Mzingazi (Case Study 4). However, it was necessary to identify specific events that had a significant effect on the groundwater, such as cyclonic systems. However, Germishuyse and Kelbe (1999) showed that the delineation of stress periods had an influence on the simulations and that there was a need to evaluate the groundwater response.

Climate

The rainfall, and hence the recharge in the Maputaland region is generally high, reaching annual rates of over 1 000 at the coast but declining to less than 800 in the interior. With highly permeable soils, high rainfall, and low water gradients, the groundwater is expected to move rapidly through the system and reflect a young chemistry. Three biennial samples from 2001 to 2004 at three depths in Richards Bay indicate an increasing TDS with depth that suggest that there is restricted vertical mixing.

Regional Groundwater

Meyer and Godfrey (1995) presented a map of the piezometric head for Maputaland derived from assumed constant head boundary conditions for Lakes Sibayi, Kosi, St. Lucia and Mzingazi and 140

borehole observations. More localised studies have been conducted in specific areas of the region. In particular:

- Kelbe and Rawlins (1992a) simulated the water table elevation for the eastern and western shores of Lake St. Lucia region
- Kelbe and Germishuys (1999c) derived the piezometric head for the Richards Bay area described in detail in Case Study 5 (Mzingazi)
- Kelbe and Rawlins (1999c) modelled the piezometric head for the southern end of the Maputaland coastal plain at Mtunzini

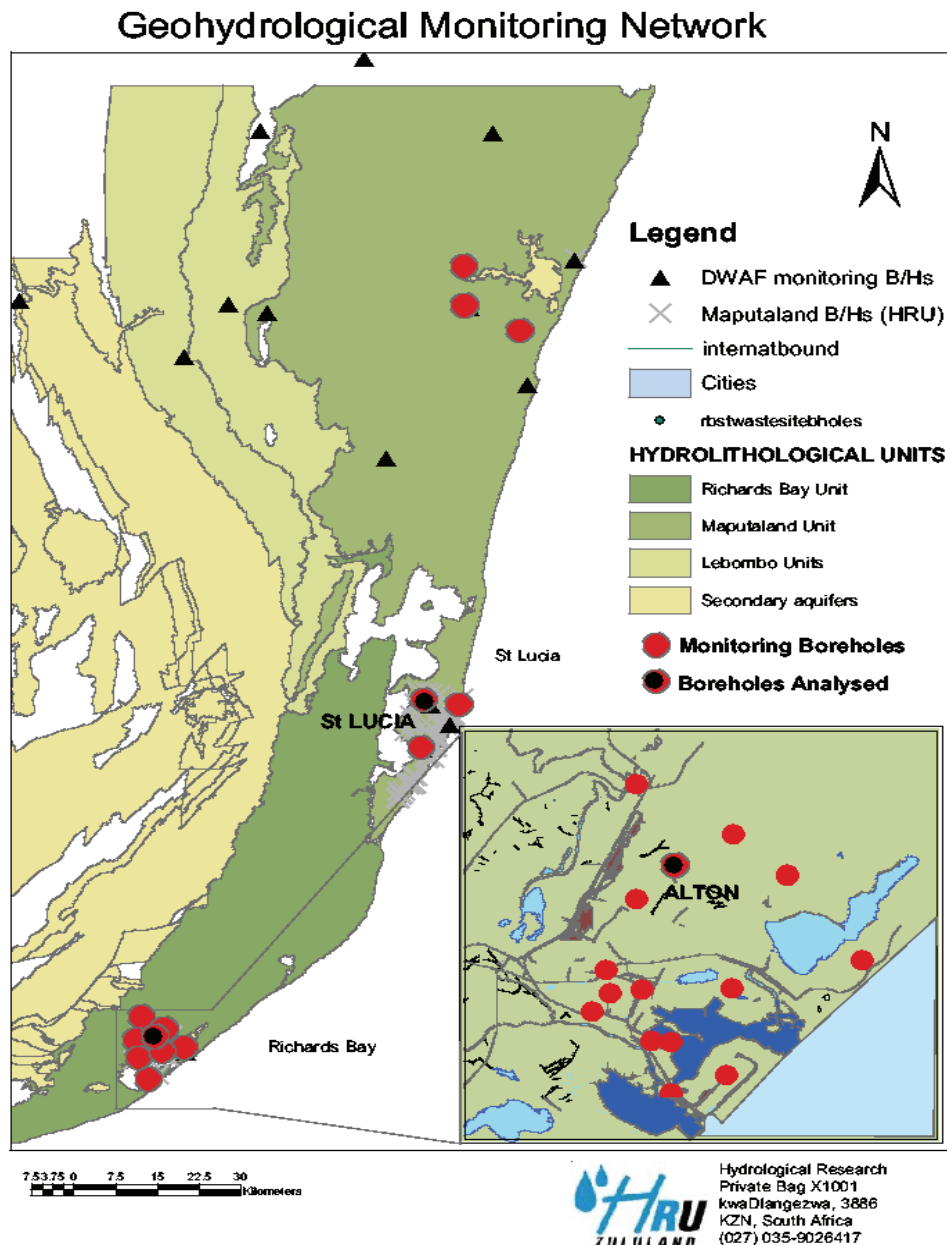


Figure (i): The location of monitoring sites for case studies in the Zululand region described in subsequent section for demonstrating specific aspects of surface-groundwater interactions. Additional monitoring sites in Northern Mozambique are not shown.

CASE STUDIES

The first case study describes a new approach to the simulation of groundwater contribution to river runoff using spatial methods. This is followed by a demonstration of the use of groundwater models to simulate the flow in stream segments in groundwater-dominated streams in the Maputaland Coastal Region. The use of groundwater models is then demonstrated in several case studies involving land use and estuaries, lakes and streams.

Case Study 1: Simulating River Runoff Components Using Spatial Modelling Techniques

This case study describes the development and application of a new technique to estimate the runoff sequence for the various flow components in a river, with particular reference to the groundwater component.

Case Study 2: Investigation of the Application of a Groundwater Model for Stream Discharge Estimation in Regions Dominated by Shallow Primary or Alluvial Aquifers

This case study looks at the application of a 3D finite difference model in a groundwater dependent system for simulating the base flow component of streamflow.

Case Study 3: Land-Use Impact on the River Flow and Estuary of the Siyaya System

This case study demonstrates the effect of land use on sediment production and water consumption that has a large impact on river and estuary functions in a small catchment on the east coast of South Africa. Sugar cane production up to the start of the 1980s led scientists to believe that the riparian function of the tributaries of the Siyaya Estuary had been severely impacted by farming practices. In the 1990s nearly all the sugar cane was replaced by *Eucalyptus* plantations. This case study looks at the effect of deforestation on the water yield of these catchments.

Case Study 4: Estimation of Groundwater Contribution to River Flows in Maputaland using Hydrograph Analysis Techniques

A previous investigation of hydrological flow components has identified several flow components in river runoff from a small research catchment in the Zululand region. Using recognized flow-separation techniques and natural tracers, these studies identified different flow components in the runoff from small rivers in non-alluvial groundwater systems. This case study demonstrates the application of flow-separation techniques in the Ntuzi Research Catchment and Richards Bay. The results of this case study formed the basis of the spatial modelling technique for separating the flow components of river runoff in the larger Mhlathuze system described in detail in the main text of this report.

Case Study 5: Groundwater Recharge and Discharge Features for Shallow Primary Aquifers in Maputaland

This case study looks at the discharge characteristics and groundwater response to storm events in a small catchment in Richards Bay that is representative of most of the wetlands in the region. A storm hydrograph is compared to the groundwater hydrograph for the same catchment.

Case Study 6: The Development of a Groundwater Model for Evaluating the Interaction of a Coastal Lake and Feeder Streams with the Underlying Aquifer

This case study looks at the development and application of a three dimensional transient groundwater model for evaluating the interaction of the lake and streams with the surrounding aquifer.

Case Study 7: The Importance of Groundwater in Sustaining the Ecological Function of the St Lucia Estuary

This case study looks at the role of groundwater seepage into Lake St. Lucia and its function in moderating the effects of salinity fluctuations that play an important role in the ecological resilience of the system.

CASE STUDY 1: SIMULATING RIVER RUNOFF COMPONENTS USING SPATIAL MODELLING TECHNIQUES

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INTRODUCTION

The South African Water Act of 1998 has introduced new regulations that required a greater understanding of the form and processes of hydrological systems and interactions. New hydrological methods needed to be developed to meet the implementation of the law. This case study presents a model that uses spatial information of a catchment to estimate the components of river runoff from a small head water catchment at the southern extremity of Maputaland. The model is built on the concept that there are a finite number of preferential flow paths down the hill slope to the catchment outlet that are frequently detectable in the runoff hydrographs of small head water catchments and coastal streams. The research catchment of the Ntuzze River in the Ngoye Hills (Fig. 1.1) has been used by Snyman (2005) to develop the model and validate the assumptions using runoff hydrographs from three monitoring weirs (W1M016, W1M01 and W1M031).

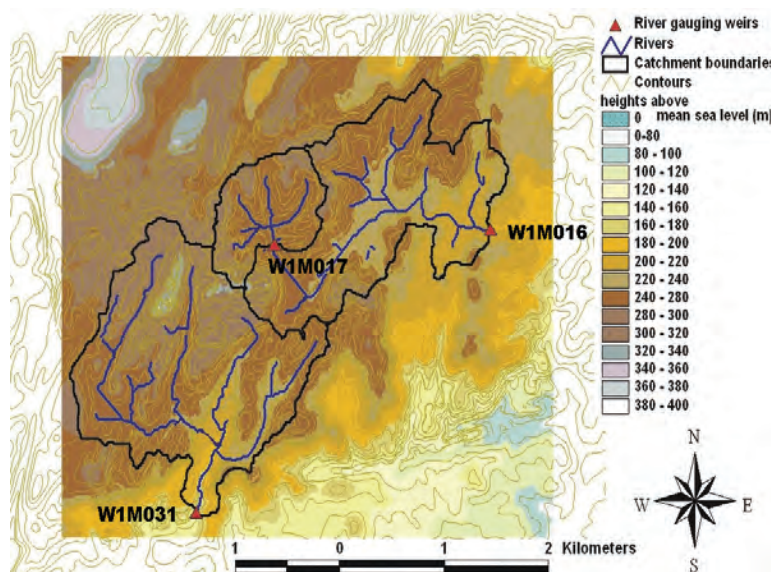


Figure 1.1: The Ntuzze research catchments and monitoring weirs

The Ntuzze River research catchments are small enough (maximum 3.2 km² each) to represent headwater conditions where the hydrological interactions have not overlapped to the extent that they have become lost in the mixing processes of river runoff derived from different contributing areas and processes.

Conceptualization of the River Flow Components

The movement of water originating as rainfall on a hill slope can travel an infinite number of pathways to reach the catchment outlet. For an impervious catchment surface, there would be a finite

number of pathways determined by the direction of the steepest slope for any point on the surface. The shape of the runoff hydrograph for short duration and high intensity storms in the Ntuzi catchment are closely related to the catchment geomorphology (Mulder and Kelbe, 1993).

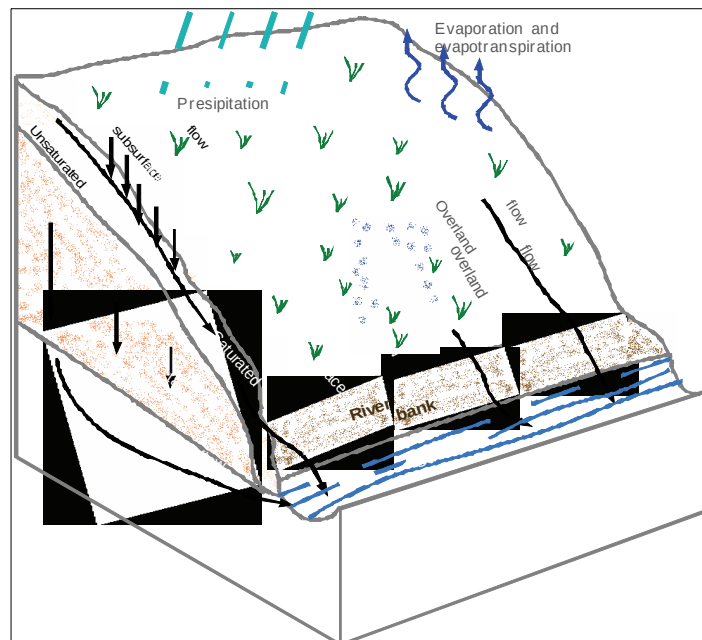


Figure 1.2: A catchment diagram of the flow component concepts.

However, the runoff process is seldom two dimensional as there is always some infiltration and percolation leading to a fully three dimensional process involving an infinite number of pathways. Nevertheless, Beven (2001) describes many of the more common flow processes that are likely to lead to preferential grouping of pathways (Fig. 1.2).

While it is well recognized that there are an infinite number of possible pathways that rainwater can travel to a discharge point in a catchment, the hydrograph analysis indicates that there are a finite number of preferential pathways that the water tends to follow down a hill slope. These are often described in terms of their transit times through the hill slope. The flow of water in an open channel is usually much more rapid than flow over or through the soil matrix because of low resistance to motion. On the other hand the slow transit times for flow that percolates down into the groundwater and then discharges as baseflow follows a tortuous pathway through the soil structure that is invariably longer and more resistive than the surface routes.

The analysis of observed streamflow of headwater catchments by Kelbe and Germishuys (1999a) indicates that the observed hydrograph in the small headwater catchments of the Ntuzi generally consists of three distinct flow components that are depicted in Fig 1.3. During dry periods with little or no rain, the flow in streams and rivers is derived entirely from groundwater. This flow is known as baseflow. The amount of baseflow in a river varies radically from region to region and within regions that depends on the storage capacity of the soil profile. Case Study 2 gives an example of the flow in a very small catchment in Richards Bay that has seldom stopped flowing even during severe droughts. Much

larger rivers in Maputaland (Nyalazi, Mzinene, Mkhuze, and others) have been known to cease flowing during the same periods.

There is a substantial difference in the groundwater contribution to these different rivers.

Following storm events, there is a large change in flow patterns when rainfall follows different pathways to the catchment outlet (Fig 1.3).

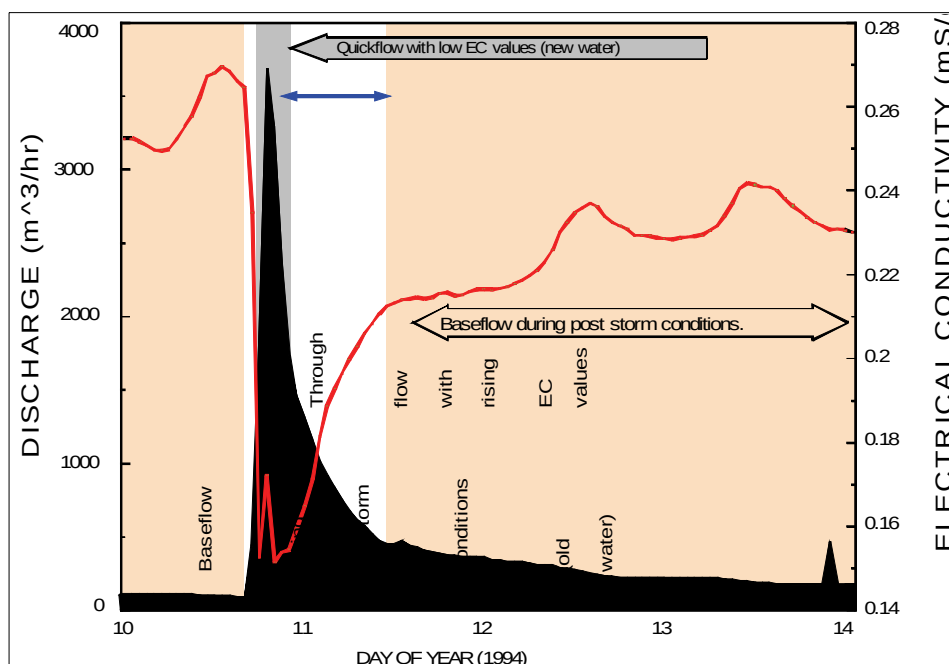


Figure 1.3 Electrical conductivity indicates the existence of the through-flow component in the runoff (Kelbe and Germishuyse, 1999)

The quick-flow component of storm flow is regarded as the water that flows fairly quickly through the catchment surface layers into channels and runs down to the catchment outlet along a 'surface' route, causing the sharp high peak in runoff after a short duration, high intensity storm. There is nearly always infiltration where the unsaturated subsurface contribution is represented by a delayed flow that moves more slowly through the top layers of the soil and back to the surface or into the channel flow. It is often referred to as through-flow, intermediate flow or macropore flow, as it is often associated with flow through preferential channels (or macropore holes in the soils) in the unsaturated soil structure. The water that percolates deeper into the soil to reach the saturated zone moves more slowly through the catchment to reach the catchment outlet, as it follows longer, more arduous paths through the saturated zone. The main flow components and the way they are viewed and conceptualized in this study, are summarized in Fig. 1.4 and described in more detail in the following paragraphs.

The three flow processes are frequently observed in most of the streams and rivers as well as shallow water tables in the Maputaland regions. An example is shown in Fig. 1.3 for runoff in the Ntuzi Catchment (W1M016). Figure 1.3 shows the rapid rise and fall of the quick-flow component that coincides with the drop in TDS when new water enters the stream channel. The throughflow is characterised by a slower recession in the flow rate for the first half of DOY 11 when the TDS starts to rise as new water is mixed with and/or displaced by old water. The quickflow and throughflow cease after about 18 h leaving baseflow (old water) where the TDS is slowly starting to return to pre-storm

conditions. The lower TDS for the post-storm baseflow suggests that there is some mixing of old and new water in the groundwater (saturated zone).

In the development of the three independent component hydrograph models, it is assumed that the shorter flow paths will reach the discharge point earlier than the longer flow paths. Consequently, the fluxes reaching a common discharge point in the stream channel at a particular time will depend on the length of the preferential pathways.

The shape of the hydrograph, following a uniform rainfall event over the catchment, will be determined entirely by the combined fluxes of all the contributing flow paths at each instant in time. Hence, the shape of the component unit hydrograph is determined entirely by the frequency (number) of flow paths discharging at each interval of time for that preferential pathway. The unit storm hydrograph (SH) will then be formed by the convolution of all three component unit hydrographs (UH). The unit storm hydrograph would then be scaled according to the actual fluxes along each pathway to generate the actual storm hydrograph.

The pre-event water represented by TDS in Fig. 1.3, which contributes to the observed storm hydrograph as baseflow, indicates that the water in the storage reservoirs of the catchment is displaced quite quickly by “new” water with low TDS, despite the slower flow velocities of subsurface water. The explanation for this contradiction is in the physics of the flow processes in the saturated zone (Beven, 2001). This phenomenon can be attributed to the wetting front in the unsaturated soil profile trapping air which effectively creates an artificial head in the saturated zone to increase the discharge rate due to a pressure wave. This pressure wave is ‘translated’ to the rest of the saturated zone very quickly. The theory proclaims that very small disturbances will propagate very quickly, and that larger disturbances will have smaller wave velocities (the magnitude of this disturbance is a function of the inverse of the effective storage capacity in the soil, the effective storage capacity being the difference between the current soil moisture and the soil storage immediately above the water table). This simply means that, in a wet catchment, the wave velocity may be much faster than the actual flow velocity of water. This is investigated in more detail in Case Study 4.

Conceptual Catchment Flow Processes

The hydrograph analysis of runoff from the Ntuzi Catchment as well as small streams in Richards Bay suggest that the infinite number of flow paths are often associated with preferential grouping that lead to only a few conceptual flow components. In particular, flow over an impermeable surface as often found in urban areas, can be effectively conceptualized as a 2D flow plane where there is no infiltration. This is indicated in the schematic diagram in Fig. 1.4a as the upper surface plane. This concept can be extended into a quasi 3D model (Fig. 1.4) where the infiltrating water is partitioned into three preferential pathways that follow deterministic flow planes that lead to the three component storm hydrographs.

The surface flow pathways of quickflow are usually the fastest moving flow components and will reach the catchment outlet first. Along these pathways, the flow path and flow rate are dependent on geomorphic features (slopes and aspects). The travel velocity along these flow paths is also a function of the resistance to flow (surface roughness and the hydraulic depth of flow).

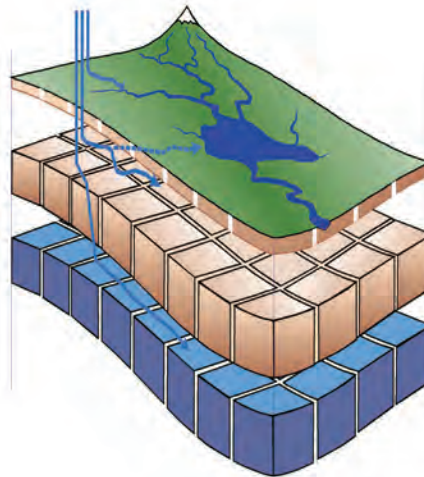


Figure 1.4: Representation of quasi 3D flow plane where three preferential pathways dominate the flow pattern (adapted from Abbot et al.)

The travel velocities for surface flow have been estimated by Kelbe and Snyman (1996) using the Manning equation, for a uniform rainfall on a small catchment in the Ngoye hills. They assumed that:

- The land use indicates the surface roughness (indicated by Manning's n),
- There is a representative depth of sheet flow (R) for a unit of rainfall for each unit area (pixel), and
- The slope of the unit area indicates the slope of the surface flow

Manning's equation determines the flow velocity of channel flow (V) described by the following equation (Chow *et al.*, 1988):

$$V = \frac{1.49}{n} R^{2/3} S^{1/2}$$

where:

n = Manning's coefficient (a roughness coefficient)

S = slopes of the channel

R = the cross sectional radius of the channel or hydraulic depth. The hydraulic depth of flow over a unit area will be dependent on the contributing area of the catchment above the unit area

Throughflow is assumed to be derived from matrix flow and from the movement through the macropores in the soils. The matrix flow can be described by Darcy theory (Todd, 1980) while the macropore flow is better represented by surface flow processes described by Manning's model. The flow rates of the through-flow are assumed to be a composite of the subsurface unsaturated flow matrix and saturated subsurface flow inside the macropore channels. These combined flow processes would be represented by the middle plane in Fig. 4.

Baseflow travels through the soil structure until it reaches the saturated zone, from where it follows the paths of least resistance down the saturated hydraulic gradient. The travel velocity is dependent on the soil permeability and the hydraulic gradient of the groundwater and can be described by Darcy's Law (Todd, 1980). The Darcy velocity (specific discharge) is directly proportional to the hydraulic gradient

($\partial h / \partial L$) of the drainage surface and the hydraulic properties of the soil (the hydraulic conductivity K) according to the following equation:

$$Q = AK \frac{\partial h}{\partial L}$$

where:

Q = the specific discharge
A = cross sectional area
K = hydraulic conductivity
h = difference in vertical height
L = difference in horizontal length

The three different flow components of river flow are distinguished by their different travel paths and the different travel velocities. However, a certain amount of mixing of the three components takes place during and even before the water flow down the river course.

This conceptual model describes three separate two-dimensional flow paths (planes) that, individually, would produce their own characteristic hydrographs. The actual storm hydrograph would be represented by the convolution of all three hydrographs.

This model is applied along flow paths using finite difference concepts defined by the scale of the application. Consequently it is also necessary to introduce a scaling factor in the application of these equations. With the groundwater, there is the added issue of pressure wave propagation that leads to faster travel times described as pressure induce flow (Beven, 2001). These variations on the traditional approach to flow rates are incorporated into a scaling factor that is applied to the resistance parameters in the various equations. Snyman (2005) has applied a power law to the resistance factors by multiplying the exponent by a factor 'm' that is derived from calibrations for the scale of application. This model is used in this study.

Convoluting Hydrographs

The three flow planes/routes are extreme simplifications of reality where the actual flow is represented by an infinitely large number of streamlines. The conceptualized flow path is expected to represent a range of streamlines that exhibit a characteristic (mean) travel length and time scale. A schematic representation of the three conceptualized flow components in a river flow derived from the quasi 3-D model are shown in Fig. 1.4 for the three sub-catchments of the Ntuzi River.

The overlapping of flow patterns/hydrographs belonging to the different flow components becomes more evident in larger catchments than smaller catchments, so that the different flow components cannot be observed in the river's hydrograph. Not only does the water from different flow components mix, but also water from different sub-catchments. This case study describes the development and application of a spatial flow model for separating the flow components of river runoff for any size catchment on the basis of the geomorphic response function that is used to determine the travel rates along the preferential flow paths and their ultimate integration into a storm hydrograph for different rainfall scenarios.

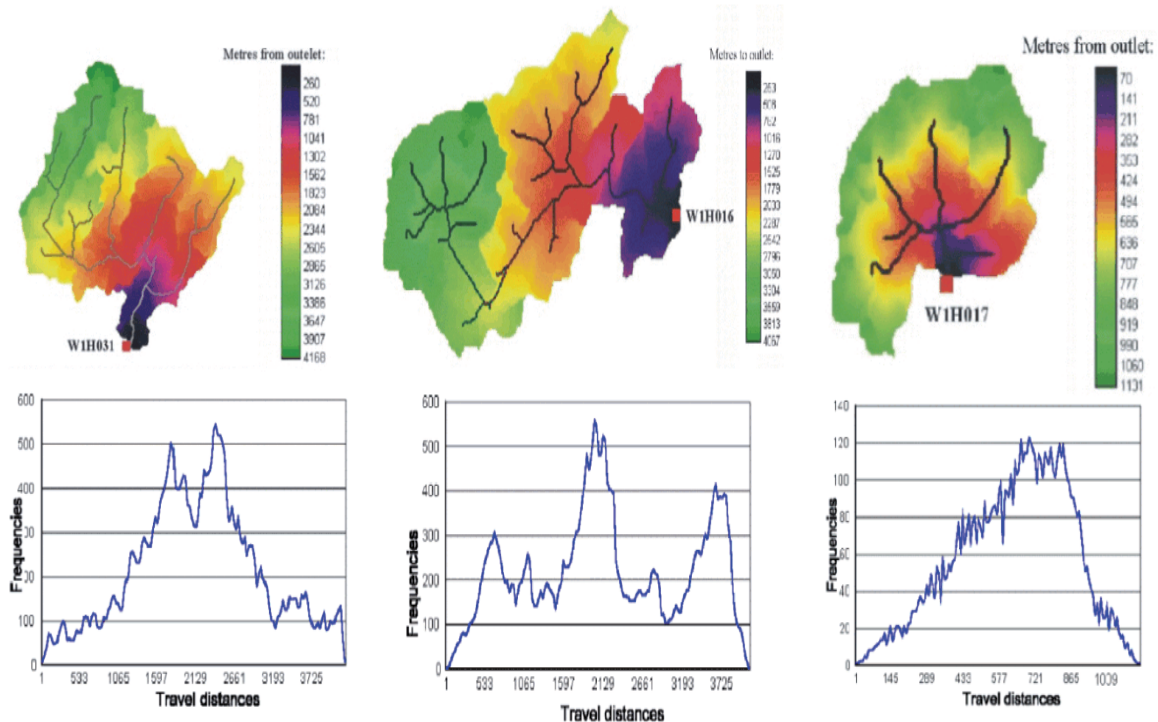


Figure 1.5: The geomorphic response function derived for the three catchments

Geomorphic Response Function

The two-dimensional flow plane has a finite number of flow paths that represent the passage of a unit incident of rainfall routed to the catchment outlet. These pathways are defined by the path of steepest slope which can be derived in GIS from a digital elevation model (DEM). The frequency of path lengths is unique for each flow plane.

Figure 1.5 is a schematic diagram showing the geomorphic response function represented by the frequency distribution of path lengths. Kelbe and Snyman (1993) showed that these geomorphic response functions were similar in shape to the storm hydrograph. The storm hydrograph from W1H031 (left graph in Fig. 1.5) represents three streams converging at the same time while the storm hydrograph for W1M016 (centre graph in Fig. 1.5) has three distinct peaks representing the progressive discharge from different sub-catchments.

Travel Time Response Function

The flow rate along each flow path is dependent on the slope and hydraulic resistance. For surface processes (quick-flow) the slope is represented by the topographical features, the resistance is defined by Manning's n and the hydraulic depth of flow is determined by the upstream contributing area. For subsurface flow paths, the slope is defined by the water table and the resistance to flow is the permeability of the porous media. It is necessary to incorporate a scaling factor for both surface and subsurface flow rates as the contributing area and the resistance to flow are scale dependent. The travel time response function for all three catchments is given in Fig. 1.6. The three response functions have been derived entirely from the physical features of the catchment (10 m resolution DEM, land use and

soils) using various GIS functions. The travel time response functions shown in Fig. 1.6 are a representation of the catchment unit hydrograph for the specific flow plane following a unit of rainfall that follows the specified flow process through the surface zone, unsaturated zone or through the groundwater zone. These flow component unit hydrographs are the building blocks for the construction of the storm hydrograph once the proportion of rainfall/runoff that contribute to the specific flow plane are estimated and partitioned.

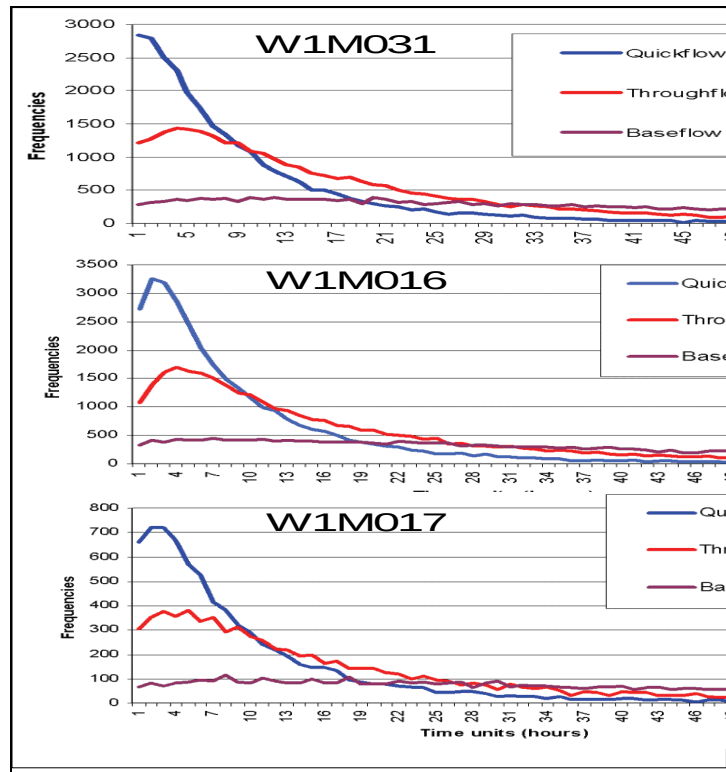


Figure 1.6: The travel time response function (unit hydrographs) for each flow component for three catchments

Concepts and Methodology

The concepts and methodology for combining the geomorphic unit hydrograph into the storm hydrograph are summarized in Fig. 1.7 and described in detail by Snyman (2006).

Simulated Runoff Components

The unit hydrograph model was applied to selected storm events to compare the actual and convoluted storm hydrographs. For the low-intensity storm, there is no quick-flow and all rainfall is partitioned between through-flow and base-flow. The distribution and partitioning of rainfall intensity and duration of storms in W1M016 is shown in Fig. 1.8.

A comparison between the observed storm hydrograph and the model prediction for all four storm categories is shown in Fig. 1.9 for W1H016 and for W1H017 in Fig. 1.10. There is good agreement between the actual storm hydrograph and the simulated storm hydrograph for the cases with long duration storms. However, the agreement is not as good for the cases for short-duration storms.

The correspondence between the actual and convoluted hydrographs will depend heavily on the partitioning of the rainfall into the various flow regimes. Further studies are required to establish the partitioning of rainfall for storm types.

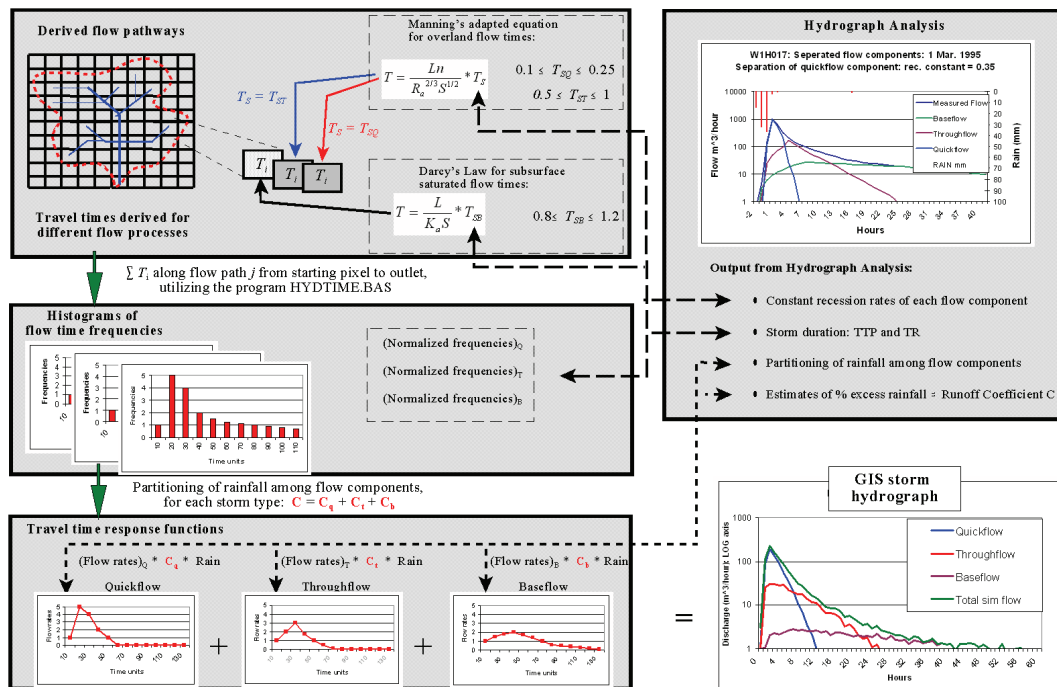


Figure 1.7: The development of the storm hydrograph using the geomorphic response function concept

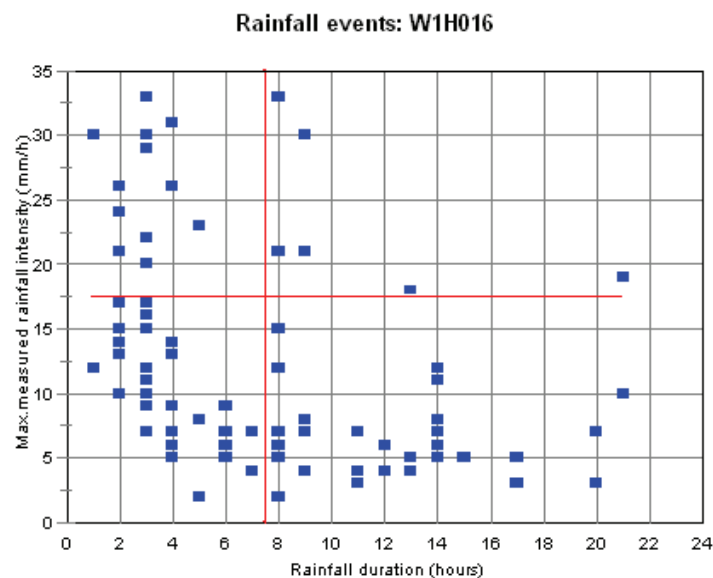


Figure 1.8: The distribution and partitioning of rainfall intensity and duration of storms in W1M016

Model Application for the Ntuzi Catchments

The model was applied to all the rainfall for the Ntuzi catchments for the period from November 1992 to March 1993. The simulated and actual discharges for W1MO 17 are shown in Fig. 1.11. The level of agreement varied between storms. However, the simulated recession rate for most storms was generally comparable to the observed rates.

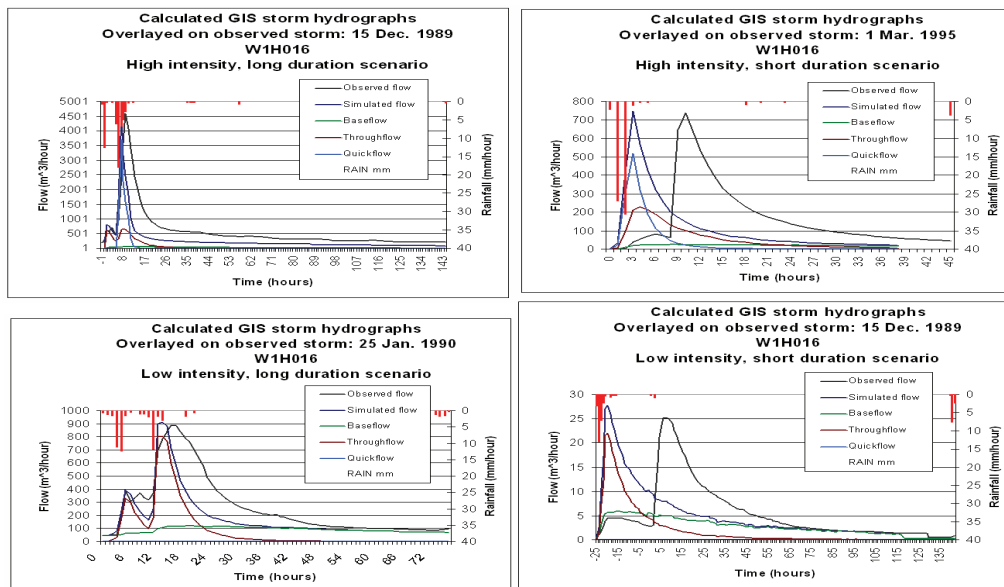


Figure 1.9: Examples of the comparison between observed and calculated storm hydrographs for the four storm categories in W1H016.

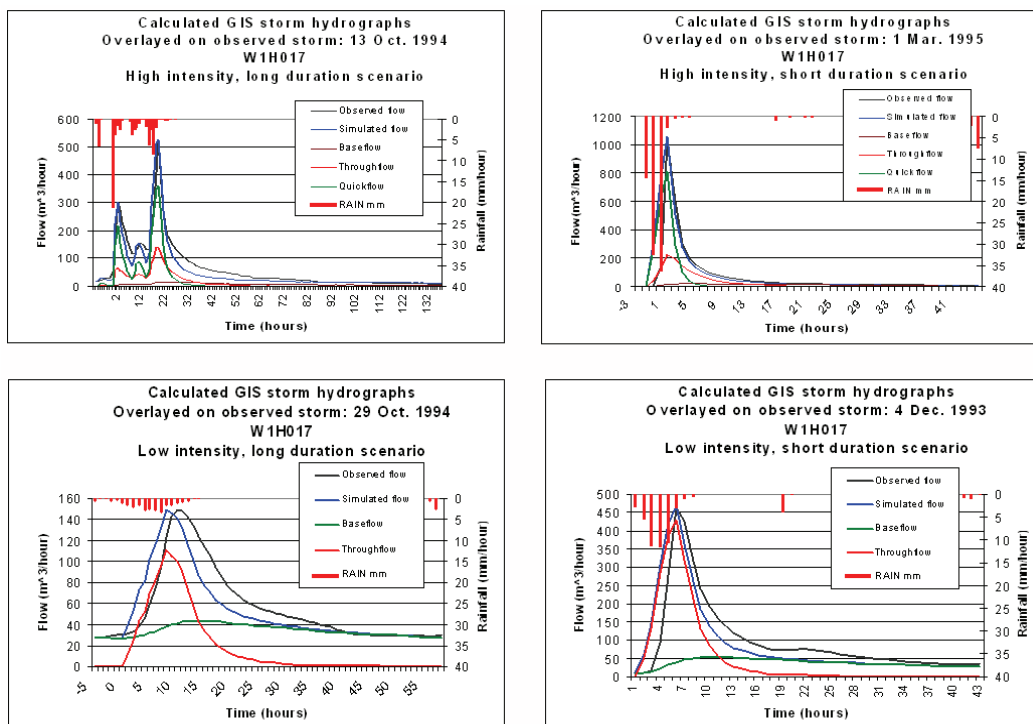


Figure 1.10: Examples of the comparison between observed and calculated storm hydrographs for the four storm categories in W1H017.

Summary and Conclusions

A model depicting different preferential pathways leading to the runoff in small headwater catchments has been presented here for three monitored catchments in the Ngoye hills section of Maputaland. The model is based entirely on the physical features of the catchments and the partitioning of rainfall into preferential pathways. The model predicts the time series of the flow components in small streams and rivers. It has been applied to larger catchments (Mhlathuze) where there is considerable mixing that of the flow components in the river discharge. This physically based model is offered as an alternative to the stochastic/empirical models currently available.

This case study demonstrates the application of the model to small- and medium-size catchments and shows potential for application to large catchments where it is necessary to quantify the level of interaction between the surface and groundwater systems.

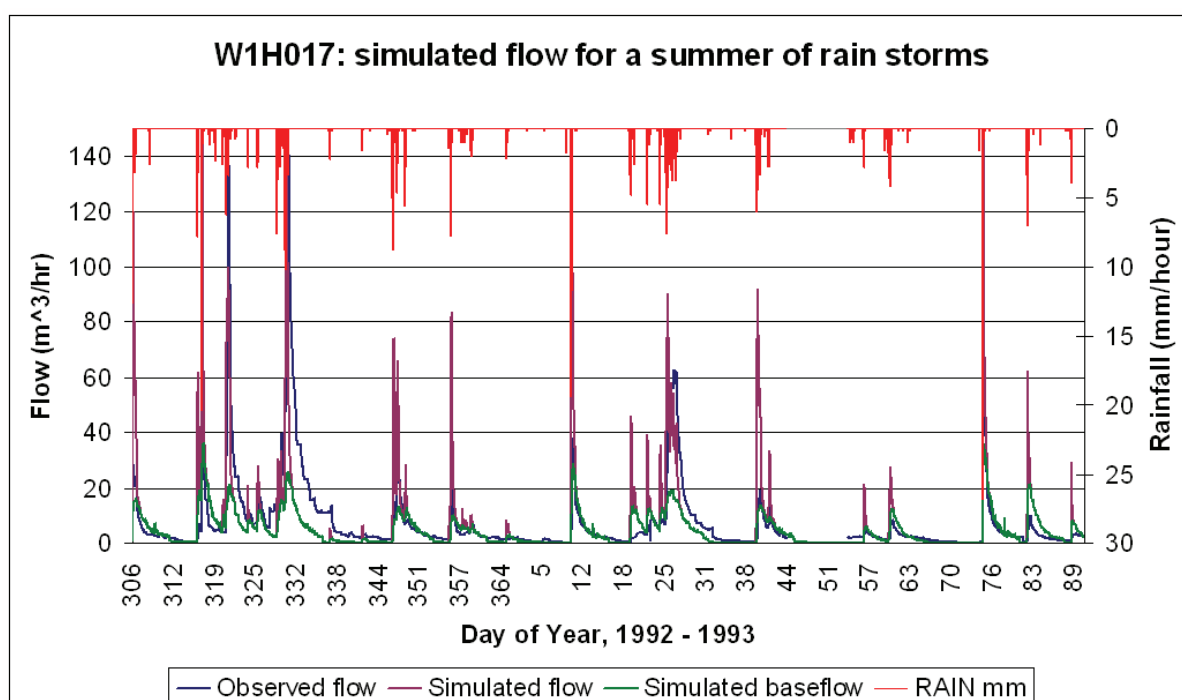


Figure 1.11: A comparison between simulated runoff and observed runoff for W1H017 for the 1992/1993 season

CASE STUDY 2: INVESTIGATION OF THE APPLICATION OF A GROUNDWATER MODEL FOR STREAM DISCHARGE ESTIMATION IN REGIONS DOMINATED BY SHALLOW PRIMARY OR ALLUVIAL AQUIFERS

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INTRODUCTION

The determination of groundwater contribution to stream discharge is complicated by several processes that occur along the hydrological pathways in different environmental settings. Methods for determining the interaction between aquifers and river systems and their application under certain circumstances need to be evaluated at different scales of interaction. This case study looks at the application of a well recognized numerical model (MODFLOW) to estimate the contribution of groundwater to the discharge in a small stream segment in a shallow groundwater environment that forms part of a local wetland along the Maputaland Coastal Plain.



Plate 2.1 Photograph of the stream that is typical of many Maputaland coastal streams

Location and Setting

The MODFLOW model with linked stream and river functions has been applied and calibrated to several stream segments in the Richards Bay area to evaluate their application in studies of the surface water/groundwater interactions.

The first stream section under investigation drains a small section of the Zululand Coastal Plain in the heart of Richards Bay (Fig. 2.1). The catchment area (approximately 1.5 km²) is very flat with a relatively large area of wetlands in the central section and it drains a large industrial development site. The stream channel has eroded the landscape significantly and lies several metres below the surrounding

banks. The cross-sectional profile near the railway bridge is significantly wider and deeper than the central section.

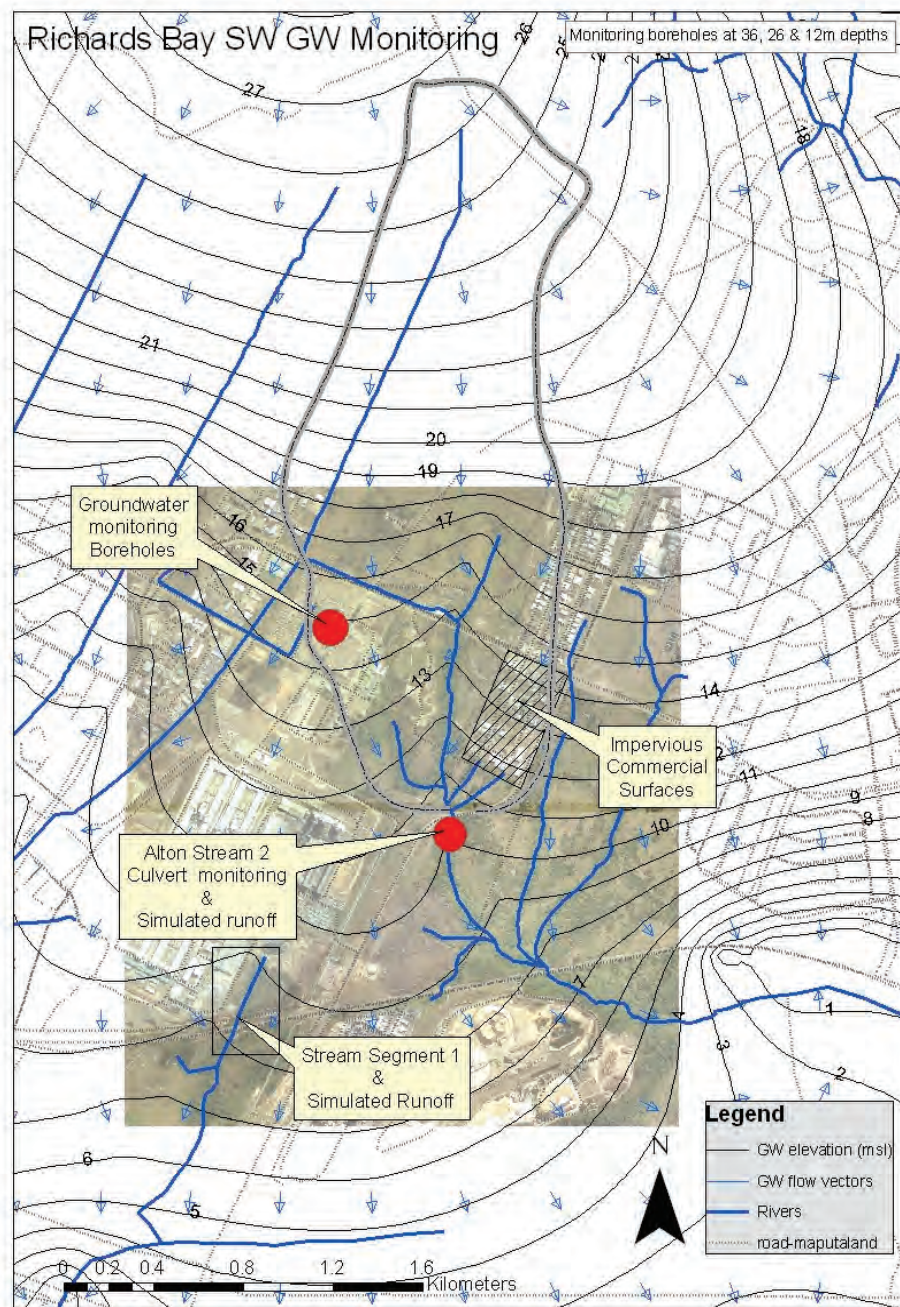


Figure 2.1: The location of the Alton streams and monitoring points for model evaluation

A typical cross-sectional profile is shown in Plate 2.1. The wetland to the north of the site (Plate 2.2) had no. However, the water table was very close to the surface and was an important component of the wetland. The Richards Bay Municipal plans indicate that the wetland is zoned as a protected area in their Municipal Open Space System (MOSS) plan.

The second stream evaluated drains the main industrial section of Alton (Fig.2. 1) and was monitored on a continuous basis at the road culvert for more than 12 months.

The third set of streams drain through road culverts into Lake Nhlabane (Fig. 2.2). These were monitored on an ad hoc basis over several months to capture high flows and base flows.

These examples formed part of an attempt to evaluate the use of groundwater models for simulating the flow in a streamflow segment and determine the role of groundwater and the wetlands in assessing the river flow and its contribution to the Richards Bay harbour.

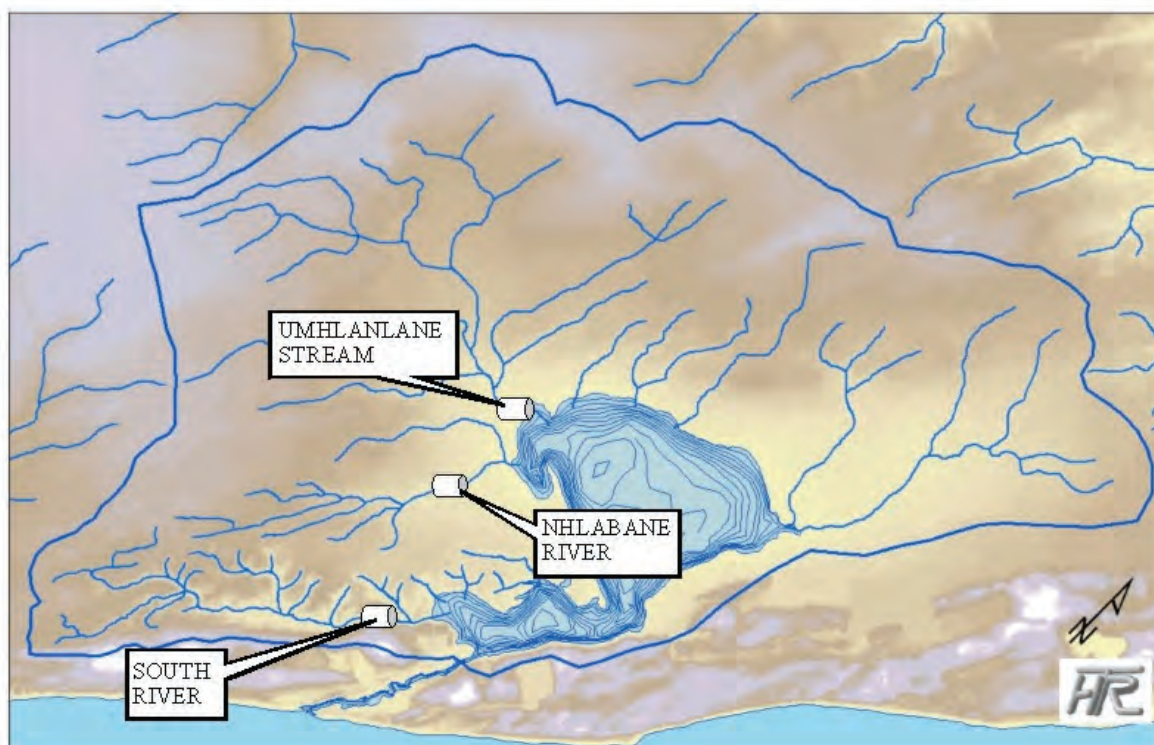


Figure 2.2: The location of the Nhlabane River monitoring points for model evaluation

The Models

The numerical model is configured to represent the conceptual geological model described in the previous sections of this report. The numerical model is based on MODFLOW 1998 and MODFLOW2000 code (Harbaugh and McDonald, 1996a;b) with the appropriate functions to simulate the flow dynamics in three dimensions over a time span varying from days to over 33 years.

The streams were configured as head-dependent boundaries using either the River Package (RIV) or the Stream Package (STR1) (described by Prudic *et al.*) linked to the MODFLOW code. In the River Package the drainage boundary functions as the discharge segment where water is lost or gained by the aquifer. In the Stream Package the flow volume and rate within the segment is also determined. In the stream package, the Manning equation was used to determine the stage of the stream. Values of Manning coefficient for the streams were derived from Arcement and Schneider (1984). All the streams in the lake catchment areas were digitized in ArcGIS to capture the stream junctions (nodes) where the stream bed elevation was derived from the contour elevation of the DEM (Fig.2.3). The stream width and stream bed thickness values were estimated for these nodes. The slope of the stream segments (sections

between nodes at the confluence of all streams) was derived from the node elevations. The Manning Roughness Coefficient was estimated from site visits to some of the stream segments using the values published by Arcement and Schneider (1984). The hydraulic characteristics of the stream bed sediment are unknown but were all assumed to have hydraulic conductivity values similar to the calibrated soil profile (upper layer of the model) values. It was initially assumed that most of these drainage features are perennial streams with little deposition to form clay or peat beds with lower hydraulic properties than the underlying aquifer. This is an assumption that needs to be evaluated because it is known that the stream draining from Hillside had a significant peat/clay bed that was seriously undercut by erosion emanating from the railway culvert which needed to be controlled before undercutting the John Ross roadway. The values of the hydraulic conductivity of the stream beds were adjusted during the calibration process to achieve the indicated levels of agreement with the few available stream flow measurements.

The evaporation mode in MODFLOW was invoked as a head dependent boundary where the evapotranspiration rate was defined as a linear function of the depth to the water table, depth of the rooting zone and a groundwater potential evaporation rate. The unsaturated zone for this wetland area implies that the groundwater potential evaporation rate must be very close to the potential evaporation rate.

Alton Stream Segment

This study area drains a small portion of the industrial area of Richards Bay (Fig. 2.1). The water table is close to the topographical surface in this area but the vegetation suggests that there may be inundation in the wetlands during large storm events and there may be a limited amount of surface flow away from the stream on the western section of the site.

Stream discharge was measured at a point just upstream of the railway bridge and at a point about 500 m upstream of the downstream monitoring point (see Fig. 3). The difference in flow between these two measuring stations is expected to indicate the groundwater contribution to the stream section bordering the study site since there are no surface streams draining the study site. The flow measurements are given in Table 2.1.

Table 2.1: Stream discharge measurements (m³/d)

Date	Upstream	Downstream
08 Oct 1999		2 154
14 Oct 1999	297	555
22 Oct 1999	5 513	4 301
26 Oct 1999	2 873	4 287
27 Oct 1999	3 070	6 061

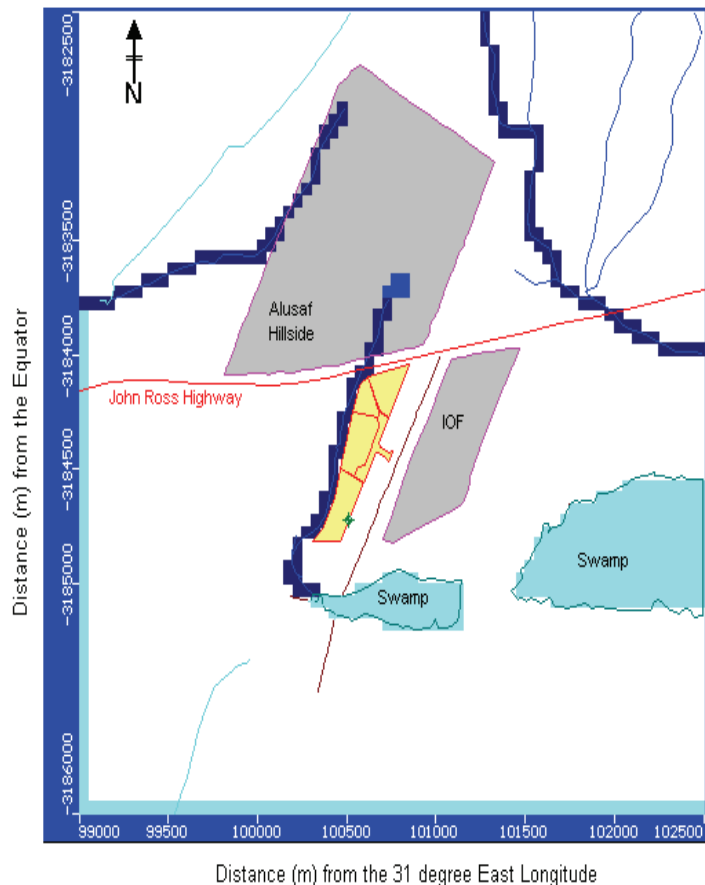


Figure 2.3: The boundary conditions for the numerical simulations. The lines of blue rectangles are the free draining streams. The light blue sections are constant-head boundaries

In order to achieve sufficient accuracy in the flow measurements to determine the difference in discharge rates at the two sites a fixed cross section was inserted into the stream bed and flow rates were measured. With the exception of one day, all measurements indicate that the flow downstream was significantly greater than the flow upstream.

Rainfall

The mean annual rainfall field for the Richards Bay area was used to create a regional groundwater model where the water table elevation was calibrated against several borehole measurements in the model domain. These have been described by Kelbe and Germishuyse (2000c). During this study period, the rainfall was taken from a rain-gauge site near the airport in Richards Bay (~5 km away).

Evaporation

Regional evaporation estimates have been obtained from WR90 (Midgley *et al.*, 1994) and the published maps of the region by Schulze *et al.* (1997). Both these studies indicate the potential evaporation. The actual evaporation is limited by the potential rate and the plant physiology. The numerical model requires some partitioning of the actual evaporation between the surface and subsurface processes. No measurements are made of these components and consequently, the evaporation is based on a conceptual model described by Germishuyse (1999) and Kelbe and Germishuyse (199b). The potential evapotranspiration for the region has been estimated as 1 400 mm/yr from WR90, Schulze *et al.* (1997) and measurements at the University of Zululand (Kelbe, 1990).

Numerical Boundary Conditions

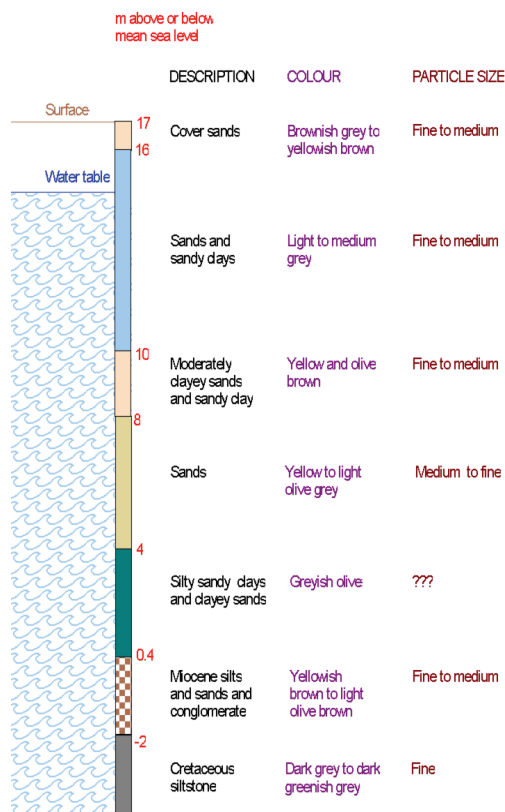


Figure 2.4: Borehole log for the study area

All the available borehole logs were used to confirm that these surface boundaries coincided with the observed data at several points in the model domain. The available log for a borehole in the eastern section of the study site is given in Fig. 2.4 (NB. The location of the borehole is on the northern end of the Alusaf Hillside site shown in Fig. 2.3). The aquifer depth is generally less than 30 m in most areas of the model domain. This is considerably less than the grid spacings that are greater than 100 m. The sediments comprising the components of the unconfined aquifer are all sandy deposits with varying clay contents. Consequently, a single layer model was considered adequate for this study of the flow dynamics.

Model Calibration and Simulations

The numerical model was constructed in Visual MODFLOW using the MODFLOW 98 kernel. MODFLOW was used to simulate the discharge along the river segments downstream of the Alusaf effluent pond. The water balance for the specified stream segment was construed as the groundwater contribution to streamflow for each stress period. The measured river discharge at the upper and lower ends of the stream segments were used together with the piezometric surface of the aquifer as the primary calibration factors. The simulated flow for the period of rainfall observation and streamflow measurements are shown in Fig. 2.5. Also shown are the measured flow rates that were taken to coincide with rainfall events. The simulated flow for both the upstream section and the discharge near the railway bridge are shown. The difference between the two series indicates the contribution to the total flow from the river segment bordering the study site.

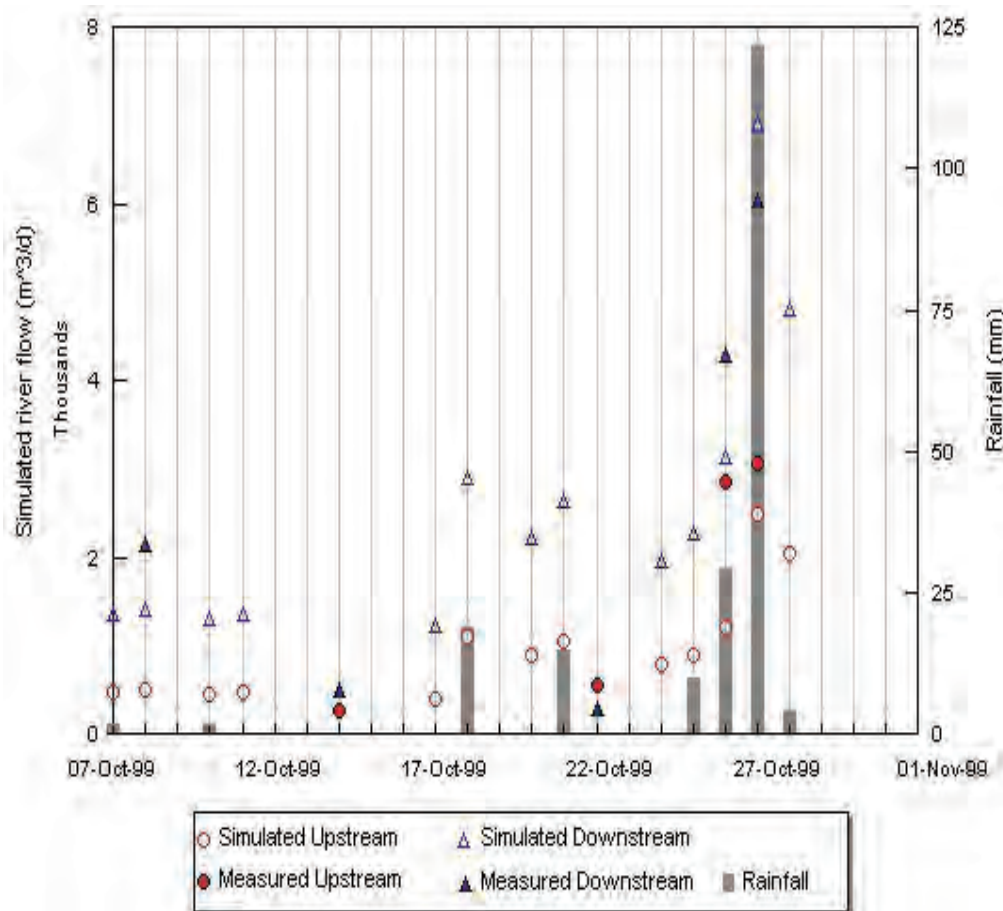
The numerical boundary conditions have been chosen to cover the entire catchment area of the stream. The horizontal extent of the model domain is constrained by the harbour and the surrounding streams shown in Fig. 2.3. The specified internal boundaries are the effluent pond and the stream section flowing into the swampy areas just to the south of the study site. The swamp area has been specified as a constant head boundary.

The vertical boundaries are described by the geology and geomorphology. The topographic surface describes the upper surface of the model domain. This surface was constructed using digitized 2 m contours from orthophotos. The lower boundary of the primary aquifer was chosen to coincide with the upper surface of the Cretaceous siltstone deposits. This surface has been described in detail by Germishuyse (1999) and Kelbe and Germishuyse (2000a).

Table 2.2: Simulated flow (m^3/d) characteristics for 10 years

Characteristics	Upstream	Downstream
Mean	1 014	2 558
Max	3 492	6 906
Min	323	1 074

The simulated flow rates vary in accordance with the observations although the correspondence is not precise. The difference in flow between the upstream and downstream monitoring points is reflected in the simulations. Consequently, the model is considered to be adequate for an initial (rapid) assessment of groundwater contribution to streamflow in these conditions at this scale of application. The general flow characteristics for the stream discharge at the two points spanning the study site are shown in Table 2.2. The minimum (base) flow was $323 \text{ m}^3/\text{d}$ which occurred on 23 August following a prolonged period with little rainfall. The high flows on 27 October 1999 indicate a substantial difference between the upstream and downstream points which may be partially attributed to some surface flow. A detailed analysis of the groundwater and river hydrographs is given in greater detail in Case Study 3.

**Figure 2.5: The measured and simulated streamflow rates (m^3/d)**

The Alton Catchment – Stream 2

An assessment of the application of the groundwater model to estimate the base flow component of a small catchment in Alton is presented in this example. The streamflow was measured on an hourly basis (see Fig. 2.6).

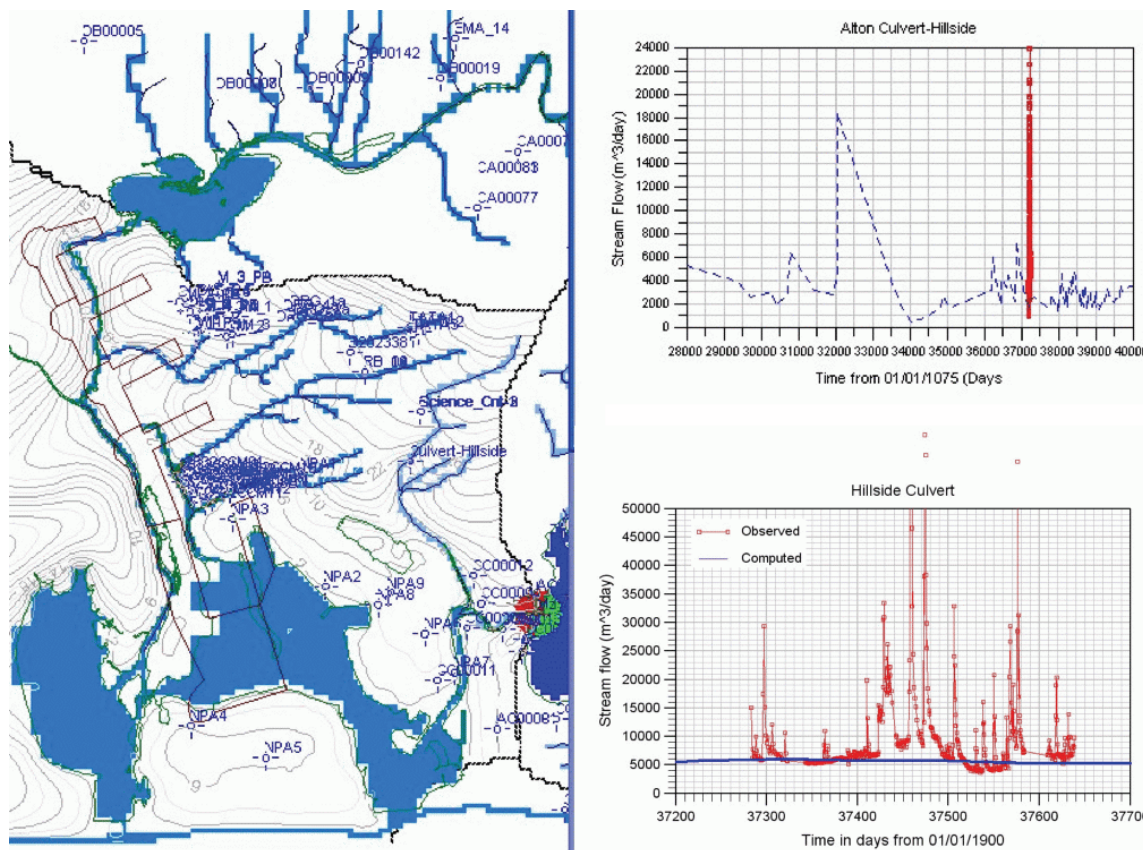


Figure 2.6: The simulated baseflow contribution to the runoff from the small catchment in Alton, Richards Bay using MODFLOW 2000 and the STR1 package.

Note the time frame of the model in the upper graph and the detailed period of the observation in the lower graph.

The Nhlabane Streamflow

An independent assessment of the application of the groundwater model for simulating the total streamflow was done in the Nhlabane catchment. The model was developed by Kelbe (2008) as part of a yield analysis of Lake Nhlabane and the total stream flow rates were required for the mass balance of the lake and used extensively for the model calibration. This model was calibrated against available piezometer and borehole readings and the stream flow in the drainage boundaries modelled using the Stream Package (STR1). The location of the catchment and monitoring points are shown in Fig. 2.7. The model has been described in more detail by Kelbe (2008).

The ad hoc measurements of the streamflow were intended to cover the range of hydrological conditions that represented baseflow and storm flow to support the model calibration. The monitoring points and streamflow measurements are given in Fig. 2.7. The model applied the STR1 package that

calculated the baseflow as a head dependent boundary and incorporated the storm flow as a function of the rainfall in the catchment. The storm flow component was derived from an analysis of the stormflow hydrograph discussed in the main report.

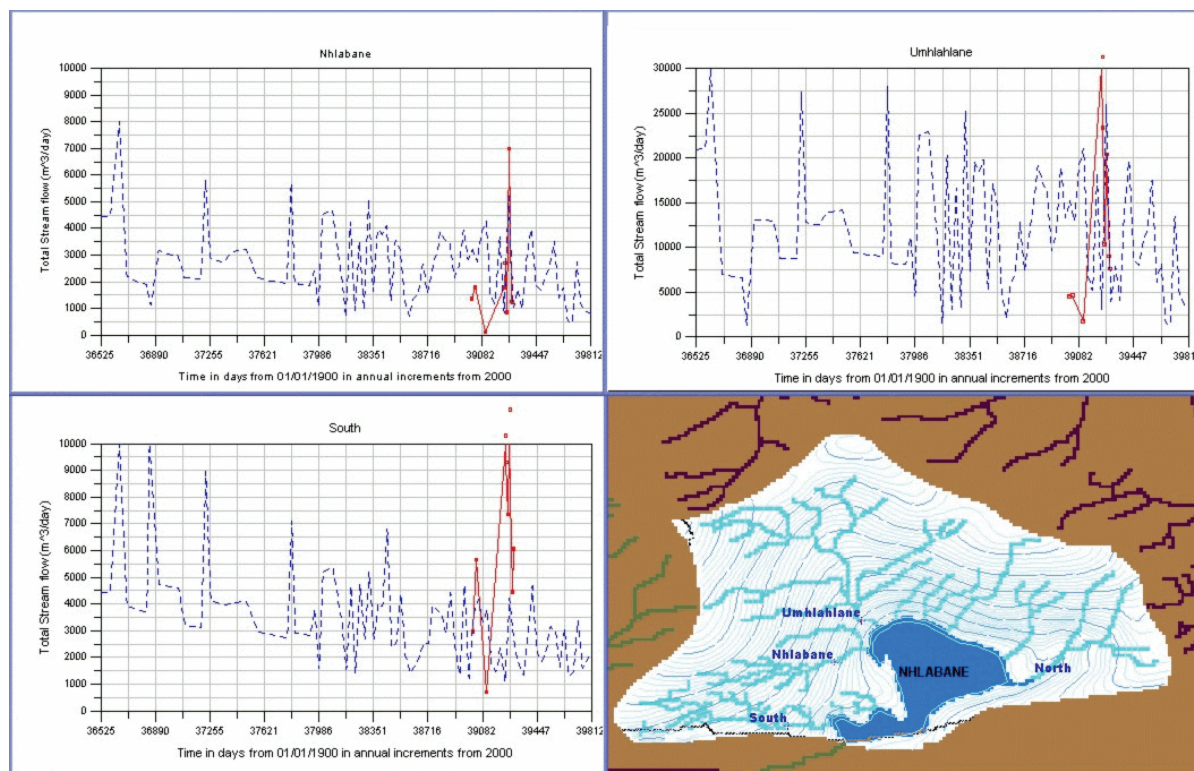
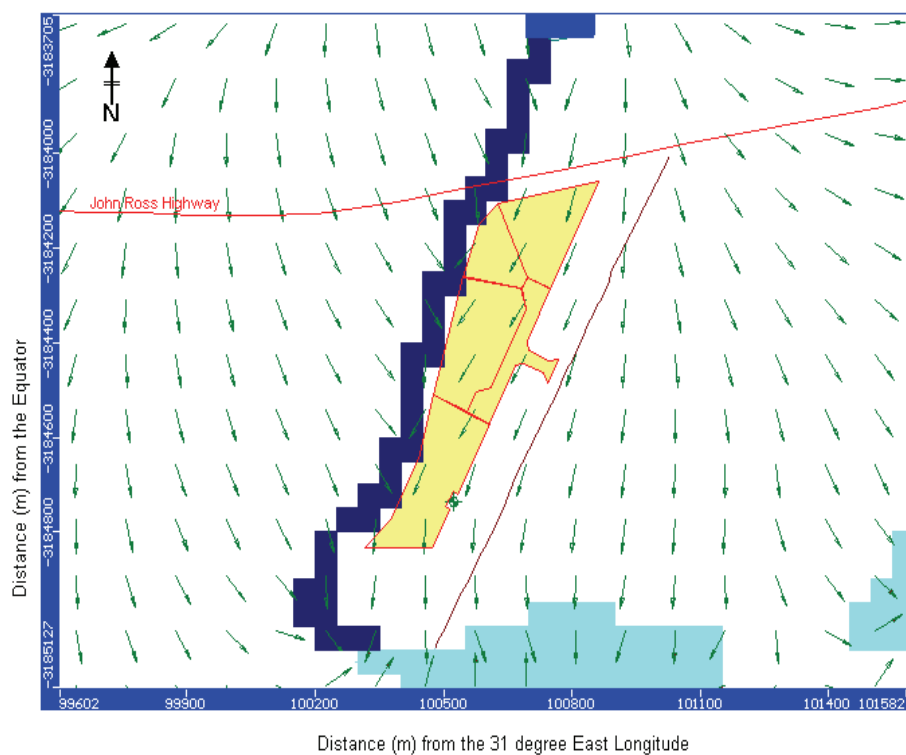
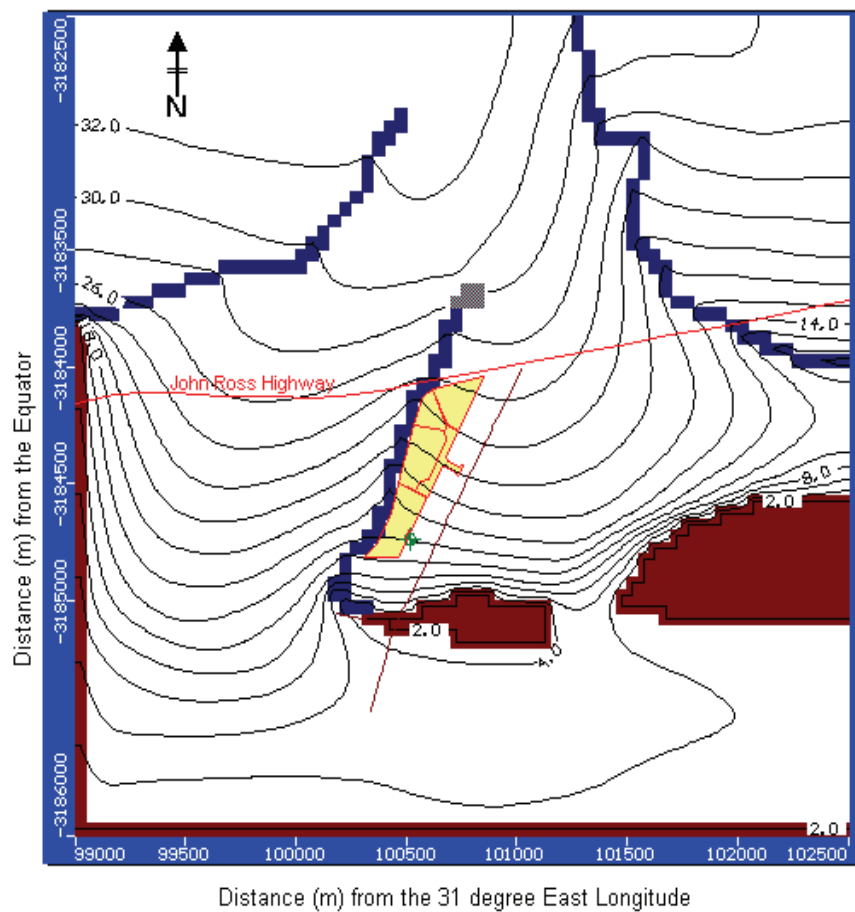


Figure 2.7: The simulated and measured total streamflow at three locations in the Nhlabane catchment

Summary and Conclusions

The need for methodologies to determine the groundwater contribution to surface water resources is well recognized. Suitable methodologies are needed for the groundwater reserve determinations at various scales in areas where the surface water dependence is important. There is also a need to be able to estimate the groundwater dependencies of different vegetation types and under various conditions, particularly where the vegetation is likely to be in hydraulic contact with the water table (see Case Study 6 for Lake St. Lucia).

This case study examined the role and applicability of using a well-defined groundwater model to generate estimates of groundwater discharge in very small un-gauged streams where there is a well recognised groundwater dependency. However, the level of confidence is dependent on the information that may be available for the estimation.



CASE STUDY 3: LAND-USE IMPACT ON THE RIVER FLOW AND ESTUARY OF THE SIYAYA SYSTEM

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INTRODUCTION

The Siyaya Estuary is situated near Mtunzini on the Zululand North Coast (Fig. 3.1). It forms an integral part of the Umlalazi Nature Reserve (renamed the Siyaya Coastal Park) and has had a pivotal role in the local community. It is used extensively by the local community as a recreation facility and many years ago it was the venue for such events as the local school gala. However, its physical features have deteriorated and its recreational functions are diminishing.

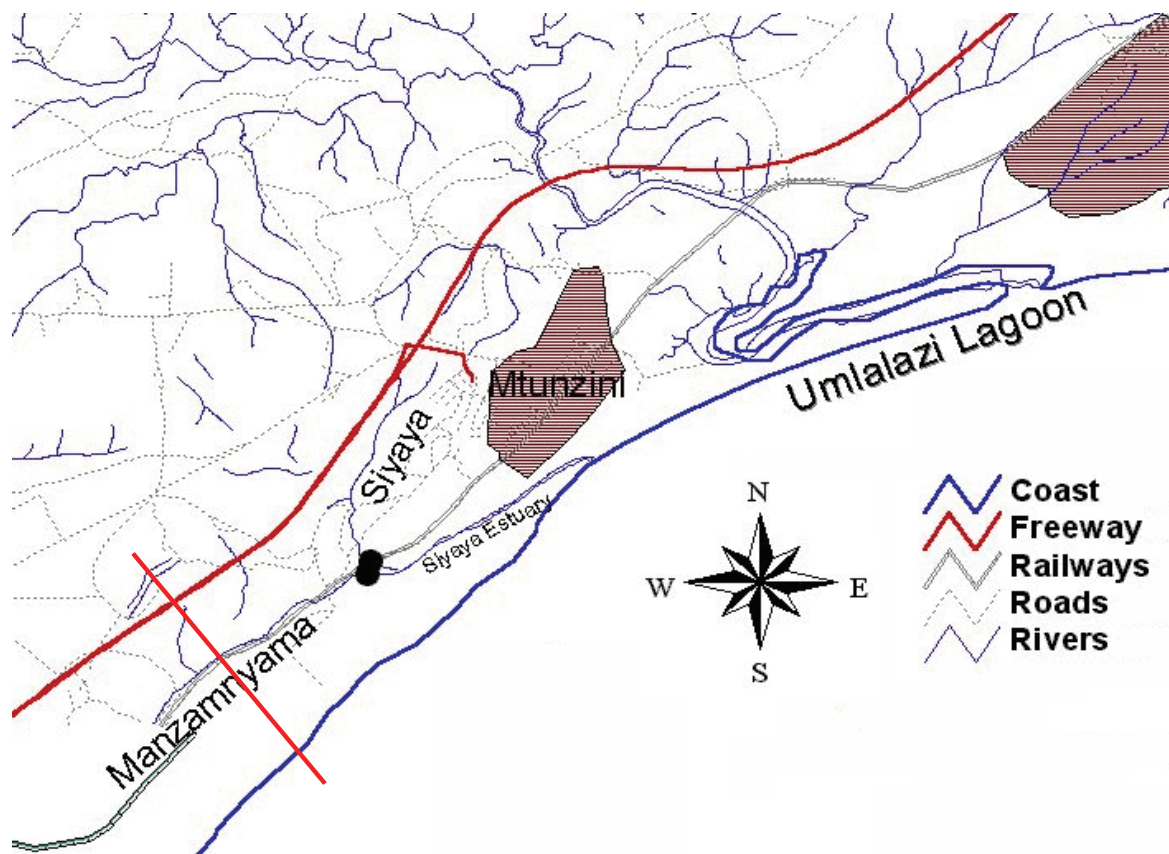


Figure 3.1: The location of the Siyaya Estuary. The red line denotes the transect in Fig. 3.3.

Beggs (1984) wrote of the Siyaya lagoon: 'Today a rusty old fridge lies dumped on the water's edge and chocolate coloured water streams toward the sea, building up slowly against the closed mouth of the lagoon or, during the rainy season, bursting out into the sea in a muddy mass before being transported northward by the current. Silt together with fertilizers, herbicides and bolides leached from the land upstream, pollute the water'. Beggs (1984) goes on to say that 'the conditions in the lagoon are symptomatic of what problems lie above it.' The identification of these problems prompted an in-depth investigation of the Siyaya under the coordination of the Natal Town and Regional Planning Commission.

The conditions in the estuary and the catchment have changed since Beggs (1984) wrote his report over 20 years ago. The lagoon is almost completely covered in reeds (Fig. 3.2) and the Manzanmyama tributary to the Siyaya has stopped flowing except during large storm events.

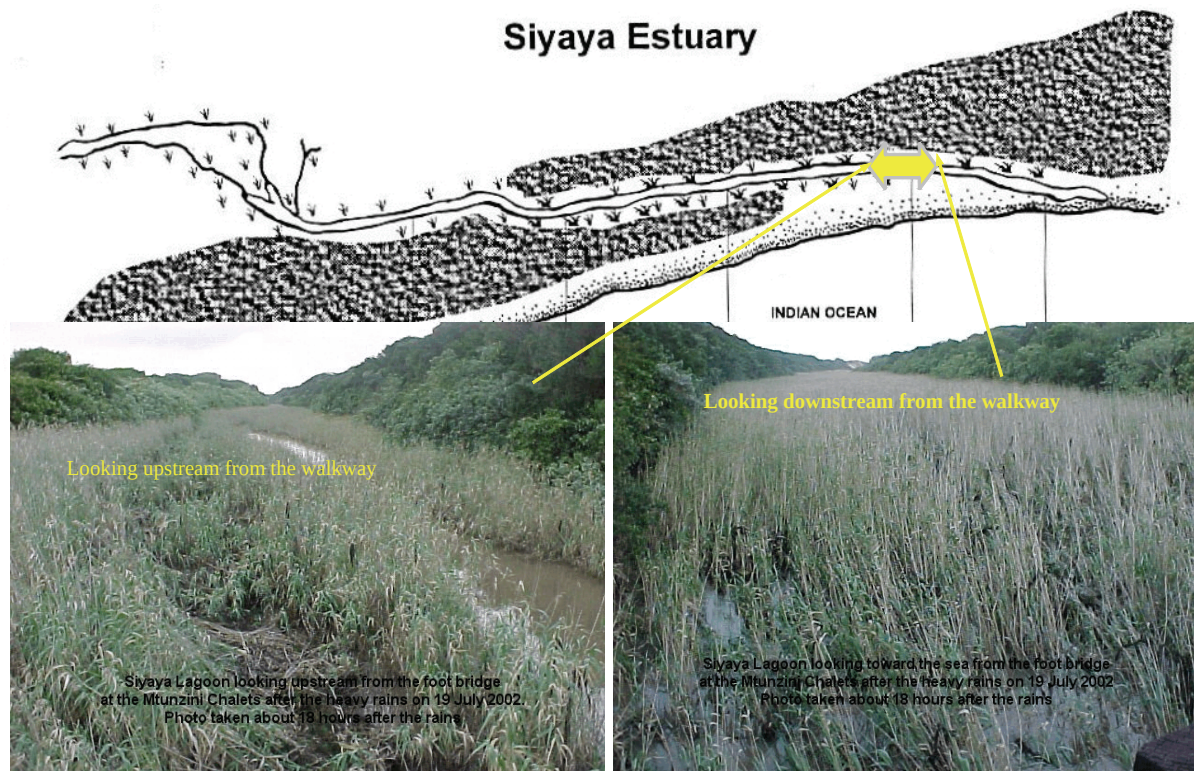


Figure 3.2: Present status of the Siyaya lagoon. Photographs taken from foot-bridge.

This case study is presented as an example of the degradation of estuarine and riverine system due to the reduction in baseflow. It shows the importance and present status of the Siyaya lagoon and its functional components. The status of the lagoon is shown to depend on the role and problems of the catchments that contribute to the conditions in the estuary.

Physical Features

The Siyaya Estuary, like most of the estuaries along the east coast, is driven northwards by the local near-shore currents. It is fed by the Siyaya and Manzanmyama Rivers (Fig. 3.1) that drains the coastal strip composed of old dune relics (Fig. 3.3) that are comprised of potentially erodible sandy soils.

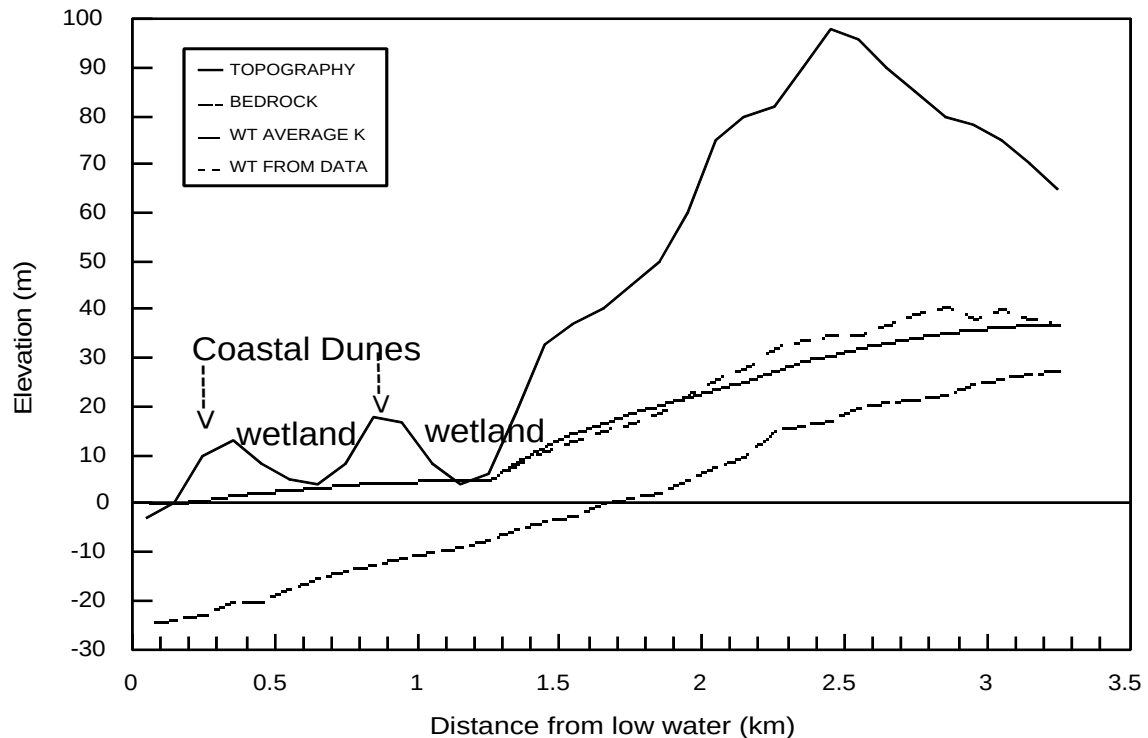


Figure 3.3: Vertical profile through the catchments of the Siyaya system

The Manzanmyama catchment drains the steep dunes towards the sea and into the coastal wetlands. It is prevented from flowing directly into the sea by a series of low coastal dunes. Instead, it flows northwards parallel to the coast to the confluence with the Siyaya. In contrast, the Siyaya Catchment drains south to the confluence with the Manzanmyama (Fig. 3.1). Part of the Siyaya Catchment is occupied by the town of Mtunzini.

The Manzanmyana Catchment lies between the N2 freeway and the Umlalazi Nature Reserve. The hydrological conditions of the Estuary have been created by conditions upstream in the two feeder catchments. The conditions in the catchment have been monitored for specific periods to determine the runoff and sediment characteristics in integrated studies of land use effects (Van der Elst *et al.*, 1999). The runoff from both catchments was monitored by two weirs near their confluence (Fig. 3.4). The weirs repeatedly filled with sediment and were eventually abandoned in the late 1990s. However, rainfall and runoff measurements for a period of intensive observation were conducted during much of the 1980s (Van der Elst *et al.*, 1999).

Land use and hydrological conditions in the early 1980s were well documented by Van der Elst *et al.* (1999). Both catchments were almost entirely covered by sugar cane (Fig. 3.4) which led to severe sedimentation that eventually impeded the flow in the Siyaya Estuary. There was one small stand of forest along the coastal strip. The sugar cane production was particularly degrading along the riverine section where nearly all streambeds were incorporated into the cultivation. The complete lack of indigenous riverine vegetation is clearly visible in the aerial photographs in 1975 (Fig. 3.5)



Figure 3.4: Weirs WIH018 and W1H019 on the Manzanymya and Siyaya Rivers



Figure 3.5: Aerial photograph of the catchments under sugar cane production in 1975

There was a radical change in land use in the 1990s when all but one of the sugar-cane farms was converted to *Eucalyptus* plantations in the Amanzanyana as well as many of those in the Siyaya (Fig. 3.6). Under afforestation, there was the complete re-establishment of riparian strips under forestry that was completely lacking under sugar plantations.

Previous Studies

Through the conservation efforts of Mr Ian Garland and the support of many organizations (SASA, Cane Growers Association, Department of Agricultural Technical Services, Department of Water Affairs, CSIR, University of Natal and the Oceanographic Research Institute) the Siyaya Catchment Demonstration Project (Van der Elst *et al.*, 1999) documented the physical conditions of the estuary and contributing

catchment during the 1980s. This project provided several years of hydrological measurements that describe the characteristics of both catchments and highlighted the differences in their response.

In the 1990s, all but one of the cane farms in the Twinstreams Catchments were transformed from sugar cane to *Eucalyptus* plantations. The Siyaya Catchment Demonstration Project (SCDP) was terminated and the runoff measurements stopped. The Hydrological Research Unit endeavoured to re-establish the weirs but the serious problem of siltation in the weir dam made the measurements too unreliable and only ad hoc measurements were conducted. At the turn of the century, DWAF recommenced the weir measurements.

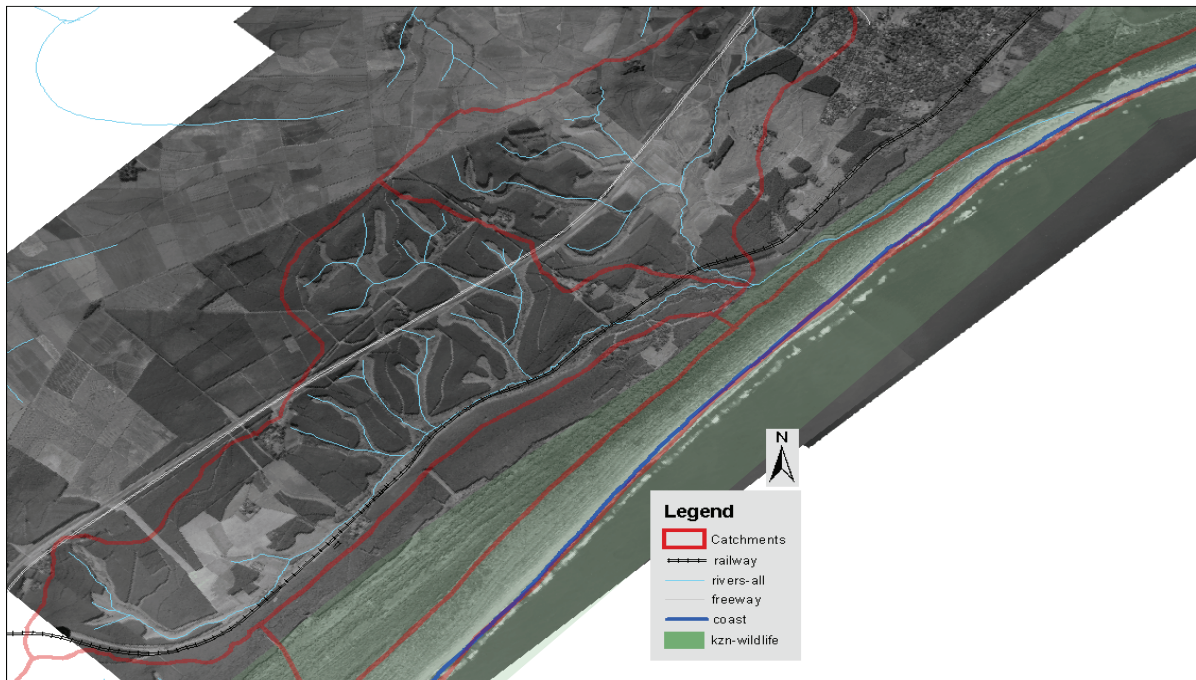


Figure 3.6: Aerial photograph of the catchments under *Eucalyptus* plantations in 2000

During the 1980s, the Siyaya River discharge was less than the flow in the larger catchment of the Manzanymama although some individual storm events showed higher runoff in the Siyaya Catchment. However, the low flow (baseflow) was consistently higher in the Manzanymama River reflecting the relatively larger size of the catchment. More importantly, neither river ever stopped flowing during this period of observation.

Table 3.1 Spot measurements of flow rates (m³/d) in January 1996

Weir/Date	21/1/96	27/1/96
Manzanymama	604	14 000
Siyaya	2 074	15 400

Some ad hoc measurements were taken following the previous study. The flow rates on 21/1/96 and 27/1/96, at the Manzanymama and Siyaya weirs are given in Table 3.1. The flow measurement on 27/1/96 was taken about 12 to 14 h after a thunderstorm and accompanying front which produced approximately 70 mm rainfall in both catchments. There was no evidence at the weir of the flow

exceeding this stage so it is assumed that this measurement is close to the peak discharge for the rainfall event on the 27 January. The flow on 21/1/96 represents baseflow conditions while the flow on 27/1/96 is a flood event.

Since the 1990s the baseflow in the Manzanmyama has ceased but the runoff from the Siyaya has seldom stopped indicating a significant shift in the hydrological response of these two catchments between these two periods. Figure 3.7 shows the cumulative discharge for both streams for the two periods of measurements, and the corresponding rainfall taken at Mtunzini (in the catchment). Figure 3.8 shows the probability of the discharge volumes for the same periods in 1980 and 2000.

Sediment Yield

According to Van der Elst (1999) the bulk of the sediment in the 1980s (~1 600 kg/ha·yr) is derived from the Manzanmyama while the Siyaya produced only one quarter of the yield (400 kg/ha·yr). The suspended solids (SS) load for the Amanzanmyana and Siyaya Rivers was also generally much higher than for the other rivers along the Zululand coast. The frequency distribution of suspended solids (SS) for the Amatikulu River just south of the study area indicates a mean concentration of about 60 mg/ℓ with an upper limit of 200 mg/ℓ. This upper limit is close to the mean sediment concentration for the study catchments during the 1980s. Despite the huge efforts by Dr Garland and the Foresters at establishing the riparian vegetation, the almost complete cessation of baseflow and with only one or two storm surges a year in the Manzanmyama, there is no flow to transport any sediment that may be produced.

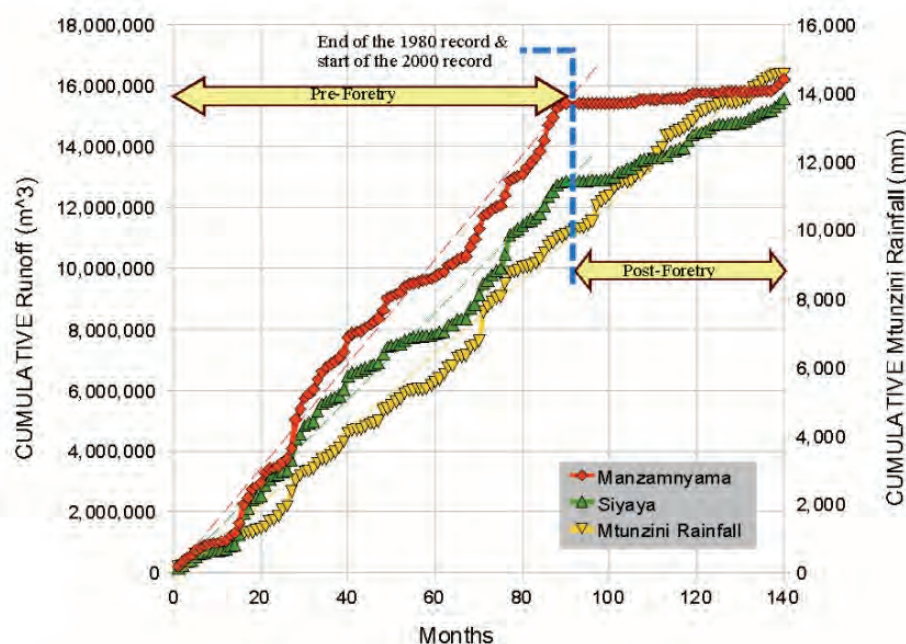


Figure 3.7: The cumulative runoff from the Twinstreams catchments for the 1980s and 2000 periods. No simultaneous runoff measurements were conducted during the 1990's. Also shown is the cumulative rainfall at Mtunzini for the same periods. The blue dashed line represents the 1990s.

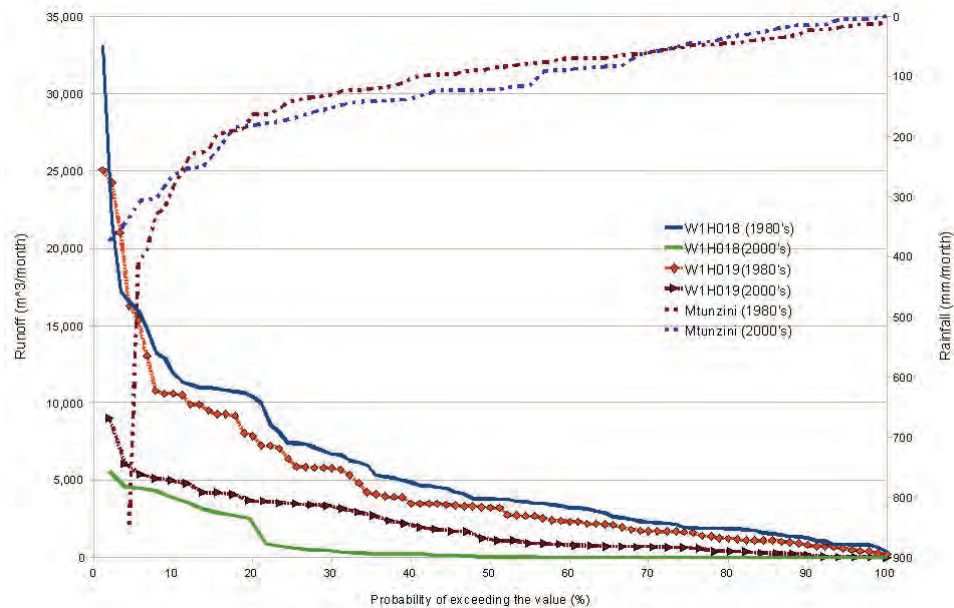


Figure 3.8: The probability of discharge volumes for the Manzanymyama and Siyaya catchments for the 1980 and the 2000 period of measurements. The corresponding rainfall curves are also included.

Summary and Conclusions

The Amanzanymyana and Siyaya Rivers provided the flow into the Siyaya Estuary. Consequently, they played the major role in sediment transport into and out of the estuary and played a pivotal role in the mouth dynamics. There was substantial production of sediment that can be attributed to sugar-cane production up to the early 1990s when the land use changed to forestry. However, the change in land use caused a radical change to the hydrological flow characteristics while significantly reducing the sediment production and transport potential of these rivers. This has changed the flow dynamics of the estuary. Despite the adherence to recognized water conservation practice in establishing the forests, the plantations have severely impacted on the baseflow/groundwater discharge. This has negatively impacted on the restoration potential of the estuary.

The residual sediment and reduced flow in the estuary has provided substrate for extensive reed infestation that have stabilised the river channel (Fig. 2). Even extreme rainfall events that scour the mouth (Fig. 3.9) had no impact on the upstream section of the estuary. The restoration of the estuary and rivers to acceptable EFR and IFR conditions will require intervention in the catchment management practice to restore base flow in the rivers and estuary.



Figure 3.9: Siyaya Estuary Mouth erosion/breach following heavy rainfall events in 2002. The impact on the estuary further upstream was minimal.

The Siyaya Estuary is ranked only 176 out of 248 estuaries in SA (Maree *et al.*, 2003). Consequently, its demise is unlikely to have huge repercussions on the marine environment. However, the effect of extensive sugar cane production and afforestation on the other catchments in Maputaland could seriously affect the most important estuaries in South Africa and subsequently the marine resources. This is an example of what could happen if these other estuaries are impacted in a similar manner.

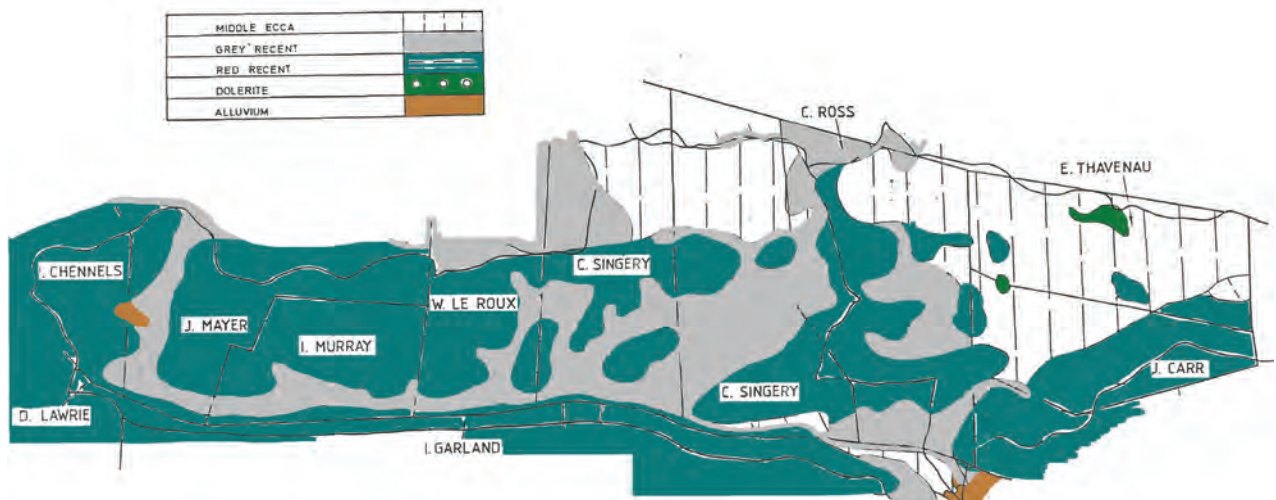


Figure 3.10: Map of the soil pattern for the Siyaya and Manzanymama catchments (adapted from SASA Expt Station, Anon)

CASE STUDY 4: ESTIMATION OF GROUNDWATER CONTRIBUTION TO RIVER FLOWS IN MAPUTALAND USING HYDROGRAPH ANALYSIS TECHNIQUES

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INTRODUCTION

Most of the Maputaland coastal streams are perennial, even the smallest examined in this study continued to flow during severe droughts which indicate their strong association to groundwater discharge. The Ntuze research catchments in the Ngoye hills (Fig. 4.1) have seldom stopped discharging, even under extreme droughts. To the best of our knowledge, the flow in the small catchment monitored by W1M031 has never stopped flowing while W1M016 has been known to have ceased flowing for short periods on only two occasions during extended drought periods since 1980. This indicates that the flow has a very significant groundwater contribution. Similarly, a small stream monitored in Alton, Richards Bay (Fig. 4.2) on a continuous basis during the severe drought of 1995 never ceased flowing when many of the larger rivers in Zululand had no measurable flow. This case study examines the flow components of the Ntuze River and the Alton Stream in Richards Bay to determine the extent of the groundwater contribution to streamflow in two different hydrogeological settings through an analysis of the runoff hydrograph components.

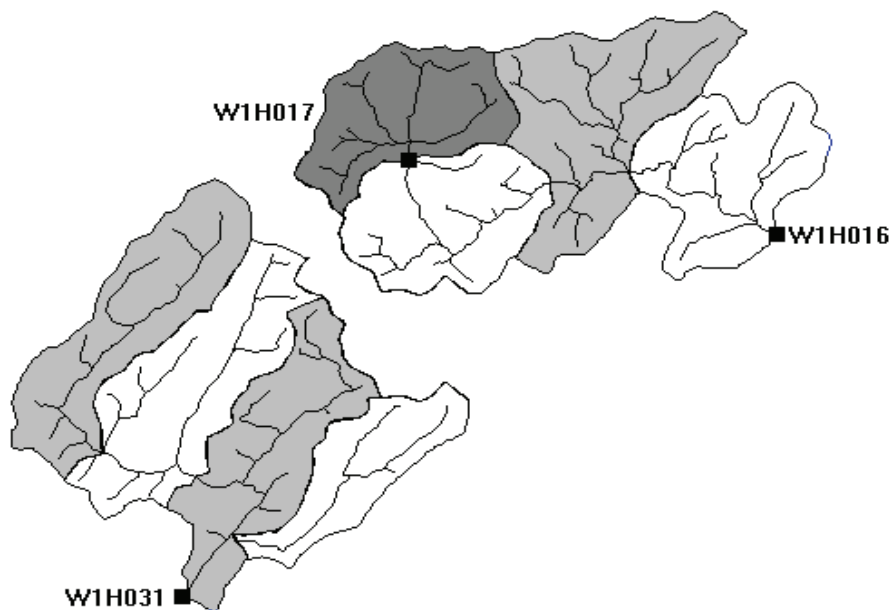


Figure 4.1: River networks and sub-catchments of the Ngoye research study area

Location and Setting

The Ntuze catchments in the Ngoye Hills are shown in Fig. 4.1 while the Richards Bay catchment is shown in Fig. 4.2. The Ntuze catchments are both $\sim 3\text{km}^2$ in area while the Richards Bay catchment is considerably smaller and harder to define but has an area of about 1.5 km^2 .

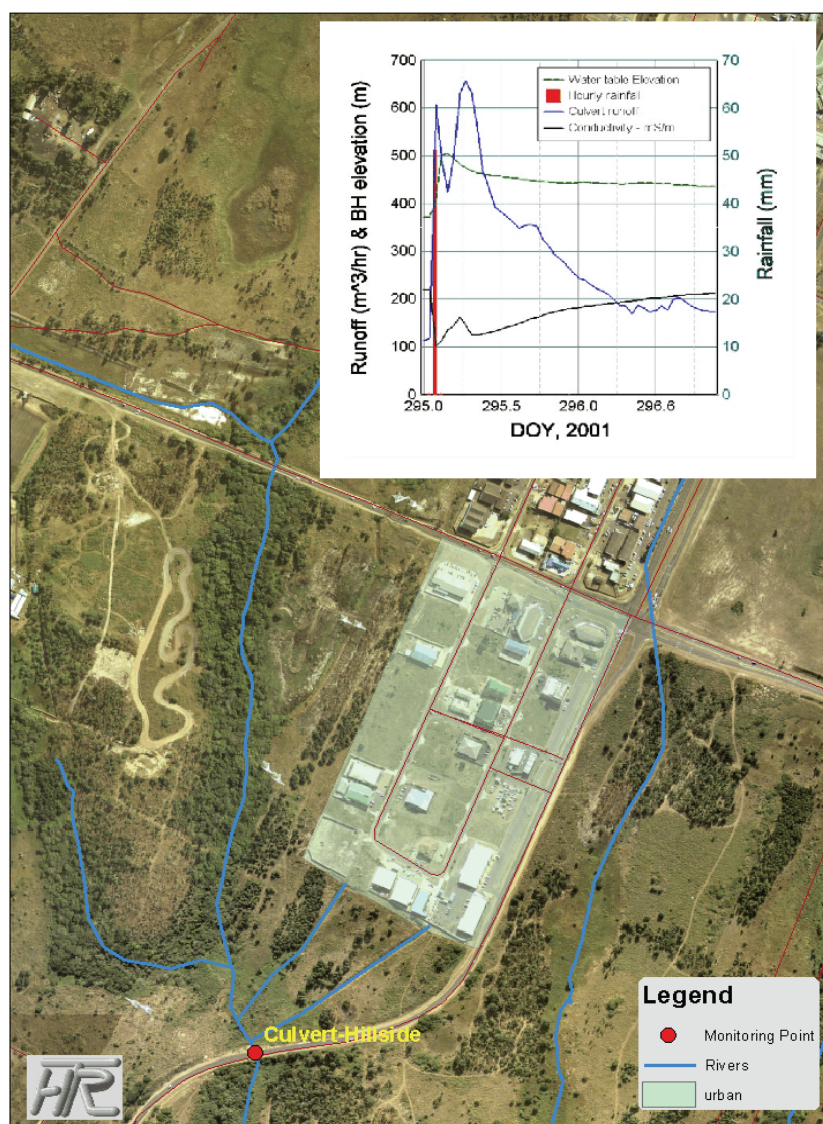


Figure 4.2: The catchment area and surrounding hydrological features for the Richards Bay site in Alton

The Ntuzi catchments have steep gradients (Fig. 4.3) while the Richards Bay catchment is very flat. The soils and geomorphology of the Ntuzi catchments have been described in detail by Kelbe *et al.* (1992) and Kelbe and Snyman (1993).

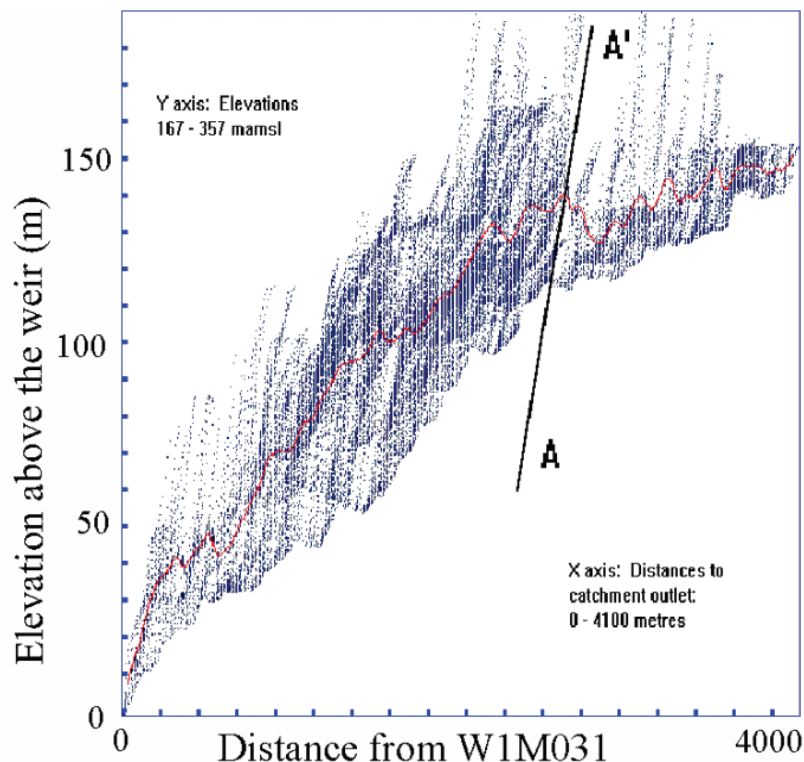


Figure 4.3: The elevation vs. distance from the outlet. AA' represents a typical slope profile

Hydrograph Analysis

Storm runoff has distinct characteristics that have been used to parameterize specific features of the hydrograph. The time to peak and the recession rates traditionally have been used to model storm hydrographs for specific rainfall events. This case study looks at the recession rates of storm hydrographs to determine the different components of the runoff from small catchments.

The recession component of a hydrograph is related to the actual depletion of storage water in the catchment, and it is expected to follow an exponential decline of the form given by Schultz (1978):

$$Q_t = Q_0 \cdot K_r^t$$

where:

Q_t = discharge at time t

Q_0 = discharge measured t days earlier

K_r = recession constant

t = time in days between Q_0 and Q_t

K_r can be evaluated using those parts of the hydrograph plotting as straight lines on a semi-log plot

Alternatively, K_r can be evaluated for $t=1$ day using:

$$K_r = \frac{Q_1}{Q_0}$$

The storm hydrograph usually comprises three components, i.e. quick flow, intermediate flow (also called throughflow or macropore flow) and baseflow (groundwater seepage), with each component having a characteristic recession rate for the catchment.

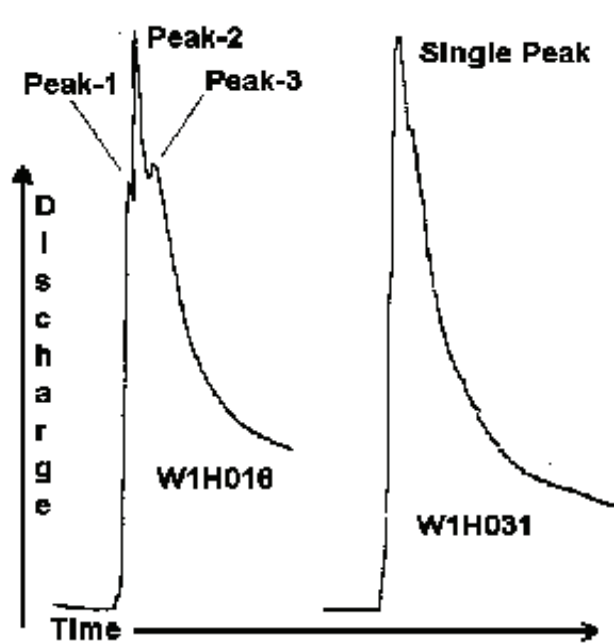


Figure 4.4 Linear plot of runoff hydrograph for two catchments following a short-duration high-intensity storm on 16-11-1976

The recession constants can be evaluated by analysing the recession curve beginning a long time after the peak when it is obvious that only baseflow is contributing to the discharge. A typical storm hydrograph for a short duration, high intensity storm is shown in Fig. 4.4 for the two catchments in the Ntuze Basin. The hydrograph analysis using Eq. (1) for a typical storm for W1H017 is shown in Fig. 4.5. The baseflow component includes the diurnal oscillations assumed to be due to evaporation. Variations in the storm hydrographs and their analysis are discussed further in this case study. The derived recession coefficients for all storms from 1976 to 1986 are given in Fig. 4.6.

Ntuze Morphological Analysis

Mulder and Kelbe (1993) have described the geomorphological characteristic features that influence the shape of the typical storm hydrographs in the Ntuze catchments. Kelbe *et al.* (1991) have analysed many of these storms and identified three distinct components that have been linked to separate hydrological pathways that are predominantly associated with surface, unsaturated soil and groundwater conditions described in earlier section of this report. These components of the stormflow are considered distinct features that can be derived from spatial information of these catchments.

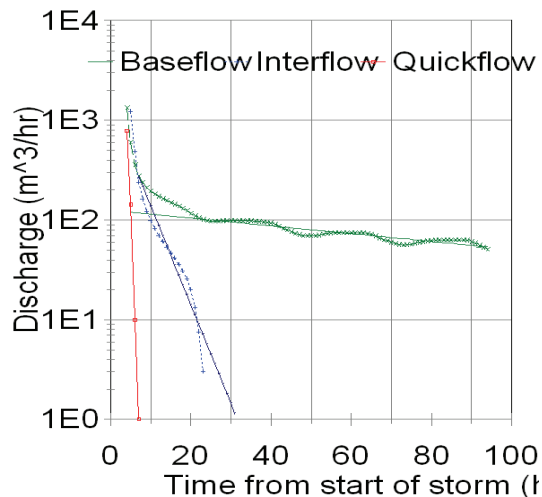


Figure 4.5: Hydrograph recession analysis for W1H017 using the model in Eq. (1).

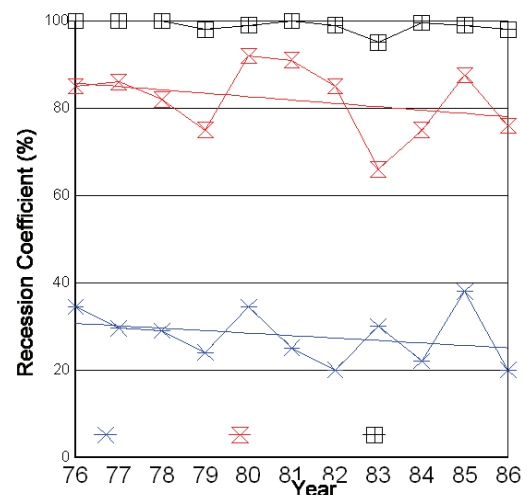


Figure 4.6: The derived recession coefficients for all available storms in the Ntuzi catchments.

There is a large difference between the two research catchments in both the volumetric discharge and the water quality which is a result of the morphological and land use differences. A detailed description of the discharge hydrograph, sediment graph and chemograph has been presented by Kelbe and Germishuys (1996). This case study focuses on the discharge hydrograph to determine the contribution of the groundwater component.

The typical storm hydrographs from both catchments for a short duration high intensity storm on 21 January 1994 are shown in Fig. 4.7. The cumulative rainfall distribution is almost identical but the hydrological response is very different. The response at W1H031 is very rapid and is sharp peaked. The response from W1H016 is not as sharp and shows some discontinuities which are examined in more detail in this section.

The hydrographs at W1H016, from short duration, high intensity storms show several distinct peaks (Kelbe and Germishuys, 1996; Kelbe *et al.*, 1991) which have been attributed to the geomorphic features of the catchment by Mulder and Kelbe (1992). These storms generally produce triple-peak hydrographs for spatially uniform rainfall. These hydrographs can be considered as the convolution of three separate unit hydrographs emanating from three specific areas of the catchment.

Model of Unit Hydrograph for W1H016

The hydrograph emanating from a catchment reflects the integrated effects of all the hydrological processes operating in the catchment in response to a specific rainfall event. The discharge hydrographs from the research catchments have been analysed according to Chow *et al.* (1988) and Herrmann *et al.* (1987) in order to identify aspects of the hydrological processes dominating the runoff from the small research catchments. Distinctly different unit hydrographs have been identified for the two research catchments representing the natural undisturbed system (W1H031) and a catchment with subsistence agricultural settlements (W1H016) which is considered to be typical of large sections of Zululand (Mulder

and Kelbe, 1992). Typical discharge hydrographs from both catchments for a short-duration high-intensity storm in are shown in Fig. 4.7.

W1H031 Hydrograph

The 3.18 km² catchment draining through W1H031 has geomorphic features (Kelbe *et al.*, 1992) that obey Horton's Laws of stream number and stream length (Chow *et al.*, 1988). The main tributary is 4.00 km in length and the catchment has a perimeter of 8.57 km. The stream network can be divided into the four relatively equal-sized sub-catchments shown in Fig. 4.1. Each sub-catchment is separated by a stream length of about 500 m that would have a very short response time so that the discharge from each sub-catchment is effectively routed in parallel. This would suggest that there would be little separation between the runoff hydrographs from each sub-catchment and that the hydrograph from W1H031 would normally exhibit a single sharp peak as shown in Fig. 4.7.

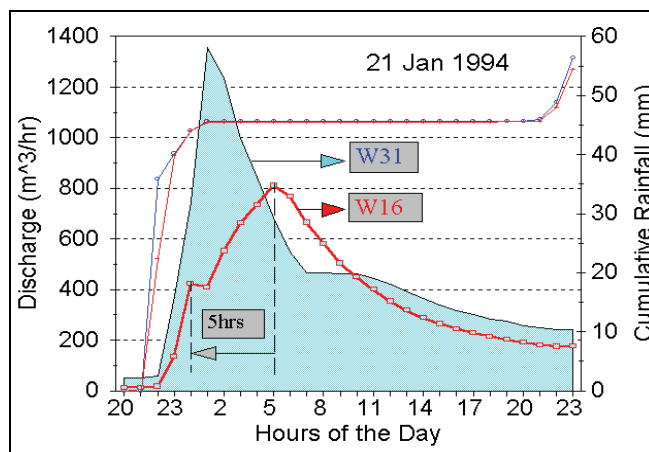


Figure 4.7: Comparative discharge and rainfall for W1H016 and W1H031 for 21 Jan 1994

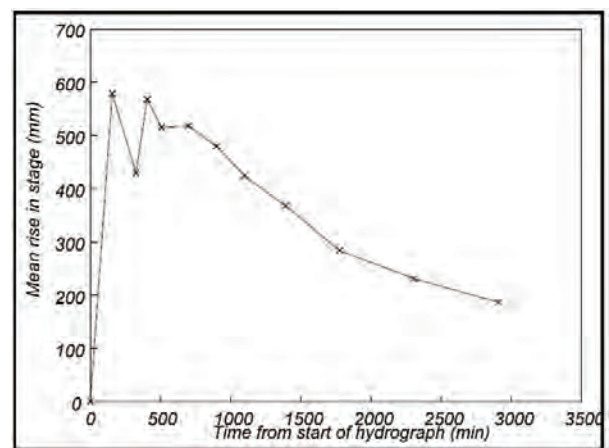


Figure 4.8: Characteristic storm hydrograph from W1H016 for short-duration high-intensity storm

W1H016 Hydrograph

W1H016 monitors a 4th order catchment with an area of 3.23 km² that has very similar Horton Numbers to W1H031 (Kelbe *et al.*, 1992). The main tributary is 4.25 km and the perimeter is 9.55 km. However, the catchment shape is distinctly different to W1H031. In W1H016 there are also four basic sub-catchments shown in Fig. 1 which route the discharge in series rather than in parallel as in W1H031. Consequently, the typical hydrograph from the disturbed catchment W1H016 for short-duration high intensity storms generally show three distinct peaks (Fig. 4.8) which are identified with the different contributing sub-catchments within W1H016.

The autographic records of five short-duration high-intensity storm hydrographs were digitized specifically to determine the characteristic lag times of the three peaks for W1H016. A composite hydrograph was derived from these events through a determination of the mean stage and the mean time between discontinuities (peaks and trough). The composite hydrograph was constructed by normalizing the discharge relative to the initial base flow conditions and setting the time relative to the start of the storm hydrograph. There is considerable variability of the peak flow between storm events. This is due to the incident rainfall volume and the antecedent conditions.

Each peak in the discharge from W1H016 is only identifiable under very specific rainfall conditions that are assumed to produce the rapid runoff response in each catchment. It is suggested that the three peaks in the discharge hydrograph emanating from short-duration high-intensity storms are due to the specific morphology of the catchment. There are two main restrictions in the drainage pattern that are believed to induce separation in the contributing area of the catchment.

The distribution of hydraulic path lengths and associated travel times derived by Kelbe *et al.* (1996) show three distinct peaks (Fig. 4.9) which they have linked to the triple peaks in the discharge hydrograph. If these geomorphic features induce the triple peaks then they may be treated as three separate hydrographs that can be superimposed to produce the final discharge hydrograph at the outlet. Consequently, changes in physical and chemical properties at specific points during a hydrograph from short duration high intensity storms may be linked to specific sub-catchments. This is examined numerically in more detail by Kelbe *et al.* (1996) who present specific case studies for several short-duration high-intensity storm events.

For W1H016, the 1st peak is assumed to be associated with the discharge from the sub-catchment immediately adjacent to the outlet weir (A) in Fig 4.10). The 3rd peak (C = shaded area) is linked to the discharge from sub-catchment W1H017 after it has been routed along the stream for about 3.5 km. The 2nd (B) and 3rd (C) peaks undergo some attenuation during the translation and consequently, the second and third peaks would normally be smaller if they contributed the same volume of runoff. However, the 2nd peak is generally larger than the other two peaks for the same rainfall events and therefore it can be assumed that the contributing area is generally greater. It is assumed that the whole of the 2nd and part of the 3rd sub-catchments in Fig. 4.1 are responsible for the runoff associated with the second peak.

Using standard hydrograph separation techniques (Eq. (1)), the baseflow was removed and the remaining storm flow was modelled as three separate triangular unit hydrographs routed along a stream channel (Fig. 4.10). The recession curve shown in Fig. 4.10 is a composite of digitization points at arbitrary time steps. However, the mean discharge and time of discharge do agree with the model of a draining reservoir described by Eq. (1).

Kr varies for the various reservoirs of surface runoff, throughflow and baseflow. For the storm event the through-flow was estimated to have $Kr=0.35$ and the surface runoff were estimated to be $Kr=0.015$ and the resulting model is compared with the observed storm discharge in Fig. 4.10.

The peak flows will depend on the spatial distribution of rainfall (Fig 4.11). However, there will be some attenuation of the peak flows as the hydraulic wave propagates through the catchment. Consequently, the peak flows will generally decrease for the subsequent peaks if the rainfall distribution is uniform over the catchment. In the analysis it is assumed that the first peak is given by the initial peak (mean) discharge and that the subsequent peak flow rates are halved by attenuation. In order to compensate for the reduced peak flows the recession rates must also be lower so that the total discharges are approximately equal from the different contributing areas and combined to give the actual composite storm discharge.

The storm discharge hydrograph for W1H016 is shown in Fig. 4.8 after the baseflow component has been removed using Eq. (1) with $k=0.3$. This storm produced an average of 35.6 mm of rainfall. By

superimposing the rainfall pattern on the catchment boundaries, the average rainfall for each sub-catchment was derived and the recharge volume estimated (Table 4.1).

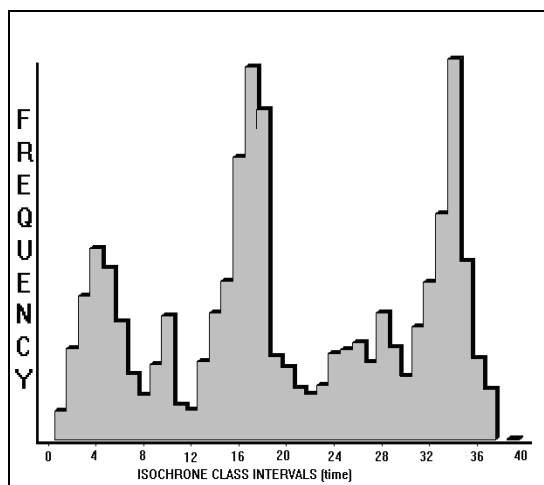


Figure 4.9: Frequency distribution of travel times from the outlet at weir W1H016 (from Kelbe et al., 1996)

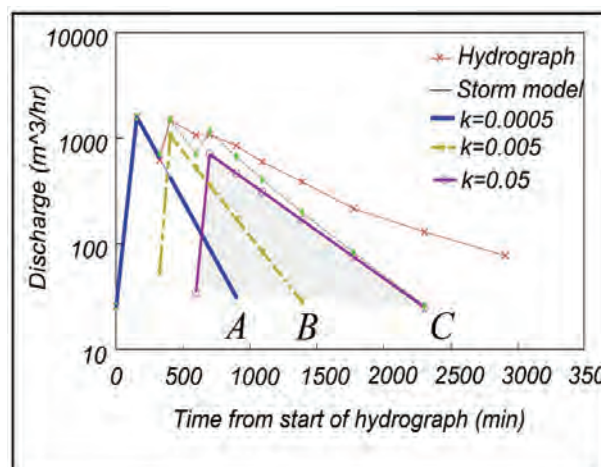


Figure 4.10: Composite model of hydrograph for W1H016 for short duration high intensity storm. A, B & C refer to the triangular unit hydrographs for the three peaks shown in Fig.4.9.

The second sub-catchment (Central (B) in Table 4.1) has the greater recharge but not the greatest peak discharge. Both the theoretical model of triangular unit hydrographs and the recharge model based on rainfall and contributing areas have some agreement in their distribution of runoff volumes but they both are in conflict with the observations and consequently will need further refinements.

Table 4.1: Sub-catchment rainfall

Sub-basin relative to weir	AREA (ha)	RAIN (mm)	Volume (m ³)
Closest (A)	84.6	33.0	27 920
Central (B)	123.3	27.6	34 030
Overlap	27.6	25	6 900
Furthest (C)	84.9	25.1	21 310

Volume of Discharge

For this storm the rainfall was measured and integrated over the whole area to give a total volume of 113 900 m³ (Fig. 4.14). This implies a runoff to rainfall ratio of 4, 8 and 9%, respectively, for the three runoff periods above. This indicates that less than 10% of all the rainfall leaves the catchment as storm runoff. This particular example for a storm in 1992 occurred during one of the driest periods on record when the stream at W1H016 ceased to flow on some occasions. Consequently, it is likely that groundwater storage was an important factor. However, other estimates for 1992 indicate that the storm

runoffs only exceed 10% of the rainfall in the late summer. One individual storm that was found to exceed 10% of the rainfall occurred in November 1989 and produced 218 mm of rainfall that resulted in approximately 180 515 m³ of runoff for a runoff-rainfall ratio of 36%.

Hydrograph Recession Rates: Ntuzze catchments

While the geomorphology of the catchment will influence the shape of the storm hydrograph, another controlling factor is the rainfall duration and intensity. Snyman (2006) grouped all the storms of the Ntuzze catchment into four categories to determine the influence on the shape of the storm hydrographs through their characteristic recession coefficients. The typical recession constants for the Ntuzze catchments derived by Snyman (2006) are given in Table 4.2 and presented for each storm type in Fig. 4.14. These include analysis for long duration, high intensity storms (Fig. 4.14(A), short-duration, high-intensity storms (Fig. 4.14(B), long- duration low- intensity storms (Fig. 4.14(C) and short- duration low-intensity storms (Fig. 4.14(D).

Snyman (2006) derived the recession constant for all the storms in the Ntuzze catchments and derived a representative coefficient for each catchment for each of the storm types which are given in Table 4.2. These runoff hydrographs form the basis of the model presented in Case Study 1.

Table 4.2: Recession constants for each of the catchments from an analysis of each storm type (after Snyman, 2006)

Recession Constants:	High intensity, long duration	High intensity, short duration	Low intensity, short duration	Low intensity, long duration
W1H016: quickflow	0.66	0.55		
W1H017: quickflow	0.59	0.35		
W1H031: quickflow	0.73	0.64		
W1H016: Throughflow	0.81	0.82	0.80	0.79
W1H017: Throughflow	0.86	0.73	0.71	0.77
W1H031: Throughflow	0.92	0.85	0.78	0.82
W1H016: Baseflow	0.98	0.98	0.98	0.99
W1H017: Baseflow	0.97	0.97	0.98	0.99
W1H031: Baseflow	0.99	0.99	0.99	0.98
Mean: Quickflow	0.66	0.51		
Mean: Throughflow	0.82	0.80	0.76	0.79
Mean: Baseflow	0.98	0.98	0.98	0.99

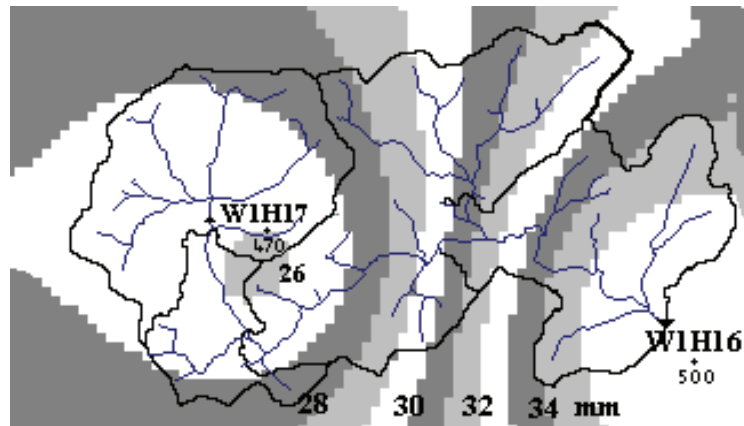


Figure 4.11: Storm hydrograph for DOY 317, 1992 from W1H016

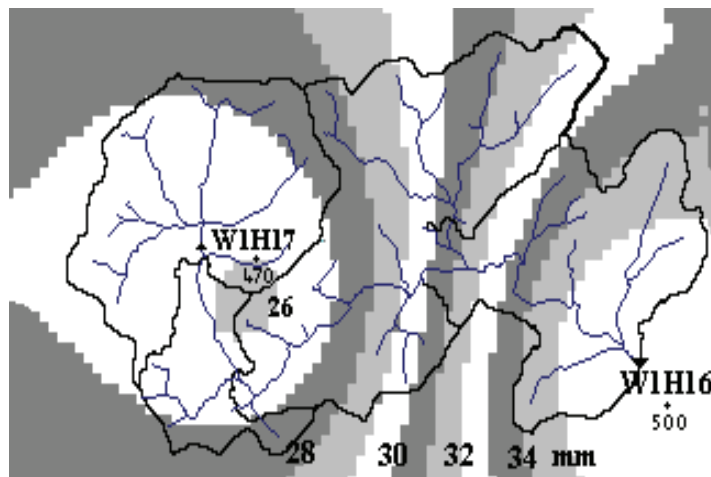


Figure 4.12: Hyetograph of storm on DOY 317, 1992

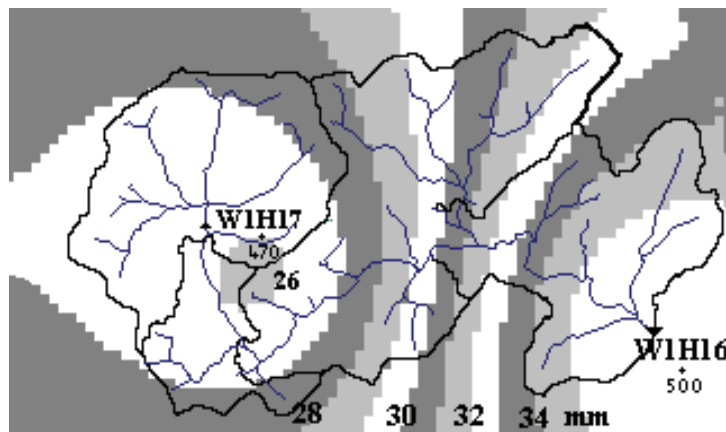


Figure 4.13: Seasonal runoff/rainfall ratio for 1992 at W1H016

Hydrograph Recession Rates: Richards Bay Catchments

A continuous monitoring station was established on the Alton stream (Fig. 4.2) and flow measurements for several months were recorded in conjunction with water table elevations and some

meteorological parameters (including rainfall) at a nearby station in the same catchment. The hourly stream discharge for the stream in Alton is shown in Fig. 4.16 for a two-month period of hourly measurements in 2002.

Figure 4.16 also shows the scaled phreatic surface fluctuations (BH3%) at a shallow borehole within the catchment area of the Alton stream. There is a remarkable similarity in the water level and runoff series which is captured in the scatter plot in Fig. 4.15, illustrating the strong linkage between the groundwater and surface water fluxes in these shallow coastal environments.

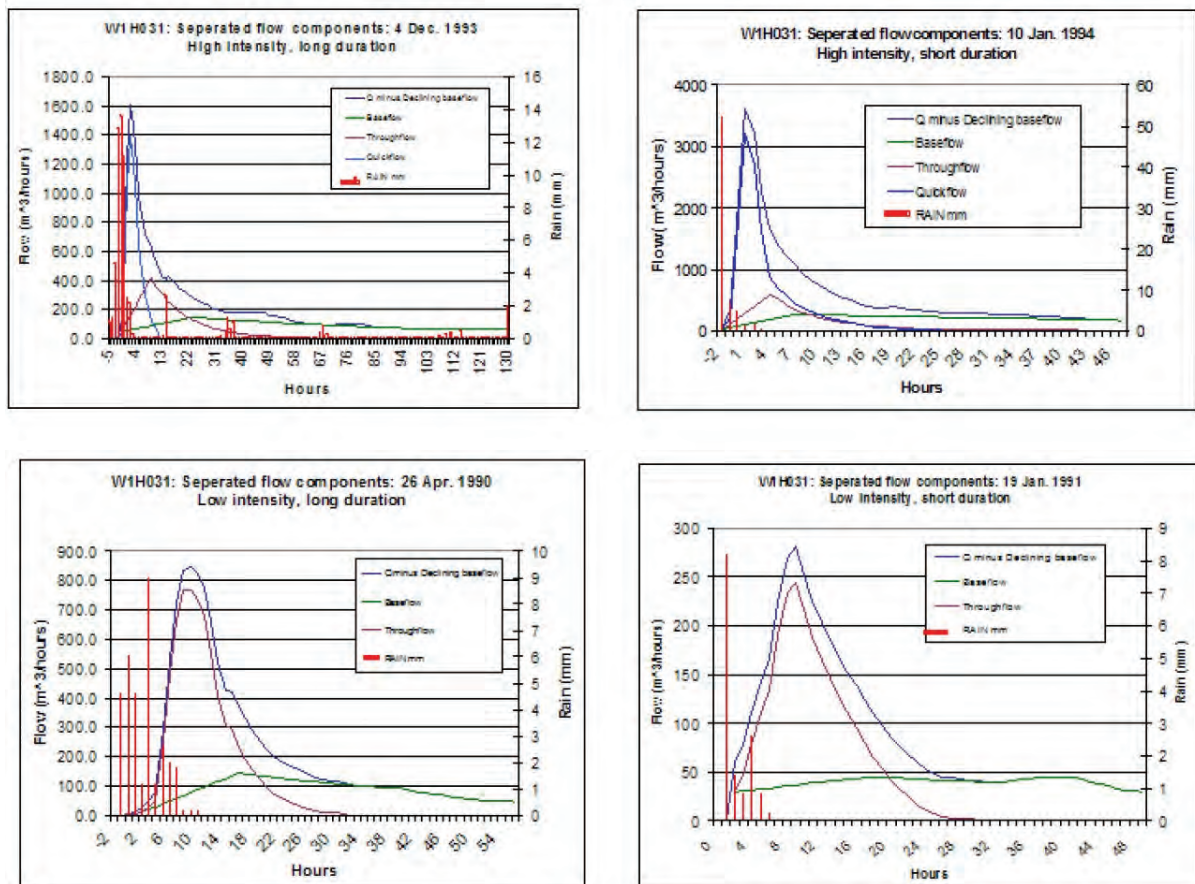


Figure 4.14: Plot of runoff hydrograph for four categories of storm types showing flow components (after Snyman, 2006)

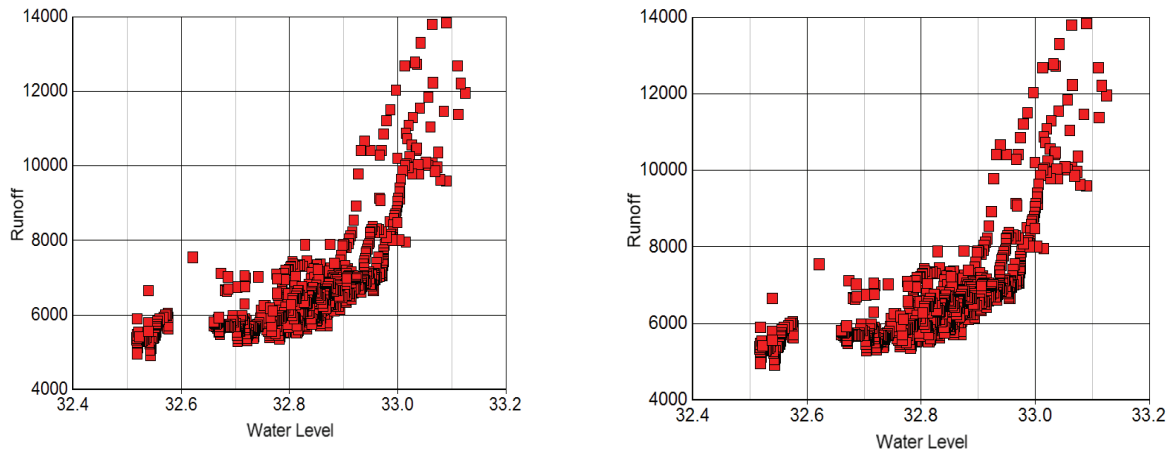


Figure 4.15: A scatter plot of stream runoff against water table fluctuations in the Alton catchment

CASE STUDY FOR DOY 295, 2002

A high-intensity short-duration storm (52 mm/h) in Richards Bay on DOY 296 has provided an excellent case study. The storm produced a peak in the runoff and an absolute minimum in the electrical conductivity (EC) within the same hour as the storm (Fig. 4.17).

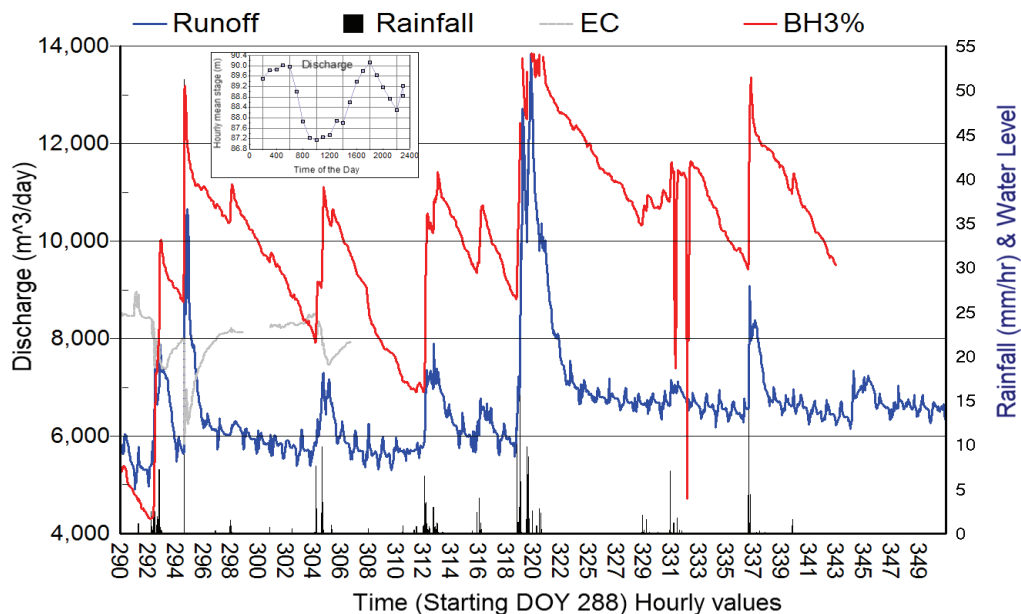


Figure 4.16: The hourly measurements of stream discharge rate, rainfall and groundwater level for a small catchment in Alton, Richards Bay. The inset graph shows the nature of the background noise in the discharge series.

This very rapid response is attributed to a small impervious urban area and paved road network (shown in Fig. 2) that flows directly to the monitoring point. The main peak in runoff occurred 5 h later. However, the water table rose to a maximum within two hours of the storm event. The leaky aquifers at

26 m and 36 m depth responded much later with peak elevations in head after 18 h and 47 h, respectively.

The change in water table elevation and the stream discharge are shown in Fig. 4.16. There are some unknown small-scale fluctuations in the streamflow measurements which are diurnal in nature but not linked to temperature (inset in Fig. 4.16). Long-term averages show peaks at 05:00 and 18:00 and a minimum at 10:00 and 23:00. It is unknown whether there are industrial discharges that produce these events.

Electrical Conductivity (EC)

The EC values shown in Fig. 4.17 clearly indicate the initial overland flow of fresh rain-water (new water) when the conductivity dropped to less than 0.12 mS/m. It increased rapidly to peak with the rising limb of the second wave before dropping again to almost 80% of the first minimum and then it started to rise with the onset of the recession limb.

The relationship between the runoff and EC is similar to the one observed by Kelbe *et al.* (2000) for the Ntuzi catchment.

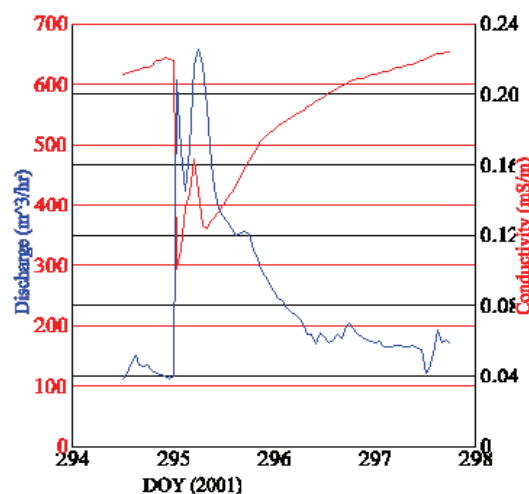


Figure 4.17: Storm hydrograph and corresponding changes in electrical conductivity

In many cases they observed a dual rate of recovery in the conductivity values from the observed minimum which is attributed to the change in flow regime from interflow to baseflow.

This is examined in more detail in the next section.

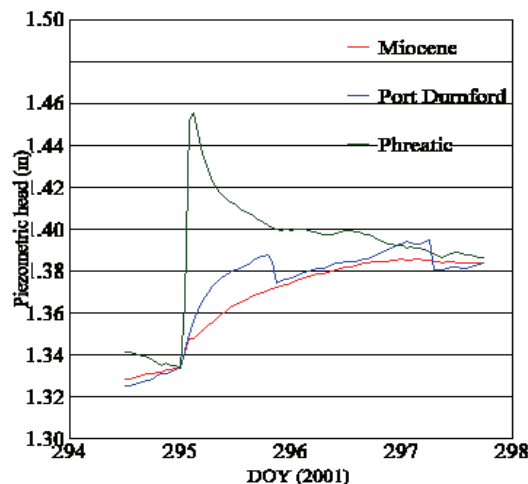


Figure 4.18: Hydrographs for three boreholes in Richards Bay

Groundwater Hydrographs

The phreatic surface responded the fastest to the storm event after the initial flush from the impervious area. However, the deeper aquifers show much longer response times. The cause of the short-term variations in the Port Durnford (clayey sand aquifer) is unknown at this stage.

The cause of the short-term variations in the Port Durnford (clayey sand aquifer) is unknown at this stage. However, the deep Miocene shows a very slow response which only peaks 2 days after the storm event on DOY 295.

Hydrograph Analysis of Streamflow

The storm hydrograph (Fig. 4.19) shows the typical rapid increase in discharge rates followed by three distinct recession rates which were analysed by the classical hydrograph separation technique using Eq. (1) (Schultz, 1976).

The baseflow recession rate shows typical small scale fluctuations. This hydrograph was analysed using a semi-log model (Schultz, 1976) and a linear model (Fig. 4.20). Both models show almost identical fits to the baseflow component of the hydrograph. The linear rate of recession was determined to be 0.966 (-0.37 m³/h) while the semi-log model was 0.957(-0.30 m³/h).

Similar analyses were conducted on the interflow and quickflow components of the recession hydrograph. However, the data does not appear to conform to the semi-log model. The linear model produced a recession rate for the interflow component that was (-10 m³/h).

The quickflow rate was closer to -47m³/h).

The r² values for all three components of the linear model are shown in Table 4.3.

Table 4.3: Linear regression coefficients for streamflow recession component

	Intercept	slope	r ²
Baseflow	191	-0.33	0.28
Interflow	363	-10.2	0.97

The poor fit for this storm event using the traditional methods of Schultz (1976) needs further investigation. It may be an indication of a difference in streamflow response between the coastal systems and the other rivers in the interior.

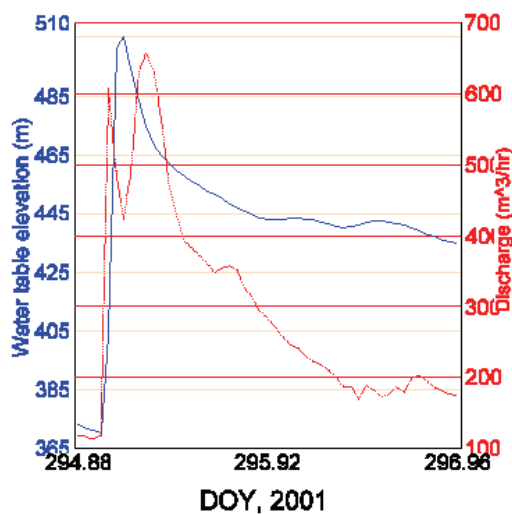


Figure 4.19: Storm and groundwater hydrographs for DOY 295

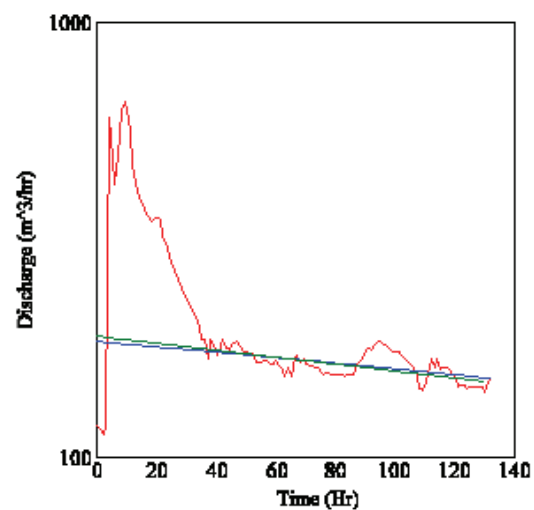


Figure 4.20: The hydrograph recession analysis for the streamflow on DOY 295

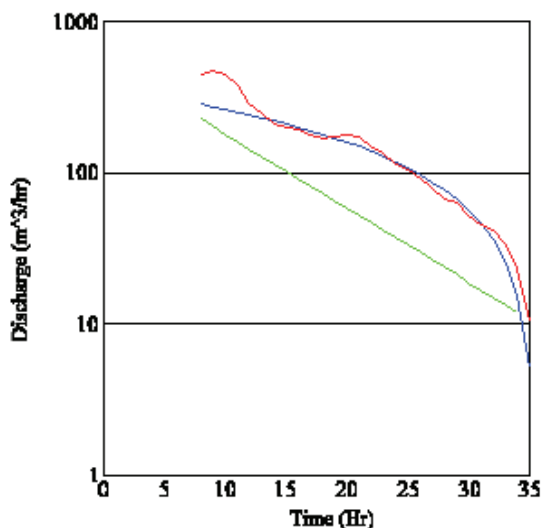


Figure 4.21: Interflow recession analysis

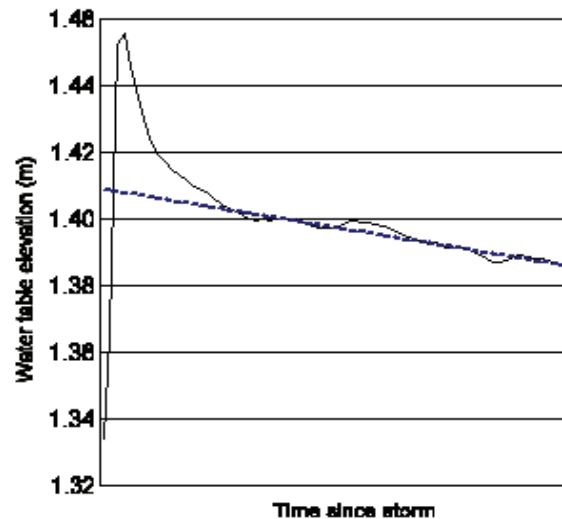


Figure 4.22: The baseflow recession rate for the water table elevation on DOY 295

Hydrograph Analysis of Groundwater on DOY 295

The water table elevation responded much quicker than the streamflow hydrograph. It reached a peak elevation three hours earlier and appears to have receded to 'baseflow' conditions within 15 h of the storm event. The hydrograph analysis for baseflow based on the semi-log model indicated a recession rate of 0.994. This is shown in comparison with the recession curve in Fig. 4.20. The linear model indicated a regression coefficient of -0.0004 m/h, which is considerably lower than the streamflow recession rate.

However, it is necessary to convert these to similar units for a direct comparison. The recession analyses for the 'interflow' and the 'quickflow' components were also done on the water table hydrograph using the semi-log model. The 'interflow' recession coefficient was estimated to be 0.02 and the 'quickflow' rate was 0.001. The plots for both models are shown in comparison with the relevant elevation series (after removal of the baseflow).

Summary and Conclusions

The hydrograph analysis for the storm on DOY 295 integrated over the duration between subsequent storm events, indicated that ~20% of the storm flow was generated from quickflow and intermediate flow (surface processes) while the remaining 80% was generated as baseflow from groundwater discharge processes.

This concurs with the results from the Ntuzi catchments. This is an important feature that has been used to simulate the streamflow using groundwater models by Kelbe (2008).

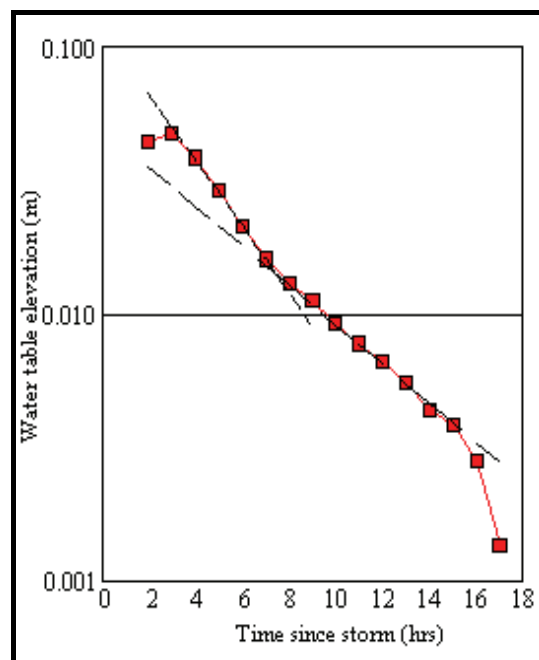


Figure 4.23: The recession analysis for the intermediate and quick recession limbs of the water table elevation on DOY 295

CASE STUDY 5: GROUNDWATER RECHARGE AND DISCHARGE FEATURES FOR SHALLOW PRIMARY AQUIFERS IN MAPUTALAND

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INTRODUCTION

The concept of groundwater recharge is generally based on individual perceptions of the hydrological system and the intended purpose of the definition. Rawlins (1991) and Kelbe and Rawlins (1991) examined the groundwater response to rainfall in the eastern shores region of St. Lucia to estimate the percolation rate (groundwater recharge) and its relation to rainfall. When the interaction between surface water sources and the groundwater are important features of a study, the model that is chosen to estimate the recharge is crucial to the intended outcome. This case study examines the groundwater response to storm events at several locations on the Maputaland coastal plain. It also seeks to relate the groundwater storm hydrograph to stream hydrographs to see if there are similarities that can be described on the basis of preferential pathways described in the model by Snyman and Kelbe in Case Study 1.

Conceptual Model of Shallow Groundwater Recharge

The concept of shallow water recharge is based on the extent to which rainfall is intercepted by surface features, infiltrates the soil and then percolates into the groundwater (saturation zone). This occurs in conjunction with a competing reverse process where groundwater rises by capillary action into the rooting zone (unsaturated zone) to be evaporated back into the atmosphere. The recharge process (Fig. 5.1) is driven by gravitational forces and is relatively faster than the return process where the evapotranspiration rate is driven by the vegetation and evaporative potential (atmospheric demand). A study of the groundwater response at the scale of storm events will show the groundwater rise before the other losses have taken effect.

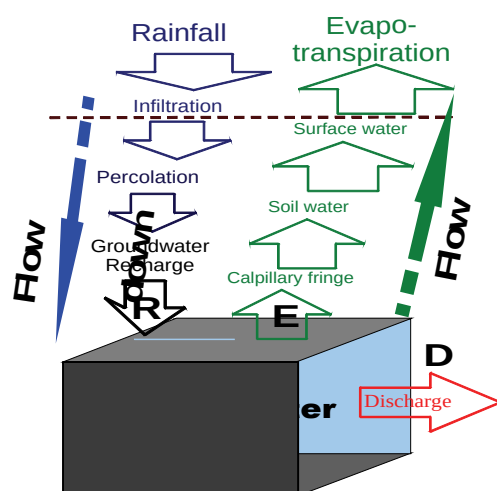


Figure 5.1: Conceptual model of surface water/groundwater interaction processes

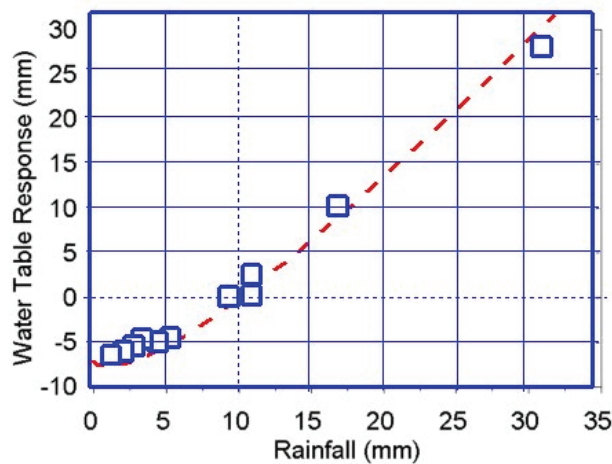


Figure 5.2: The measured water table response to rainfall at St Lucia

Study Sites

The groundwater response to rainfall (surface recharge) was examined at the Alton site in Richards Bay (Fig. 5.2).

This site is located in a region with a shallow primary aquifer composed of recent sands overlying deep (>30 m) marine sediments made up of the Maputaland Group (Fig. 5.3). All the sites were monitored on an hourly basis together with rainfall and other meteorological conditions (Fig.5.2).

The Alton site comprised three boreholes located within 2 m of each other drilled to varying depths of 12 m, 26 m and 36 m to monitor the different formations shaded as yellow, green and grey in Fig 5.4. The site was also equipped with an automatic rain-gauge. This case study examines the reaction of all three boreholes to storm events.

A summary of the screen and water-table depth for the boreholes at these sites is given in Table 5.1.

Table 5.1: Depth and piezometric head for the Alton boreholes

Borehole	Screen depth	Water level (m a.m.s.l.)
BH3	8	32.8
BH2	24	32.2
BH1	36	31.7



Figure 5.3: Continuous monitoring sites used in this case study

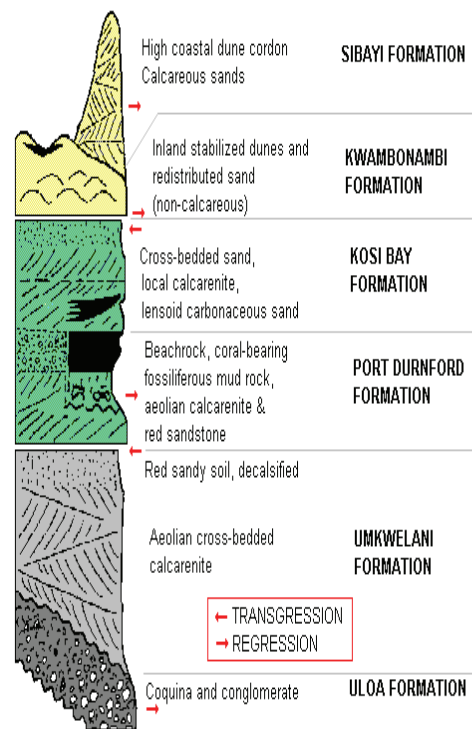


Figure 5.4: Stratigraphy of the Maputaland Group (Botha, 1987)

Recharge Model

The change in the shallow water table elevation in direct response to incident rainfall for the St. Lucia site was presented by Rawlins (1991) and Kelbe and Rawlins (1992) and is reproduced in Fig. 5.5. Their model, for a water table at a depth of <3 m is given by:

$$\begin{aligned}
 &R = \text{rain WHEN } 10 < \text{rain} < 50 \text{ mm} \\
 &\quad \text{ELSE} \\
 &R = 0 \text{ when rain} < 10 \text{ mm} \\
 &\quad \text{OR} \\
 &R = 50 \text{ when rain} > 50 \text{ mm}
 \end{aligned}$$

This model assumed that a rainfall event was the cumulative rainfall over a 5 d period. The 5 d event, with a potential evapotranspiration rate of 2 mm/d (10 mm/5d) accounted for the initial 10 mm loss. All rainfall over 50 mm/d was lost through surface flow processes (local streams and wetlands) when monthly simulations were considered.

This case study looks more closely at this model for the monitoring data at the other sites where the depth to the aquifer depth varies considerably.

Discharge Model

The discharge from shallow primary unconfined aquifers is through lateral flow under a hydraulic gradient and through vertical fluxes during evapotranspiration in the rooting zone. Figure 5.5 depicts the components of evaporation that directly affect the recharge and discharge processes.

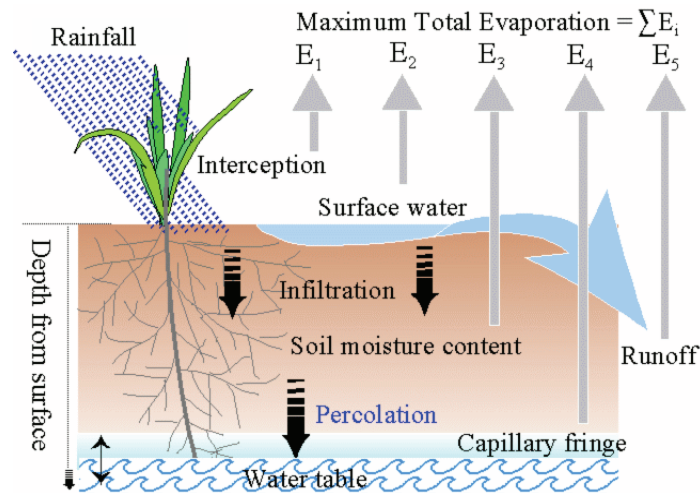


Figure 5.5: Evaporative losses from the various zones affecting the phreatic surface

The evaporative losses from a shallow groundwater system can be estimated using numerical schemes as depicted in Fig. 5.6. The extinction level depicts the elevation of the phreatic surface when it breaks hydraulic contact with the evaporating surface (vegetation). The conceptual model assumes that there is an upper level of the water table where the vegetation is adequately supplied with free water to proceed at the potential rate. This potential rate is not known but assumed to be a proportion of the total evaporative losses (Fig. 5.5). In most cases it would be necessary to calibrate this rate if suitable data are available.

The lateral flow is described by the Darcy model. The flow rate is a function of the hydraulic gradient and the hydraulic properties of the porous media. This case study examines the rate at which the water table rises and falls in response to storm events.

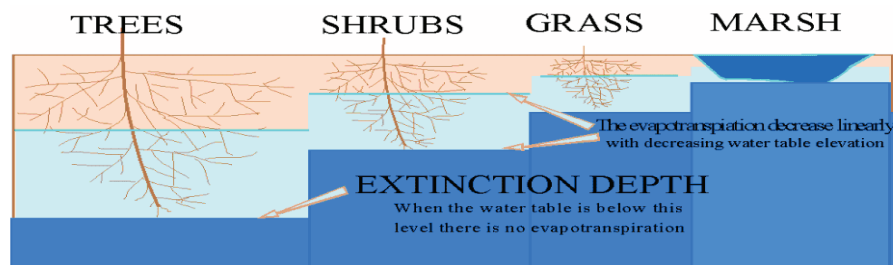


Figure 5.6: Numerical model concepts used to estimate groundwater losses to evaporational processes in numerical groundwater models

Groundwater Response to Rainfall

The response of the groundwater elevation at various depths and time frames is examined.

Long-Term Trends

The water table response to annual rainfall at a monitoring site in Maputaland (BH117:NGWDB) is presented in Fig.5.7.

The water table elevation is closely related to annual rainfall for this station in a form that suggests an asymptotic relationship with an upper maximum that is likely to be related to the topographical features (Fig. 5.4).

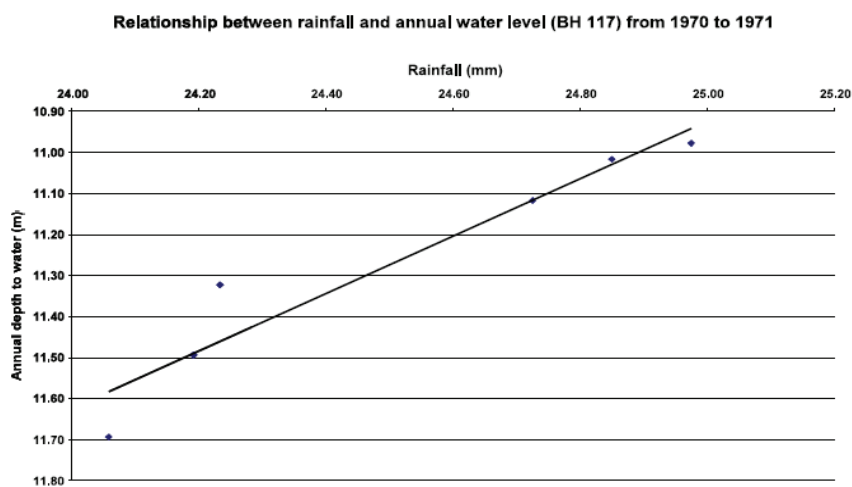


Figure 5.7:
Relationship
between the
annual rainfall
and the water
table elevation
for BH117
(NGWDB)

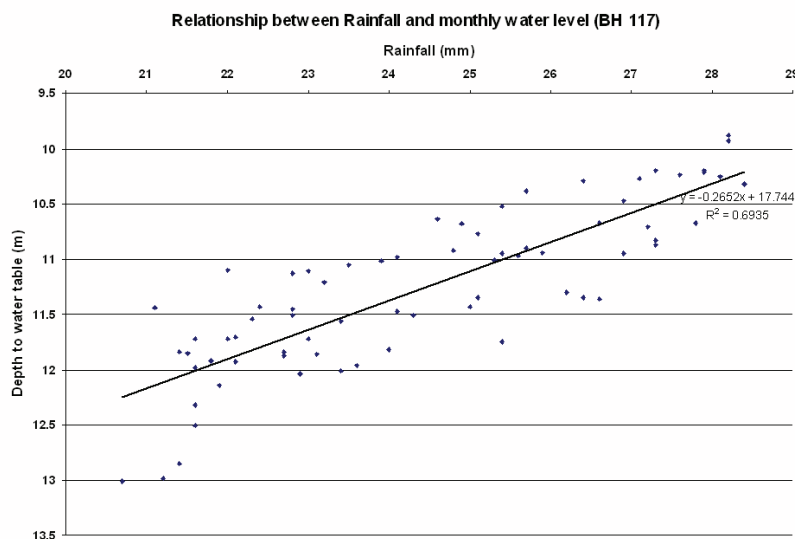


Figure 5.8: The
relationship
between the
monthly rainfall
and the water
table elevation
for BH
117(NGWDB)

Medium-Term Trends

Similar analysis for monthly data is shown in Fig. 5.8. A similar relationship exists at the monthly scale. There appears to be little or no lag in the response to the rainfall at this scale of interaction

although it is highly likely that the rainfall at the end of a month will be more likely to affect the groundwater in the next month.

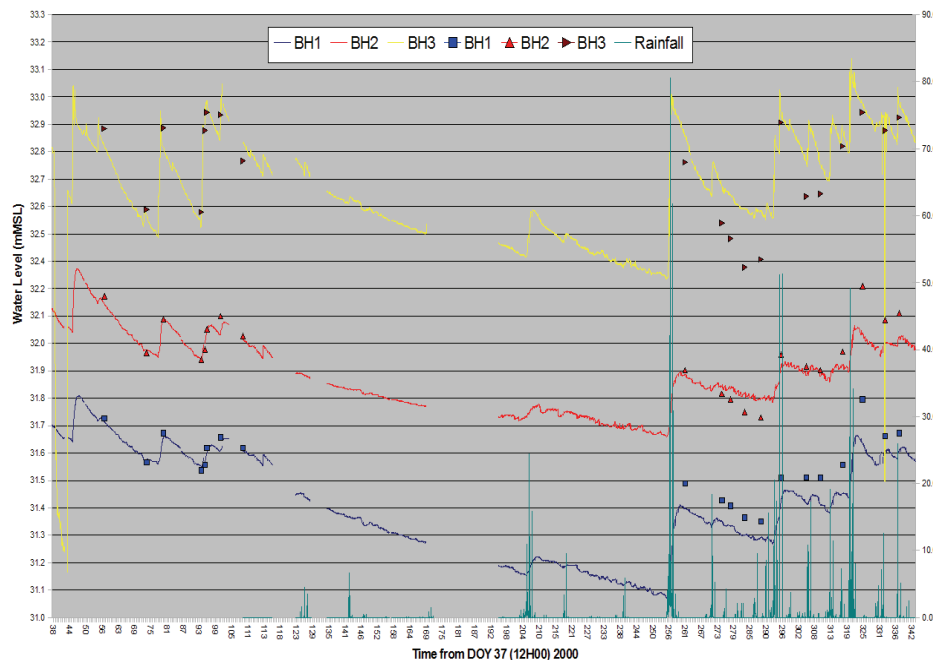


Figure 5.9: The piezometric heads (m a.m.s.l.) for three monitoring boreholes in Alton with screens at depths of 12, 26 and 36 m b.g.l.

Storm Events

The groundwater response to rainfall at the scales of individual storm events (hourly averages) is shown for the Alton boreholes in Fig. 5.10. There is a constant hydraulic head difference of approximately 0.5 between the three boreholes for the two years of observation. The deeper piezometric surface is lower than the phreatic surface by about 1 m. This suggests that there is not a good hydraulic connectivity between these hydrostratigraphic units. However, the response to storm events is clearly evident at all three levels although the peak head is much greater and sharper for the phreatic surface. The figure also suggests that the deeper storm hydrographs are attenuated and translated by several hours.

Hydrograph Analysis

The hydrograph of streamflow (river runoff) was analysed in previous case studies using a simple linear reservoir model proposed by Schulz (1978). The same concept is used in this case study to examine the change in water table elevation (groundwater storage) and stream discharge from the same aquifer.

Shape of the GW Hydrograph

The shape of the groundwater hydrograph differs significantly for the different depths and surface flow (Fig. 5.11). The peak flow for a shallow water table in Alton shows a distinct peak elevation which is not as obvious at greater depths. The phreatic surface reaches a peak about 2 h after the storm event. However, the streamflow (excluding the initial overland flow peak) only peaks about 6 h later (Fig. 5.11).

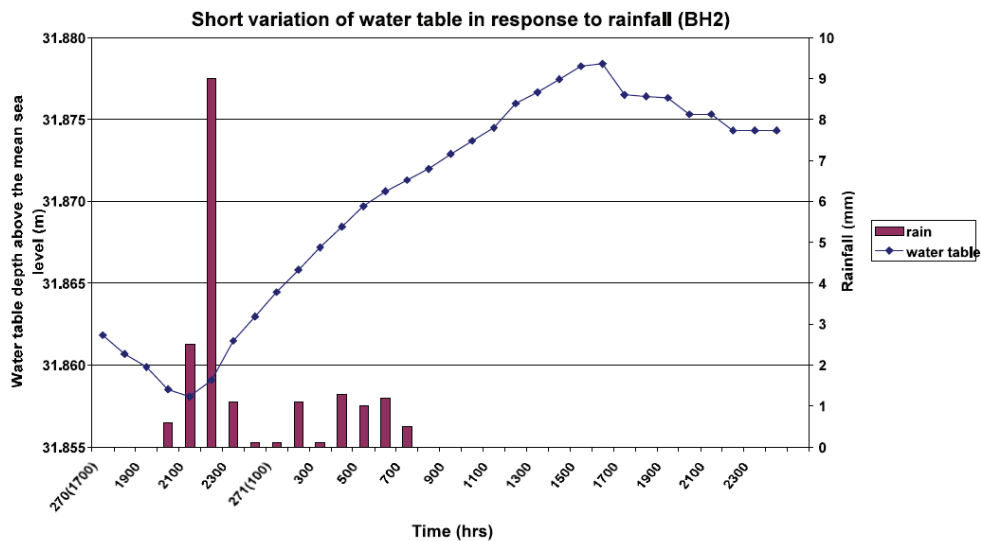


Figure 5.10: Hourly response to storm events at HRU BH3 in Alton

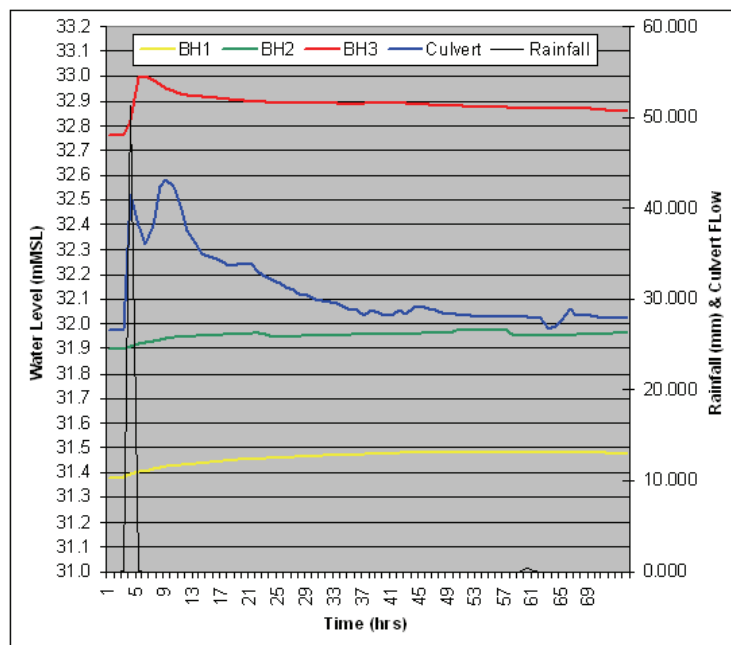


Figure 5.11: Piezometric head at three depths in comparison with stream runoff

The shape of the hydrograph for the shallow groundwater suggests a similar exponential decay as the stream but only reaches a steady state recession rate after about 18 h (Fig. 5.11). The stream discharge from the same storm and aquifer appear to have reached the same stage of recession (baseflow) about 36 h after the rainfall event.

Time to Rise (TTR) of the Hydrograph

The time to rise in the water table following a significant rainfall event gives an indication of the time it takes for the infiltrate to reach the piezometric level. The time taken to 'reach' the water table (or exert sufficient pressure to increase the piezometric pressure) will depend on the depth to the piezometric surface. The time taken to reach a maximum at three depths in the Alton Site for a

significant rainfall event (50 mm/h) is shown in Fig. 5.11. The stream (culvert) had an immediate rise to peak flow for the overland flow from an urban catchment. This was followed by the water table rise of 22 cm for the 8 m screen and then the stream quickflow after a further 4 h. The piezometric pressure peaks at 24 m depth and at 36 m depth after about 18 h and 48 h, respectively.

The borehole logs for the Alton Site show unconsolidated medium to fine sand all the way down to the deeper layers where the Uloa formation becomes coarse conglomerates with high conductivities. This implies a single aquifer with direct hydraulic connectivity between layers. However, a pressure wave at the phreatic surface is heavily attenuated and translated as it moves down into the deeper layers of the system.

The stream discharge during a storm event can reach 40 m³/h (40*24/d).

Recession Rate (RR) of the Hydrograph - Components

The recession rate of the groundwater storm hydrograph can be analysed in the same manner as the river hydrograph in Case Study 4 where the recession is treated as a linear storage reservoir that is depleted at a rate given by Schulz (1978):

$$Q_t = Q_0 \cdot K_r^t$$

where:

Q_t = discharge at time 't'

Q_0 = discharge measured t days earlier

K_r = recession constant

t = time in days between Q_0 and Q_t

On closer inspection of the recession rates for the three depths (Fig. 5.12), the hourly recession rate for the shallow water table appears to be a similar shape to river runoff with possible identifiable components (characteristics similar to river runoff. There is an initial rapid decline after about 4 h, followed by a slower rate for the next 10 h before settling into a constant recession rate over 70 h later.

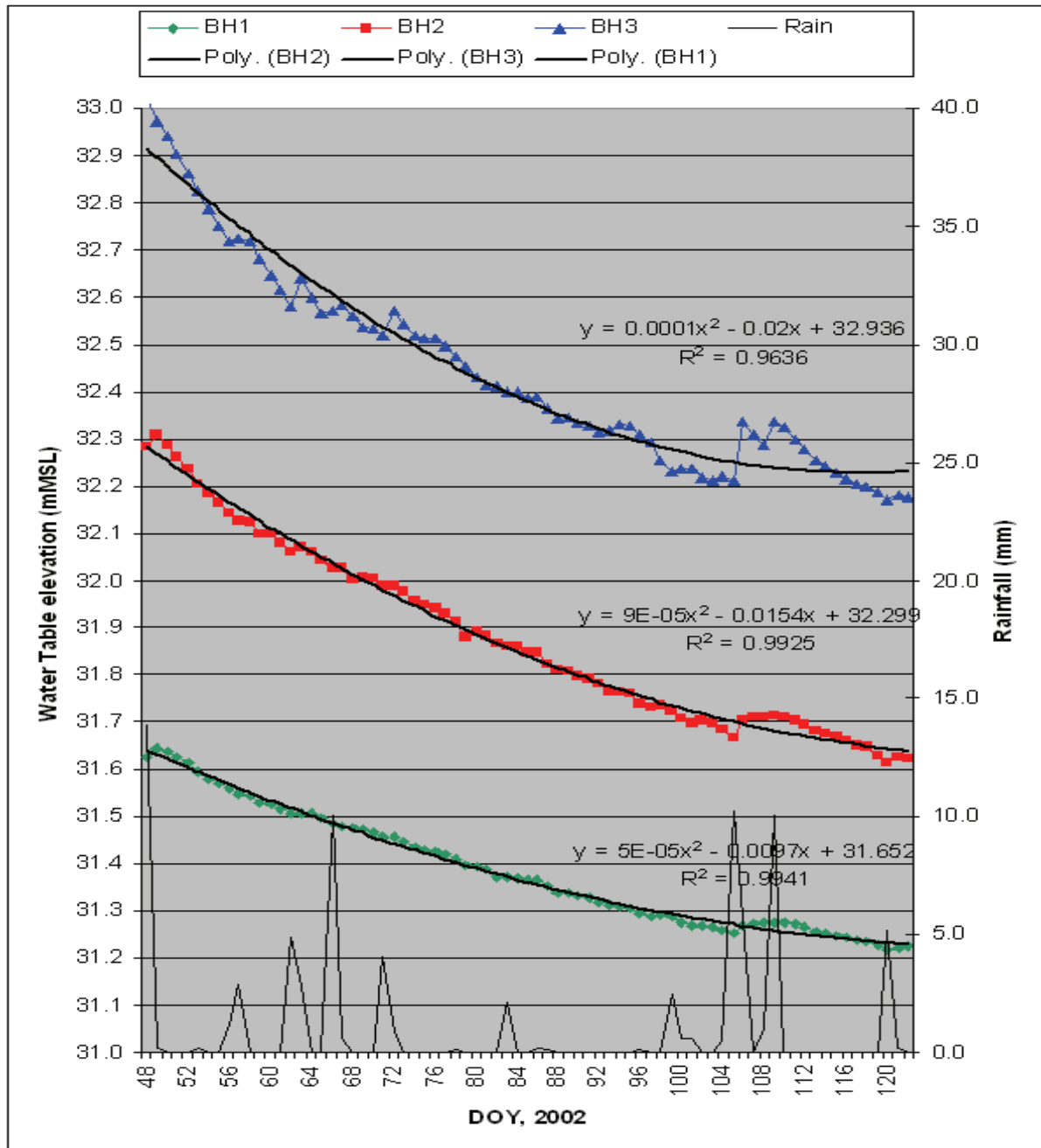


Figure 5.12: Piezometric level recession rates for the three boreholes in Alton

This is a special case of an exponential decay function (equations in Fig. 5.12). The decay coefficient (K_r) for small catchment discharge varies rapidly with cessation of storm events. However, it is expected to be more uniform for groundwater systems (water table elevation).

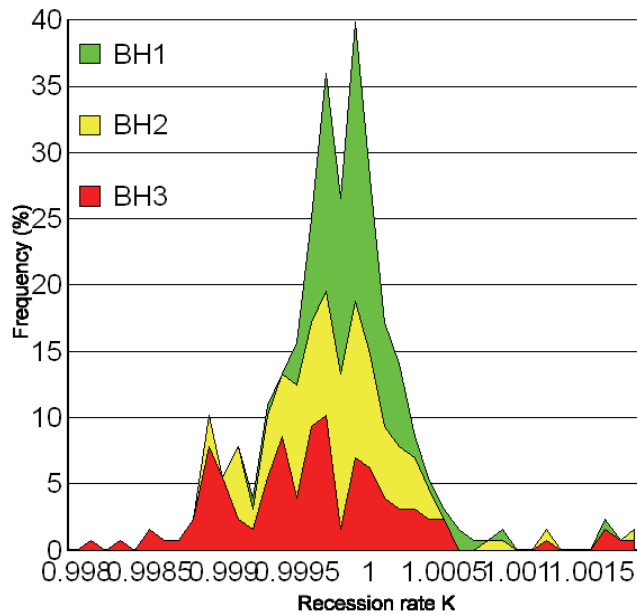
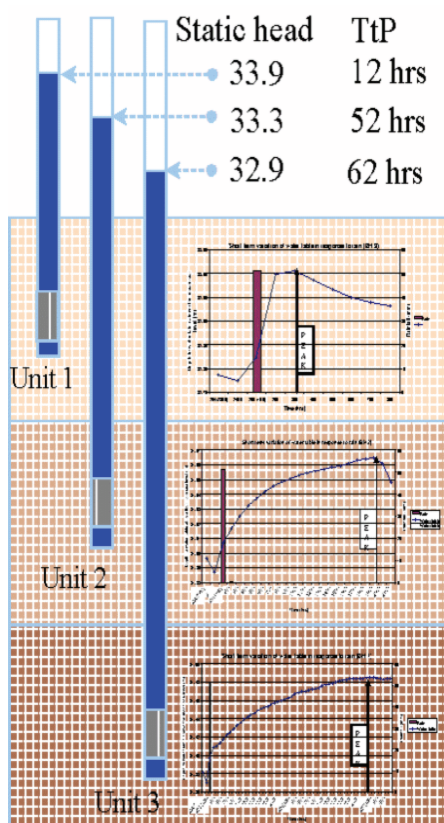


Figure 5.13: The frequency distribution of the rise and recession rates Q_2/Q_1 for the three boreholes in Alton



The frequency of the daily recession rate (K) for the period shown in Fig. 5.12 is presented as a histogram in Fig. 5.13. These rates are very much larger than any observed for base flow which is typically between $0.8 < K < 0.9$. If the recession rate for the water table is directly related to the storage depletion through baseflow in drainage boundaries, then the groundwater recession must be related to the baseflow recession.

Figure 5.14: The configuration of the Alton monitoring site and the composite of all storm hydrographs for the two-year monitoring period

CASE STUDY 6: THE DEVELOPMENT OF A GROUNDWATER MODEL FOR EVALUATING THE INTERACTION OF A COASTAL LAKE AND FEEDER STREAMS WITH THE UNDERLYING AQUIFER

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INTRODUCTION

Water resources in most areas of South Africa are very limited. Along the east coast of Southern Africa, there is heavy reliance on the coastal lake systems for freshwater supply. These natural lake systems have been over exploited and many restrictions have been placed on their abstraction limits in recent times. Very little is known about the dynamics of these systems which are believed to be extensions of the local aquifer. Exploratory studies of several lakes in the region have been undertaken by Kelbe and Germishuyse (1999 and 2000). These lakes are heavily reliant on the groundwater through the persistent baseflow in the small feeder streams and the direct seepage along the lake shoreline. Consequently, this study used a numerical groundwater model to estimate the hydrodynamics and flow components of the lake mass balance of Lake Mzingazi, one of the principle water supply lakes in the Richards Bay area.

The study area lies on an extensive primary aquifer that stretches along the east coast of South Africa. It is part of an extensive low-lying plain that stretches along the coast of Mozambique. The flat coastal plains consist mainly of highly permeable sands that encourage rapid recharge to the shallow aquifer. Exposed wetlands and lakes are abundant in the region, many of which are remnants of estuarine systems that have become closed during recent geological periods forming freshwater lake systems. These freshwater lakes, because they are generally considered to be surface extensions of the primary aquifer, are likely to exhibit varying flow through dynamics (Townley *et al.*, 1993).

A numerical study of Lake Mzingazi is presented to examine the concepts and assumptions readily employed in studies of dynamic lake system in the region where there is very limited information. The study used MODFLOW, a finite difference model described by McDonald and Harbough (1983), with several additional modules simulating internal boundary conditions that describe the interaction between the surface and groundwater components.

Study Area

Lake Mzingazi is a natural freshwater lake system whose outlet was raised to meet local water demand. The lake comprises two compartments that have been isolated during extreme drought conditions (Fig. 6.1).

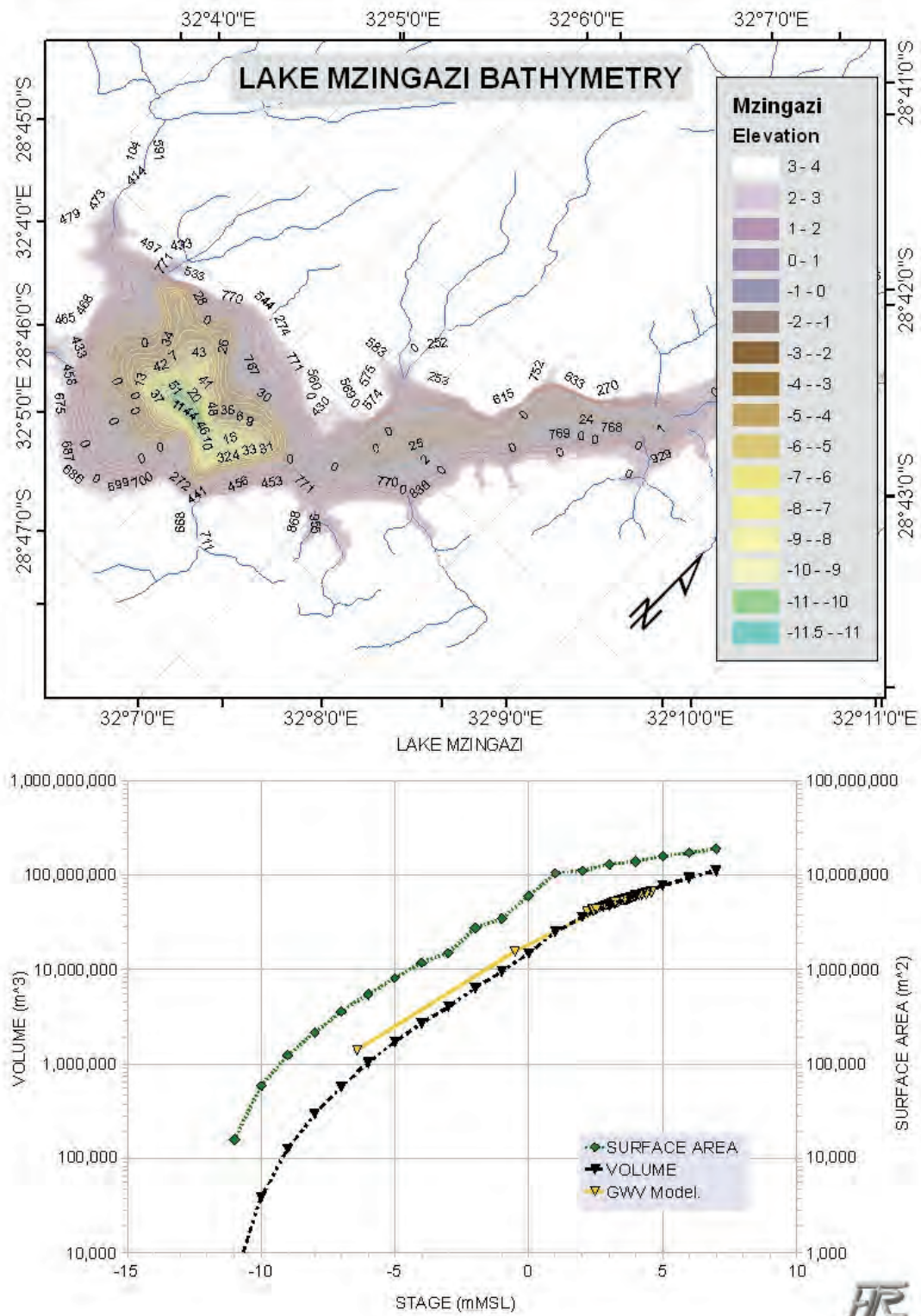


Figure 6.1: Lake Mzingazi bathymetry

The deepest compartment extends 12 m below mean sea level (m b.m.s.l.) and is protected from saline intrusion by maintaining a freshwater wetland between the lake and the saltwater estuary (Rawlins *et al.*, 1977).

Conceptual Geological Model

The lake lies over primary aquifers comprising deep marine and Aeolian sands that form the main aquifers (Worthington, 1978). These sedimentary layers overly a deep Cretaceous marine deposit of siltstone that is assumed to represent the lower boundary of the study area. This boundary formation has been deposited and incised by alluvial processes during transgressive and regressive marine conditions that have left significant palaeochannels in the region. The upper surface elevation for the Cretaceous sediments is shown in Fig. 6.2 and indicates a deep palaeochannel running through the deepest part of Lake Mzingazi.

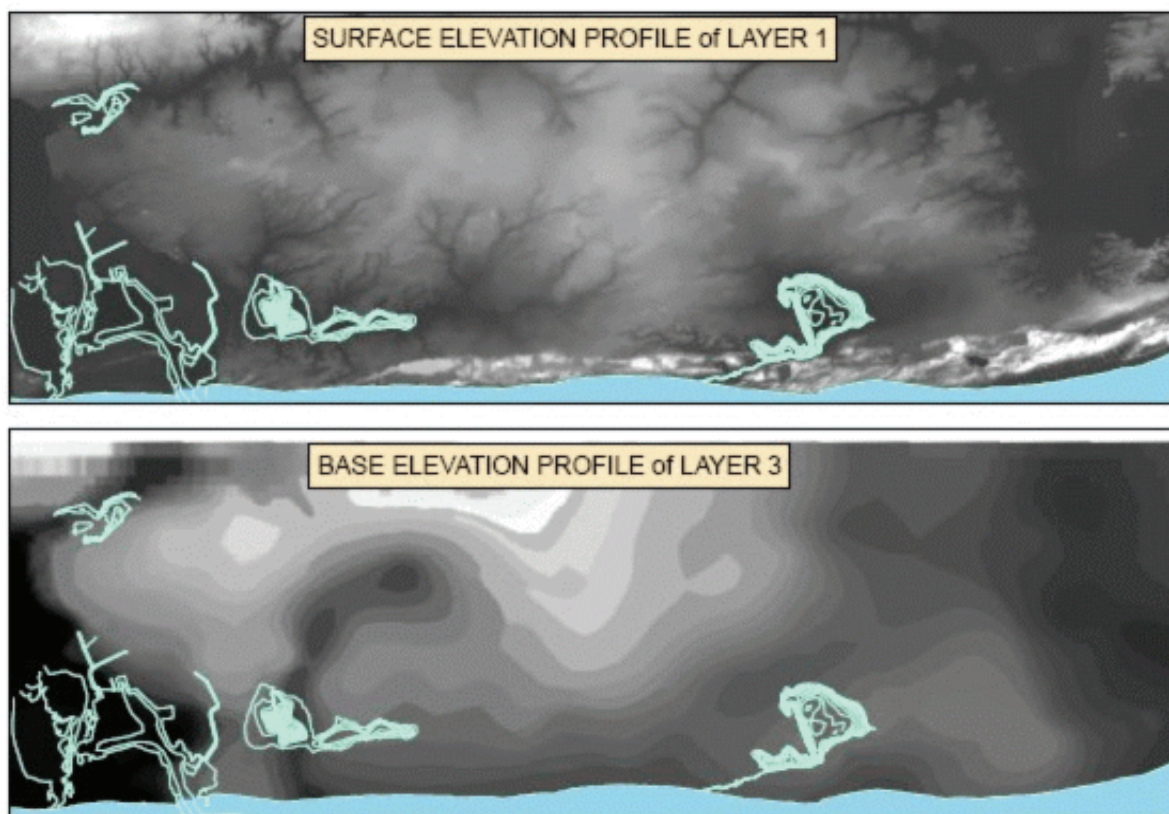


Figure 6.2: The upper and lower surface profile of the primary aquifer

Overlying the Cretaceous siltstone are intermittent layers of coarse sandy calcareous conglomerate of the Miocene age (Fig. 6.3). The spatial extent of these deposits was estimated from an inadequate borehole network by Germishuys (1999). The location of the borehole network and the spatial extent of these formations are shown in Fig. 6.3. The deposits are intermittent layers that generally occur on the higher regions of the Cretaceous siltstone formation suggesting that they were also eroded with the Cretaceous Mudstone during periods of transgression in the sea level.

The Cretaceous and Miocene deposits are overlain by more extensive layers of clayey sands of marine origin and more recent Aeolian cover sands that are highly permeable and promote rapid recharge. These lithological units have been identified from the available borehole logs and mapped in Fig. 6.4 for incorporation into the numerical model.

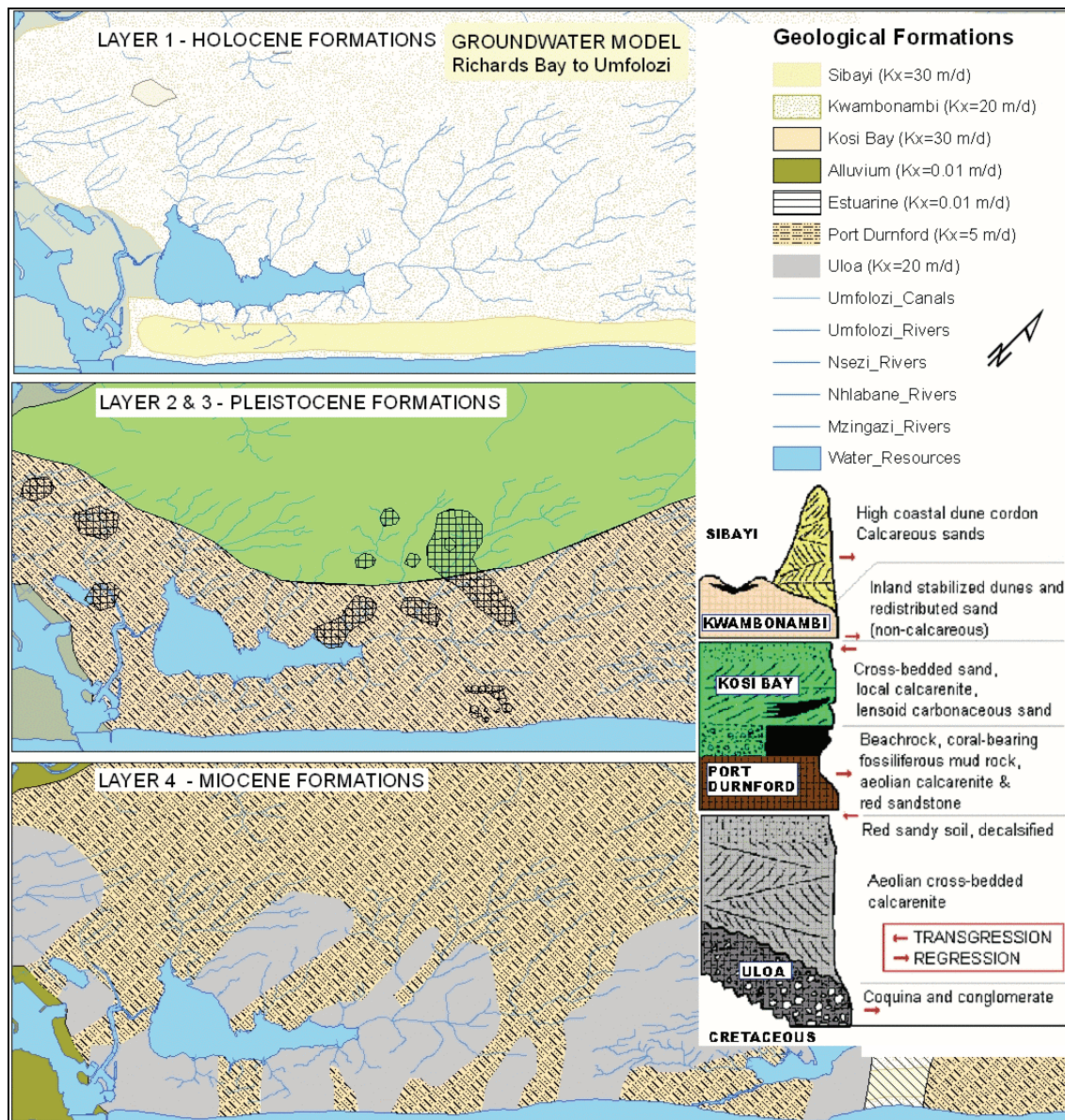


Figure 6.3: The conceptual geological model of Lake Mzingazi area

Numerical Model

The numerical method chosen for this study was MODFLOW 2K. This is a three-dimensional finite difference model that is constrained by specified boundary conditions. A lake package developed by Council (1997) was incorporated to simulate a head dependent boundary that responds to surface recharge and groundwater seepage through lake bed sediments. This package simulates fluctuating lake levels and computes a lake water budget by balancing the changing volume with multiple inflow and outflow processes. The degree of lake-groundwater interaction is controlled by the hydraulic gradient and the lake bed sediment conductance which had to be estimated from calibration. The feeder stream were simulated using the STR1 package (Prudic *et al.*, 2004) that simulated the flux of water trough the stream channel into the lake to provide the inflow component.

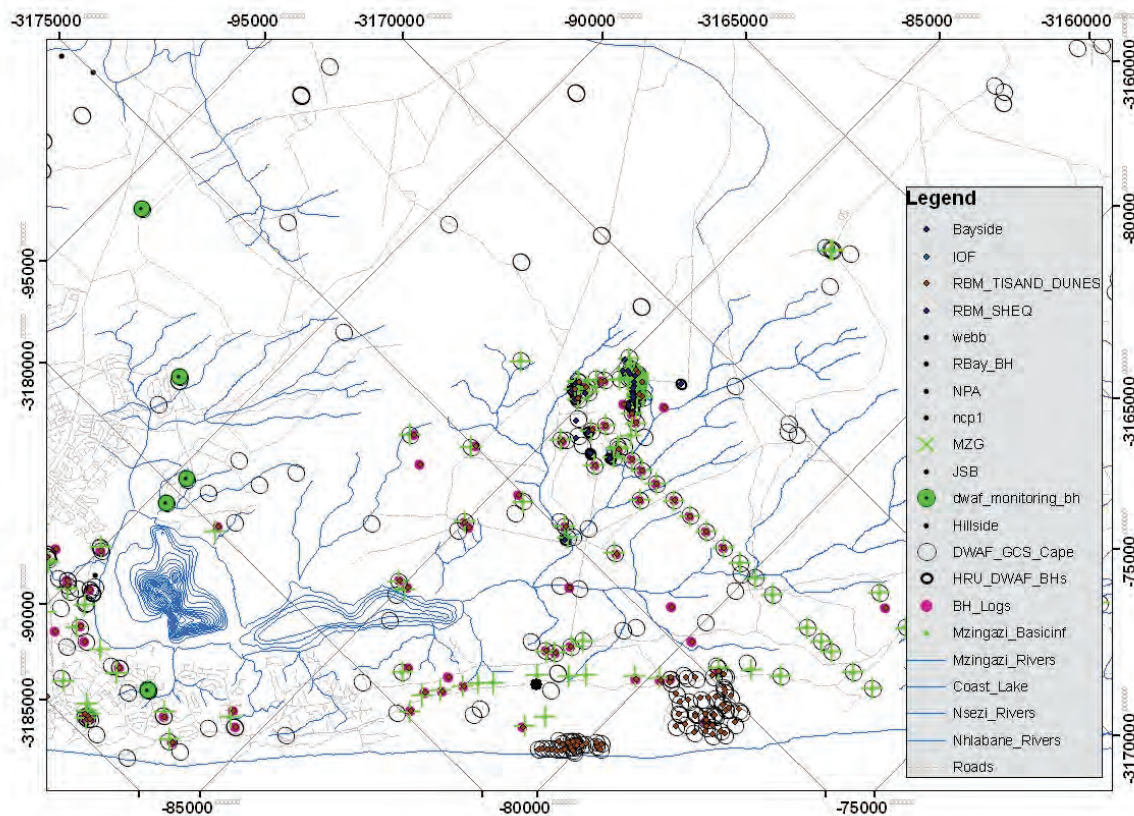


Figure 6.4: The location of borehole and piezometers from various sources with some data for use in the development and calibration of the model

The numerical model is constrained by the Indian Ocean, Mhlathuze Estuary and Lake Nsezi drainage faces which were configured as constant head boundaries on three sides of the model domain. The other side of the model domain was assumed to be a no flow boundary parallel to the piezometric gradient. Vertical flow boundaries were configured to simulate the effect of land use through evaporation from the shallow water table. The rooting depth that regulates evapotranspiration from the capillary zone (for vegetation in hydraulic contact with the water table) was varied according to the land-use type. The vertical recharge rate was assumed to be a function of the land-use type which was allowed to vary over time. The recharge and evapotranspiration rates were estimated using a crude rainfall-runoff model that accounted for surface evaporation from the unsaturated soil profile and interception, and the surface runoff on a daily basis but allowed for additional evapotranspiration when the groundwater was in hydraulic contact with the water table (Kelbe and Germishuys, 1999).

Measurements

The water level elevation for Lake Mzingazi as well as the discharge rate through a sharp crested rectangular weir at the outlet to the lake have been monitored since 1975. Measurements of streamflow within the lake catchment area have been recorded at various times. The rainfall is monitored on a regular basis within the lake catchment area.

Calibration of the Aquifer Model

The hydraulic parameters for the four layer transient groundwater model were adjusted systematically to achieve the best fit between the observed and simulated water table elevation for the available borehole measurements (Fig. 6.5). These water table measurements were assumed to represent normal climatic conditions but may have been obtained under different climatic conditions.

The lake package of Council (1997) was configured for the groundwater model where the surface groundwater interaction would be controlled by the leakage rate through lake bed sediments occupying two of the layers. This lake leakance parameter (or conductance) defines the level of interaction between the aquifer and the lake. For very large conductances the lake would act like an unconstrained boundary which would be controlled by the aquifer properties and hydraulic gradients. For very low conductances, the lake behaves like an isolated system dependent only on the surface recharge and discharge conditions. The area of each cell was known but the thickness and the hydraulic conductivity of the lake bed sediments are unknown. Calibration studies were used to derive a best estimate of these parameter values using observed lake levels (Fig. 6.6) during extreme hydrological events.

The model was further calibrated for the stream function against continuous flow records for a feeder stream in Alton and several ad hoc measurements of stream flow within the lake catchment. The simulated streamflow formed an integral part of the lake water balance which was also used to calibrate the model. This required an accurate bathymetric profile of the lake (Fig.6.1).

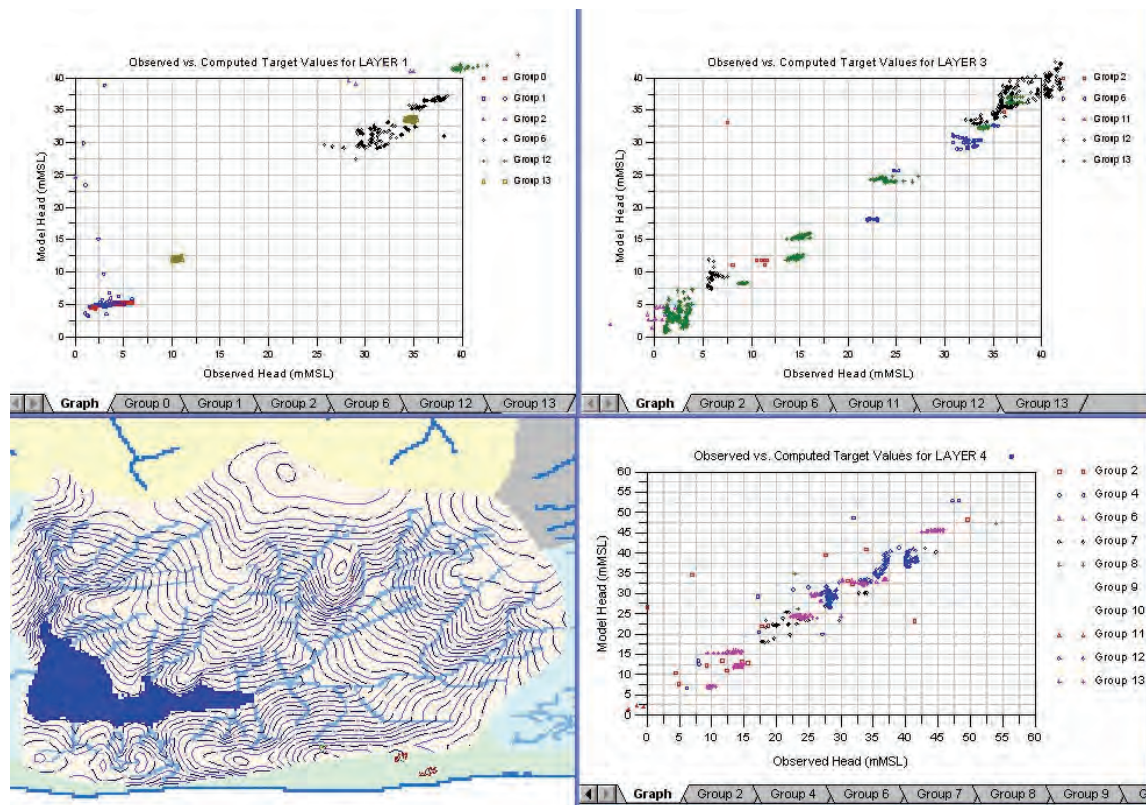


Figure 6.5: The scatter plots of observed and predicted heads for the different layers of the model

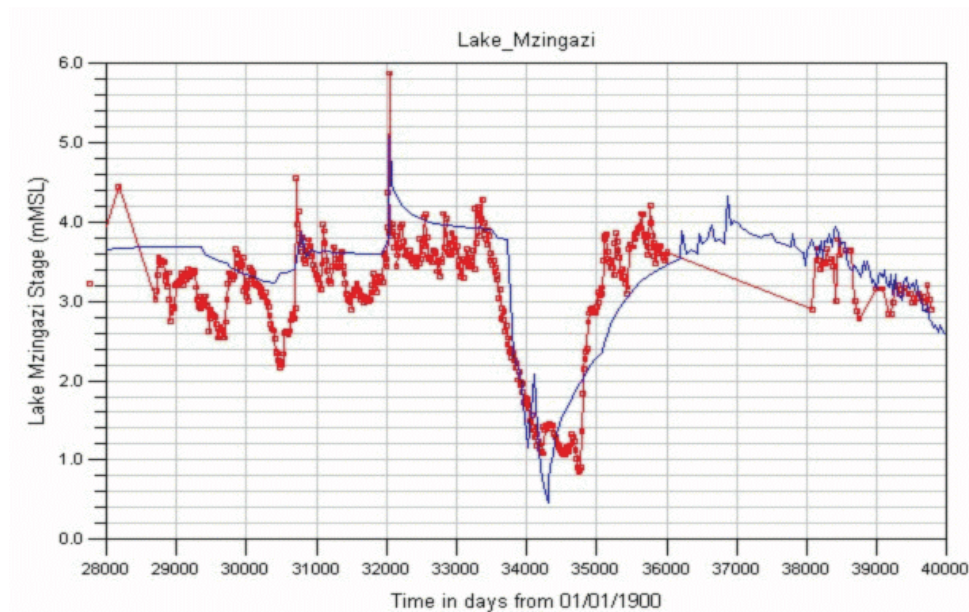


Figure 6.6: The measured and predicted water level elevations for Lake Mzingazi for the simulation period of 12 000 days

Summary

The calibrated groundwater model provides estimates of the water table and piezometric heads for the region (see map in Fig. 6.5). It has been used to predict the streamflow in several of the streams in the region (Fig. 6.7) and to estimate the impact of afforestation on the feeder streams in the region (Fig. 6.8). The model has been used to conduct particle traces and solute transport of pollutants.

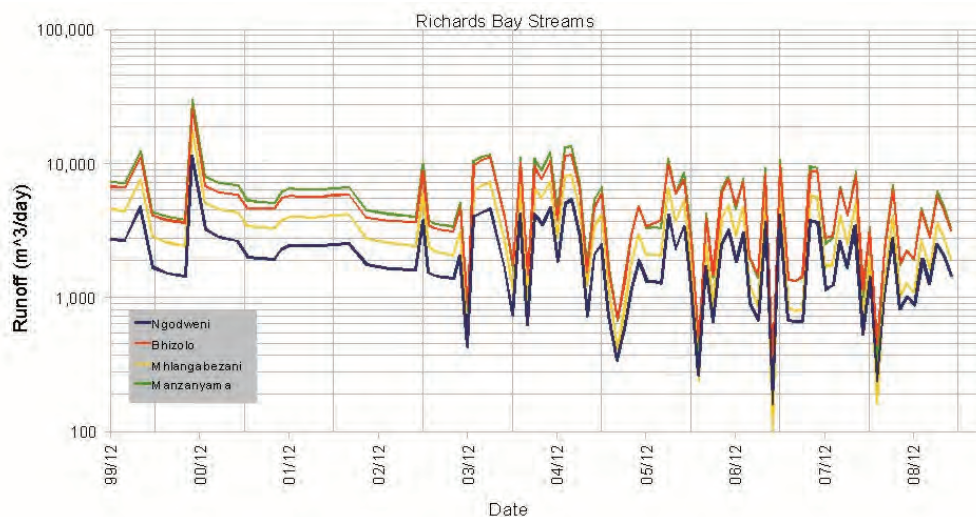


Figure 6.7: Simulated flow rates for four streams in Richards Bay using the calibrated groundwater model

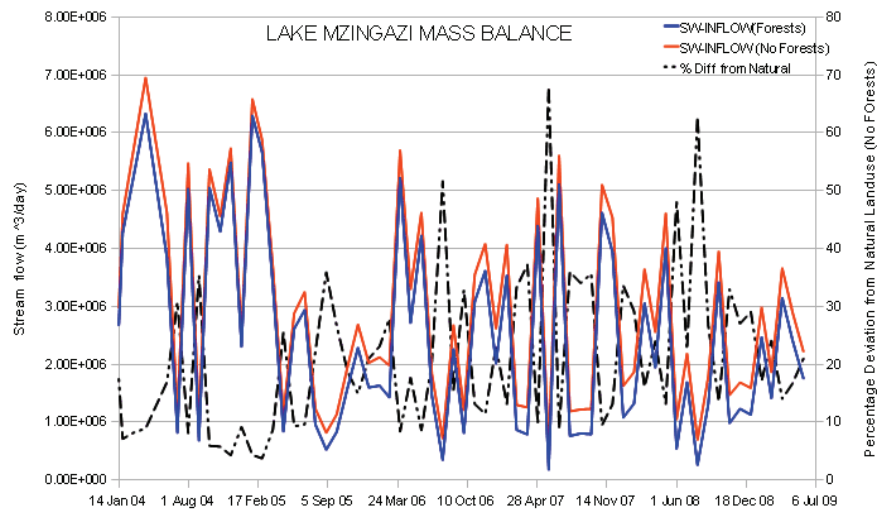


Figure 6.8: The simulated impact of afforestation on the streamflow into Lake Mzingazi using the calibrated groundwater model with the STR1 stream package

CASE STUDY 7: THE IMPORTANCE OF GROUNDWATER IN SUSTAINING THE ECOLOGICAL FUNCTION OF THE LAKE St. LUCIA ESTUARY/LAKE

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INTRODUCTION

Historically, South Africa's emphasis on water resource management had been on the maximization of exploitable reserves largely for the benefit of the agricultural, industrial and mining sectors. However, the global shift toward the sustainable utilization of limited resources has influenced the South African policy on water resources management. Following the initiatives proposed at the Rio Conference (Agenda 21) and subsequent international meetings (Earth Summit in 2002), the promulgation of the National Water Act (1998) has formalized South Africa's changing attitude to water resources management. This act, along with other important principles, has established water rights for basic human needs (currently set at 25ℓ/person·d) and the specification of the quantities and qualities of water that are required to protect aquatic ecosystems in order to secure ecologically sustainable utilization of the various water resources. However, to implement this legislation it is necessary to establish many of the functional relationships between the aquatic ecosystems and the quantity and quality of the water resources.

The implementation of the governing legislation (NWA) is hindered by the lack of methods and procedures for specifying the level of protection that must be afforded the water resource to sustain its aquatic function for future generations. Pioneering work has been done to establish the instream flow requirements for rivers (references) and estuaries (references) but little has been done for lakes and wetlands. Along the north-eastern shores of KwaZulu/Natal and Mozambique (the Maputaland coastal plain) there are numerous lakes and wetlands that are under threat from over development of the local water resources. Recent studies (Mhlathuze RDM) have indicated that the coastal lakes of Richards Bay on the southern tip of the Maputaland Coastal plain are stressed and require protection from over-exploitation. A similar study for Lake St Lucia (Van Niekerk, 2004) identified the precarious nature of the lake system (Fig. 7.1) and its dependence on the interaction between the terrestrial, marine and estuary systems. This case study looks at the role of groundwater seepage into Lake St. Lucia and its function in moderating the effects of salinity fluctuations that play an important role in the ecological resilience of the system.

Importance of the St. Lucia system

The St. Lucia system has been internationally recognized as a wetland of international importance by its designation, first as a RAMSAR site and then as a World Heritage Site. Recent studies by Været *et al.* (2009) indicate that anthropogenic impacts on the system are of more concern than climate change so it is important to understand the functioning of this system and the role that the different resources play in sustaining its important features.

Physical Features

Lake St. Lucia is the largest estuarine system in Africa, covering an area of ~36 000 ha. It is located in the subtropical zone on the east coast of South Africa. The lake is fed directly by incident precipitation, runoff from five major rivers with a combined catchment area of 6 085 km², and through groundwater seepage along the 187 km of shoreline (Fig. 7.1). It is connected to the Indian Ocean by a long, narrow (~100m wide) estuary that was once linked to the Umfolozi River that discharges into the ocean just to the south of the St. Lucia estuary (Fig. 7.3).

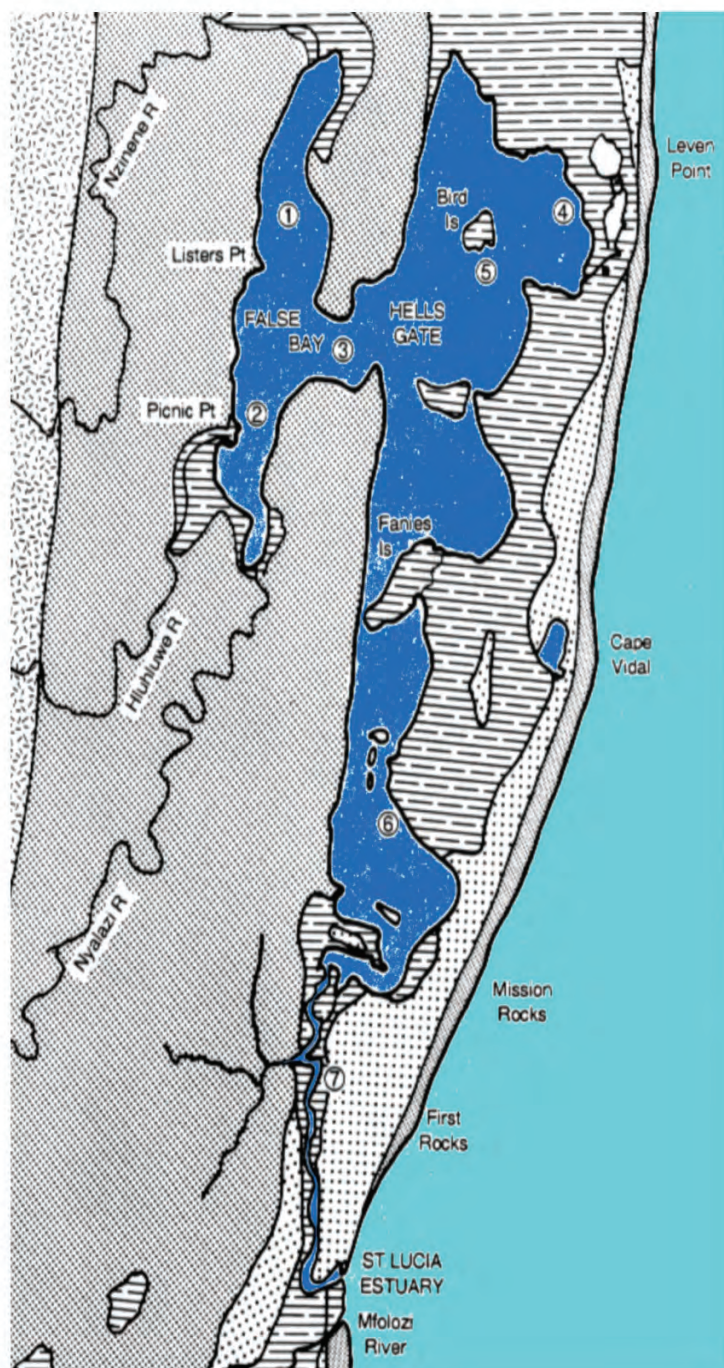


Figure 7.1: Location of the St Lucia system

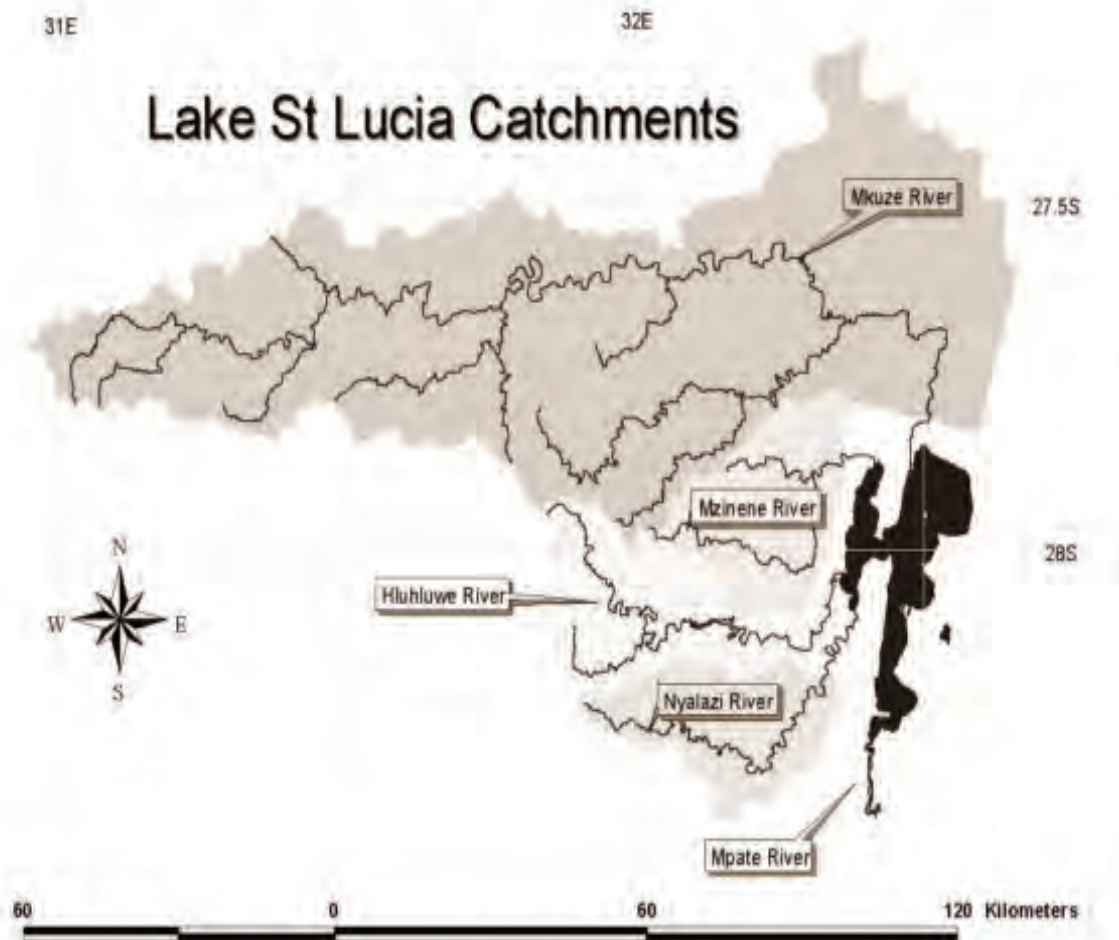


Figure 7.2: The main catchment areas for river runoff into Lake St. Lucia

The storage of water in the lake (water levels) is controlled by the balance between the freshwater inflow from the river runoff, groundwater seepage and direct rainfall, the inflow/outflow of sea water through the estuary and the direct losses through evapotranspiration. The annual average rainfall, runoff and groundwater seepage into the lake are given in Table 1 together with the estimated evapotranspirational loss. The water balance indicates a mean freshwater outflow to the sea of approximately $256 \times 10^6 \text{ m}^3$ per year during average climatic conditions. This outward flow of freshwater ceases when the combined rainfall-runoff contribution reaches about $465 \times 10^6 \text{ m}^3/\text{year}$. The small contribution from the lake shoreline groundwater seepage under average conditions has focused attention on the river systems.

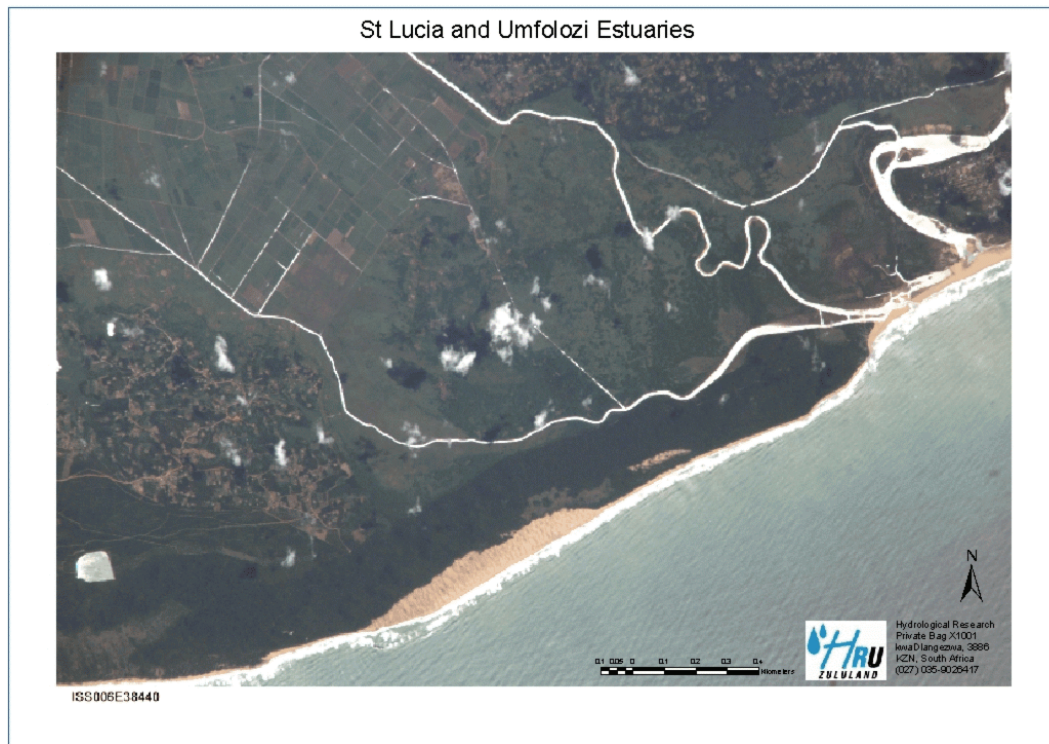


Figure 7.3: Satellite photograph showing the Umfolozi system of drainage canals, canalised Umfolozi River (northern branch) and the Msunduzi River to the south

Table 7.1: Water balance components of Lake St Lucia (*10⁶ m³/year)

Contributions to lake water balance	Ave Actual	Ave Reference
River runoff (Cornelius, 1992)	362	232
Ground water seepage along the lake shoreline (Kelbe et al., 1995)	53	53
Rainfall (1 000 mm/d)	360	232
Evaporation (4 mm/d)	518	518
Inflow/Outflow to the sea	256	0

Geology, Geomorphology and the Lake Development

The Lake St. Lucia system lies on the relatively flat Maputaland coastal plain that is dominated by several palaeodune cordons with interspersed wetlands and coastal lakes (Wright, 2002). These lakes were created during erosional (sea-level recession) periods and subsequently closed during periods of transgression. The sea level fluctuations modified the coastline and created the recent Holocene dune barrier that separates the lake from the Indian Ocean. This barrier dune formation, one of the highest in the world reaching elevation of over 120 m, caused constriction of the estuaries draining the five main rivers in the north that led to the closure of the lake/estuary in the northern parts of present system during the early Holocene (Wright, 2002). Lake St. Lucia currently drains through a long narrow estuary to the south.

The sediments from the five main rivers over geological times have reduced the current lake level to an average depth of less than 1 m. Currently, the Mkuze swamp and the very low gradients in the other river mouths insures a low level of sediment supply except in periods of extreme floods. In the early 1980s a canal was built to link the Umfolozi to the St Lucia estuary (see Fig. 7.3) but the inlet weir was covered by sediments during a cyclonic event in 1984 and the system was never restored for fear of Umfolozi sediments entering the lake system. The management of the lake has been to maintain separate mouths for the lake and Umfolozi River with a small sand bar that allows occasional overtopping when there is sufficient discharge in the Umfolozi system.

Lake St. Lucia was formed during the Holocene on the extensive coastal plain from marine and aeolian deposits on an inclined basement rock of the Karoo and Lebombo Formations. A transect (AA') through the lake from east to west is shown in Figs. 7.4 and 7.5. The underlying formations slope at an angle of about 3° . These are covered by the mudstone/siltstone deposited in the Cretaceous. This mudstone forms the basement of the primary aquifers along the coastal plain. It is exposed along the western shores of False Bay (Lake St. Lucia) and east of the Lubombo Mountains (Fig. 7.3). The Cretaceous reaches depths of about 30 m below mean sea level (b.m.s.l.) along the ocean shoreline.

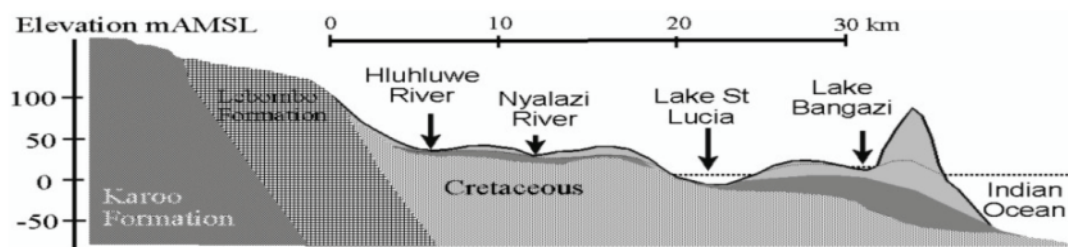
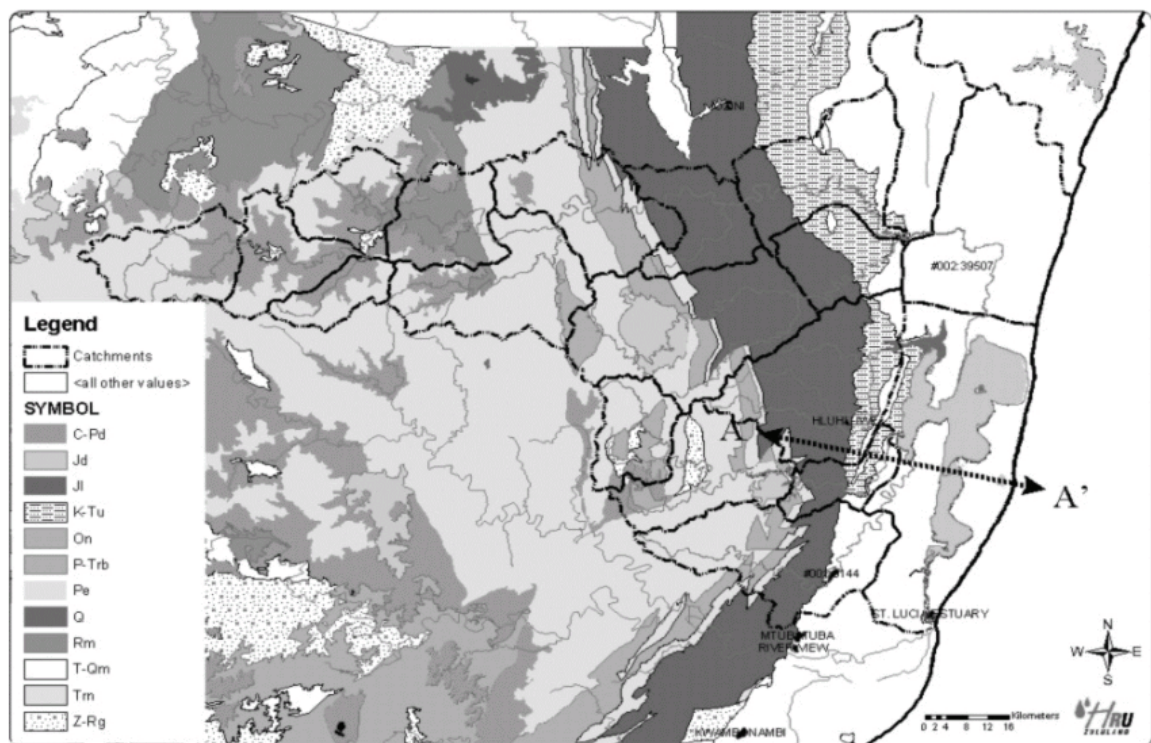


Figure 7.4: Transect AA' in Fig. 7. 5 from east to west through Lake St. Lucia

The shallow primary aquifer (illustrated in Fig. 7. 4) has a maximum depth of about 30m under the coastal dune belt but has a maximum elevation of less than 20 m a.m.s.l. This aquifer is formed by the recent marine sediments (Port Durnford Formation) overlain by Aeolian deposits. These sandy deposits generally have high permeabilities ($>1\text{m/d}$) and porosities (>0.2) with occasional interspersed layers of peat and clay.



. Figure 7.5: Geological features of the St Lucia catchment area

Rainfall Analysis

The climate of St. Lucia is subtropical with higher rainfall and temperatures during the summer months. February (summer) is generally the wettest month with occasional cyclonic activity and July the driest. In general, the winter period has about one quarter to one third of the summer rainfall. There is a strong rainfall gradient from the coast to the interior. The coast general has about 1 200 mm a year which drops to less than 900 mm within 50 km of the coast (Schulze *et al.*, 1997). There is considerable annual variability in the rainfall (Fig. 7.6) with a maximum between 1923 and 1997 of over 1 600 mm which has fallen to as low as 500 mm in the same period.

Freshwater Resources

The freshwater inflow to Lake St. Lucia from the five main rivers is very variable. The upper reaches of the Mkuze (the largest catchment) lies in the drier, older fractured rock areas of the region (Fig. 7.5) that drains into a large expansive swamp on the coastal plain before entering the lake. No monitoring exists between the swamp and lake so little is known about the amount of flow passing through the swamp. Monitoring exists on all the major rivers upstream of the swamp and flood plain but reservations have been made about the suitability, reliability and consistency of the measurements. Estimates of flow in these rivers have been made by several researchers (Stassen, 2004) that show considerable differences.

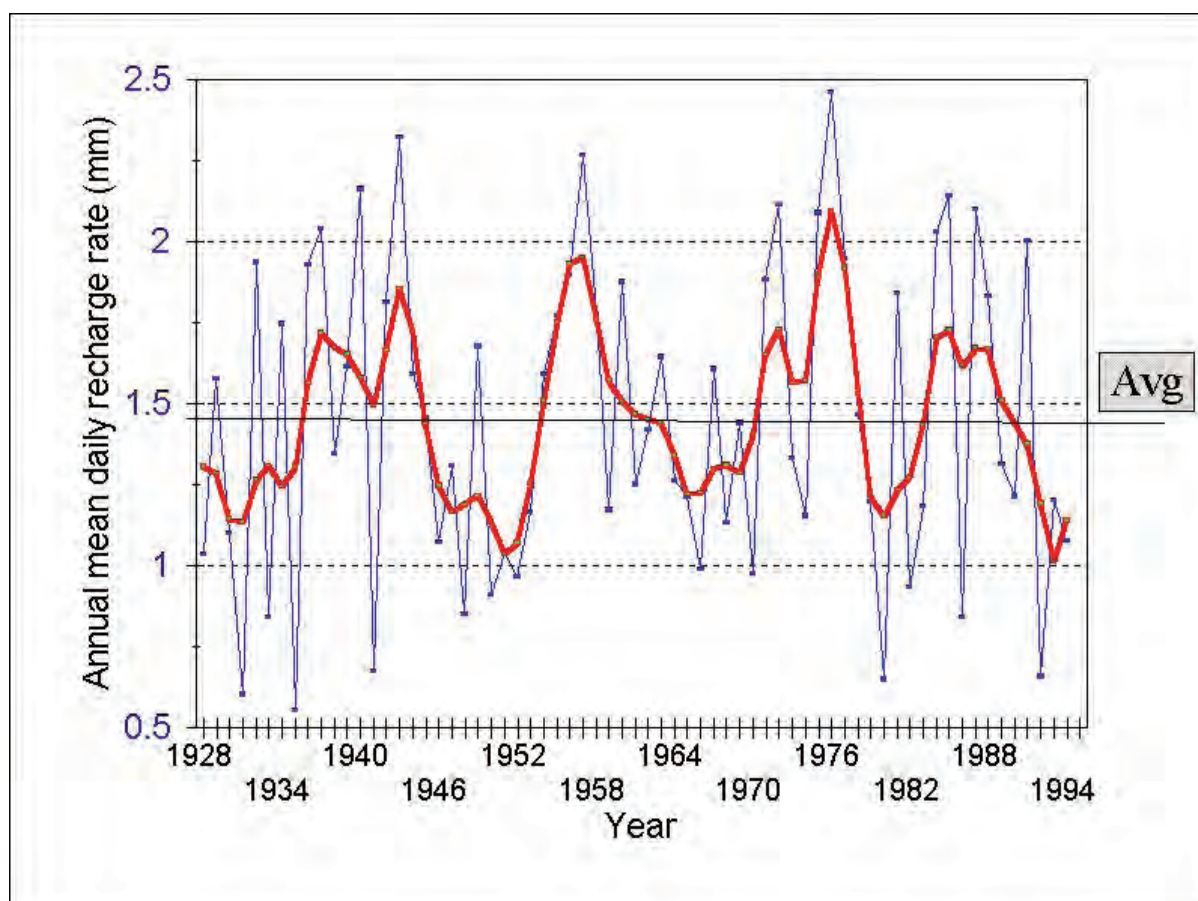


Figure 7.6: Annual variation in daily recharge rate (mm/d). Red line is the 5-term binomial smoothed series (after Kelbe et al., 1995).

The inflow of all five major rivers varied considerably over the six-year simulation period. This inflow has been significantly reduced by abstractions of various forms (irrigation, afforestation, and other forms of land use). The mean combined inflow of freshwater into the lake under present and natural (reference) conditions is given in Table 7.1. There are currently no restrictions on abstraction under drought conditions for environmental needs and the river inflow to the lake can be completely depleted. During 2004 and 2005 it was estimated that there was no freshwater inflow from the northern rivers to Lake St. Lucia. This has had a devastating impact on the water level (volume) of the lake that is described in later sections.

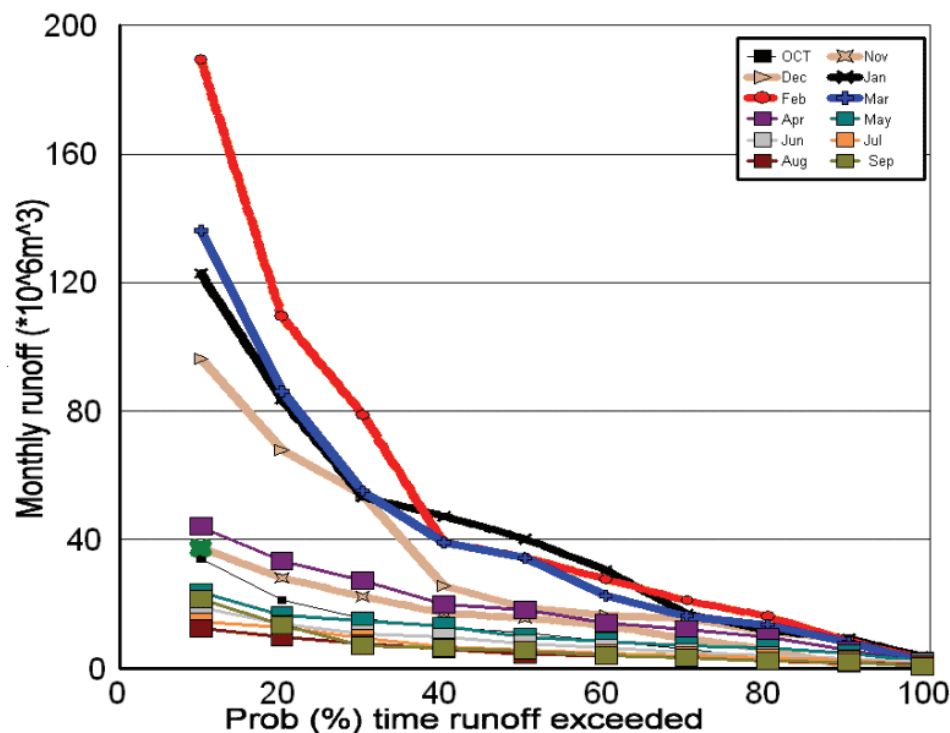


Figure 7.7: The probability (%) of monthly runoff into St. Lucia (adapted from Van Niekerk, 2004)

The Umfolozi and Msunduzi River system that drains into the Indian Ocean through the extensive flood plain to the south of the Lake St Lucia Estuary (Fig.7. 3) was linked to the lake over 50 years ago before the swamp was drained and sugar cane farming was established. The Umfolozi Rive has subsequently been canalised and is presently elevated above the flood plain (sugar farms) by levees along the lower section of the system. An intricate diversion system for high flows has been engineered above the flood plain to divert flood waters through the more southerly Msunduzi channel (Fig. 7.3). These engineering works have broken the link between the Umfolozi and Lake St Lucia Estuary and only in exception circumstances does river water enter the estuary. This happened when both river and estuary mouths were closed and a small flood wave was able to flow over the sand bar at the mouths before the Umfolozi system breached in January 2005. A canal was built in the 1980s to establish an artificial link but the intake works on the Umfolozi River were blocked by sediment during a cyclonic event before the system could be commissioned in 1984.

The potential for sediment intrusion into the very shallow estuary and lake induced management to abandon the link canal. The loss of swamp lands on the Umfolozi/Msunduzi flood plain that filtered the sediment is limiting the options for linking the system with the lake to alleviate the stress under extreme droughts. There is a call for re-establishing the linkage (Van Niekerk, 2004) but it would require renovation of the swamp lands to eliminate sediments entering the lake system.

The river inflow under mean annual conditions contributes about half of the total freshwater inflow to the lake system. The other contributions come from direct precipitation and groundwater seepage along the shoreline of the lake system. While no estimate of the groundwater contribution in the river runoff (i.e. baseflow) has been made it is probable that these rivers have at least 50% of their contributions from base flow that is assumed to be groundwater feed. This would vary from negligible amounts during wet periods to the major proportion during low flows. The baseflow contribution to the lake water balance is not considered any further in the remaining discussions on the role of groundwater in this case study.

Groundwater Components

The groundwater contribution to the freshwater inflow from the four major rivers feeding into the northern lakes (Mkuze, Nelazi, Msinene and Hluhluwe rivers) has been lumped with the estimated total river runoff in Table 7.1 and Fig. 7.7. The groundwater seepage along the lake shoreline has been estimated by Kelbe, Rawlins and Nomquphu (1995) to be a small proportion of the total freshwater inflow into the lake under mean annual conditions for the eastern and western shores region of the lake. No estimates for the northern shoreline (False Bay and Mkuze swamp sections) have been identified. However, it is assumed that the Mkuze swamps would provide similar seepage rates to the eastern shores region while the northern parts of False Bay would be similar to the western shores seepage condition.

Groundwater elevations for a network of piezometers on the eastern shores (Fig. 7.8) show a range of variability that seldom exceeds 2 m. The variability in the water table elevation over 25 years during which time there were three severe droughts and two extreme flood events is shown in Fig. 7.9 for the central section of the monitoring network where the groundwater mound reaches local maxima of about 18 m a.m.s.l.

Kelbe *et al.* (1995) simulated the groundwater dynamics for the western and eastern shores of Lake St Lucia and derived estimates of the groundwater seepage rates along the lake coast line. The groundwater seepage along the shoreline from the western region into the lake under natural (reference) conditions contributes an average of 22,000 m³/d to the southern half of False Bay, approximately 17 000 m³/d to the south lakes and almost 5 000 m³/d to the estuary under reference conditions (Fig. 7.9). The groundwater contribution from the Mpuze River is included in the river flow estimates but it is highly likely that this runoff is dominated by the groundwater contribution. The western shores aquifer is characterised by a very shallow layer of unconsolidated cover-sands overlying the deep but very low permeable layer of siltstone/mudstone deposits of Cretaceous age. These sections of the shoreline are almost completely covered by extensive commercial forests of pine and eucalyptus plantations that have had a significant impact on the



Figure 7.8: Groundwater monitoring sites

groundwater discharge to the local drainage systems into the lake (Fig. 7.9). The seepage rates into the False Bay, south lakes and estuary sections are reduced by more than 5 000 m³/d.

The simulated groundwater seepage from the eastern shores region (Fig. 7.10 and Fig. 7.11) shows that the outflow to the Indian Ocean is approximately twice the outflow to the lake. There is also a significant reduction due to afforestation in the seepage into the Indian Ocean along the southern dunes (5 to 20 km from the estuary mouth) and a 50% reduction in seepage to the lake along the central plains region (23 to 35 km from the estuary mouth). As of March, 2005, all the pine plantations on the eastern shores will have been clear-felled and will be left to return to natural conditions under controlled management.

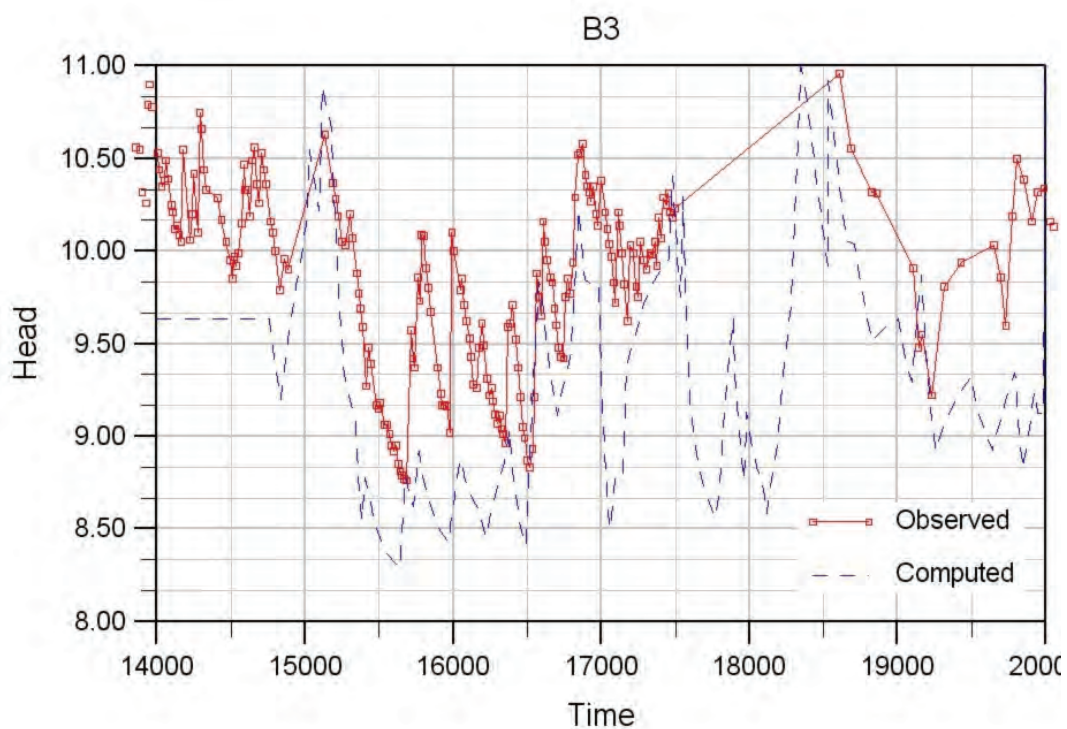


Figure 7.9: The measured (red squares) and simulated (dashed line) water table fluctuation in the central region of the eastern shores of St. Lucia

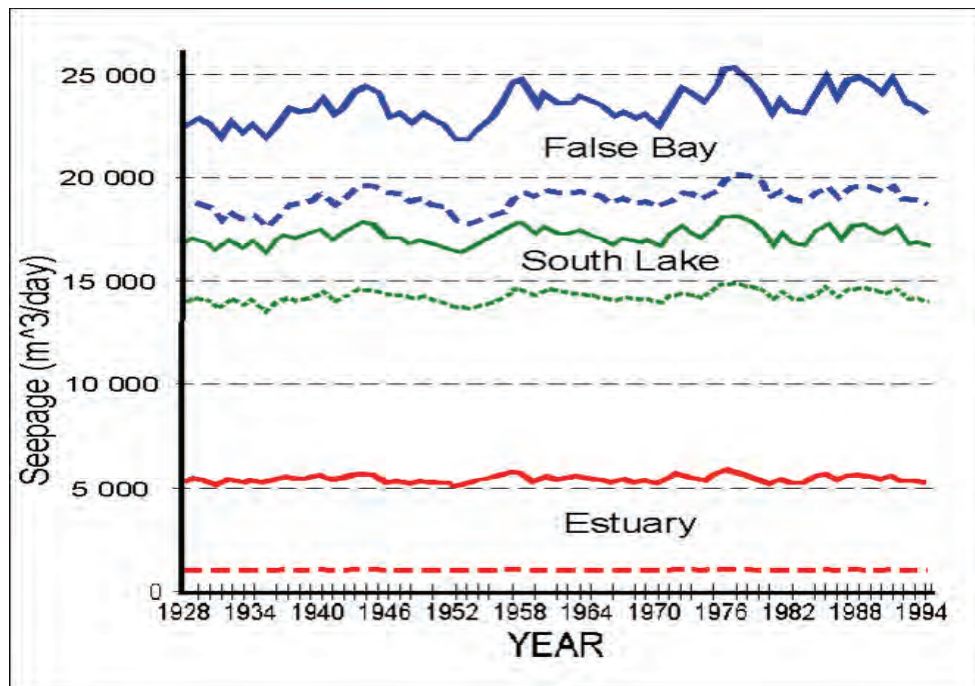


Figure 7.10: The simulated groundwater fluxes into the different lake compartments from the western shores region under present and reference conditions

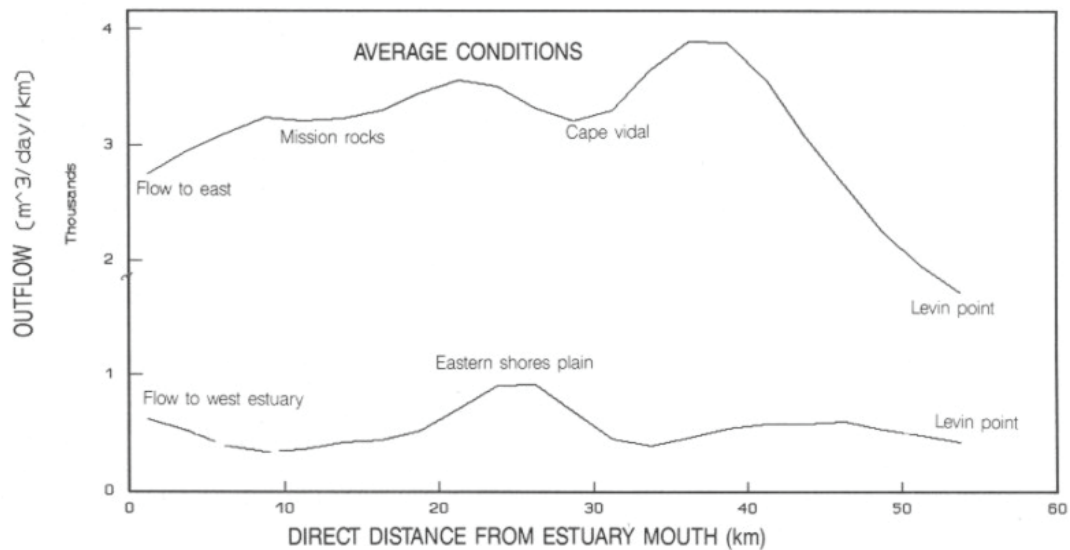


Figure 7.11: The simulated groundwater fluxes into the lake and to the sea along the length of the shoreline

Measured Groundwater Seepage along Shorelines

The groundwater seepage from the eastern shores region was confirmed during the severe drought of 2003-2005 when the lake level dropped and exposed the lake shoreline. Aerial photographs of the exposed shoreline showed the groundwater flowing down the beaches into the lake in many areas. Figure 7.12 shows the zones of exposed groundwater with electrical conductivity values of $<5\text{mS/cm}$ seeping down the exposed beaches into the lake where the EC was $>40\text{mS/cm}$. A small concentrated zone (spring) of discharge from the wetland just inland of the old shoreline was measured at $10\text{ m}^3/\text{h}$ on the 24 May 2003. The groundwater seepage down the beaches was not as strong on 7 June 2004 after $2\frac{1}{2}$ years of severe drought conditions.



Figure 7.12: Areal photograph of the groundwater seepage at the jetty on the eastern shores of St Lucia on 2003. Concentration contours of Electrical Conductivity are shown from measurements along transects.

Evaporation Losses and Water Balance

The imbalance between the freshwater inflow and outflow has a major influence on the marine connection to the lake. With a net outflow of freshwater from the lake for more than 50% of the time, the lake will function in the form of a freshwater ecosystem. However, when the rainfall and runoff provide insufficient inflow, the net importation of sea water inflow into the lake will raise the salt load of the lake system.

Groundwater Seepage Contribution to Lake Water Balance

During the periods when rainfall and runoff greatly exceed the groundwater seepage rate, the groundwater contributions to the overall water balance of the lake are negligible. However, as the rainfall falls for sustained dry periods, with concomitant reduction in river runoff, the groundwater seepage becomes an increasing larger proportion of the overall water balance. When runoff ceased in early 2004, the groundwater became the only source of freshwater into the lake with the possible exception of small intermittent showers. During this period of low runoff, the groundwater seepage played an important role in reducing the rate of increase in salinity levels for a considerable period in 2003. Figure 7.13 illustrates the relative importance of the runoff, direct precipitation and groundwater to the water balance of the whole lake.

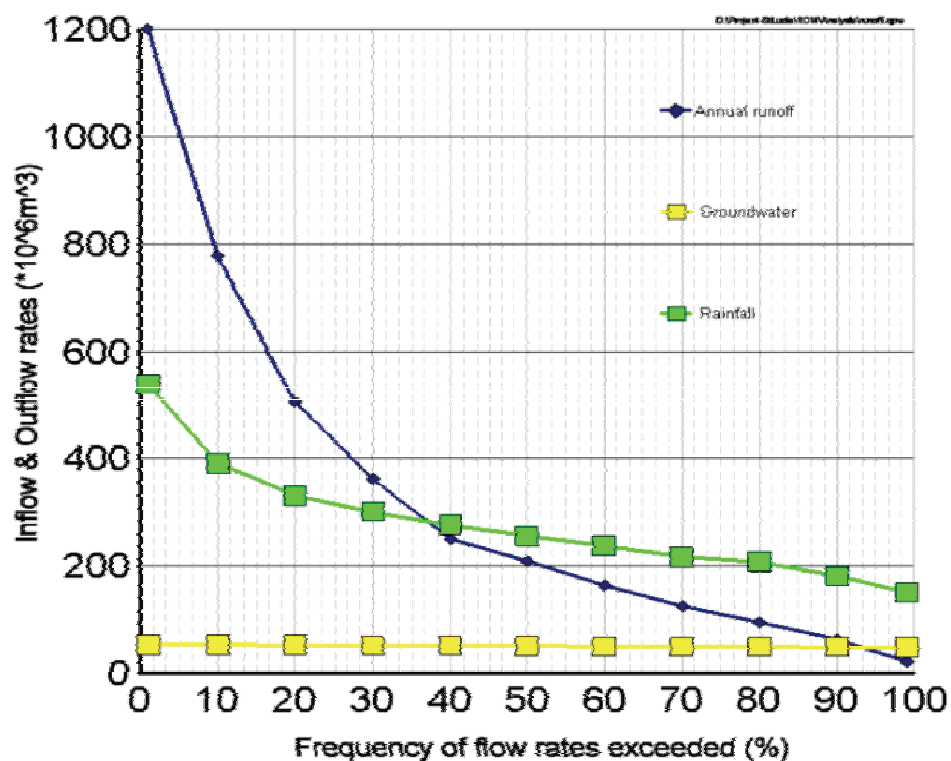


Figure 7.13: The frequency distribution of river runoff, direct precipitation and groundwater seepage

The estuary mouth closed in the middle of 2002 when the lake salinities were generally below those of sea water. With the mouth closed from 2002 to 2004, the lake levels dropped drastically to the stage where four isolated compartments remained by the end of January 2004 (Fig. 7.14). In the northern lake segment, where groundwater contribution is estimated to be the least, the salinities rose exponentially high in early January to reach peaks of nearly 300 ppt. In contrast, the southern, smaller lake compartment showed a much lower rise in salinities in the latter part of 2003 when the northern lake salinities were already twice that of sea water. Water balance models of the shrinking lake area indicate

that the groundwater played the dominant role in regulating the salinity levels during the extremely dry conditions when the link to the sea was lost.

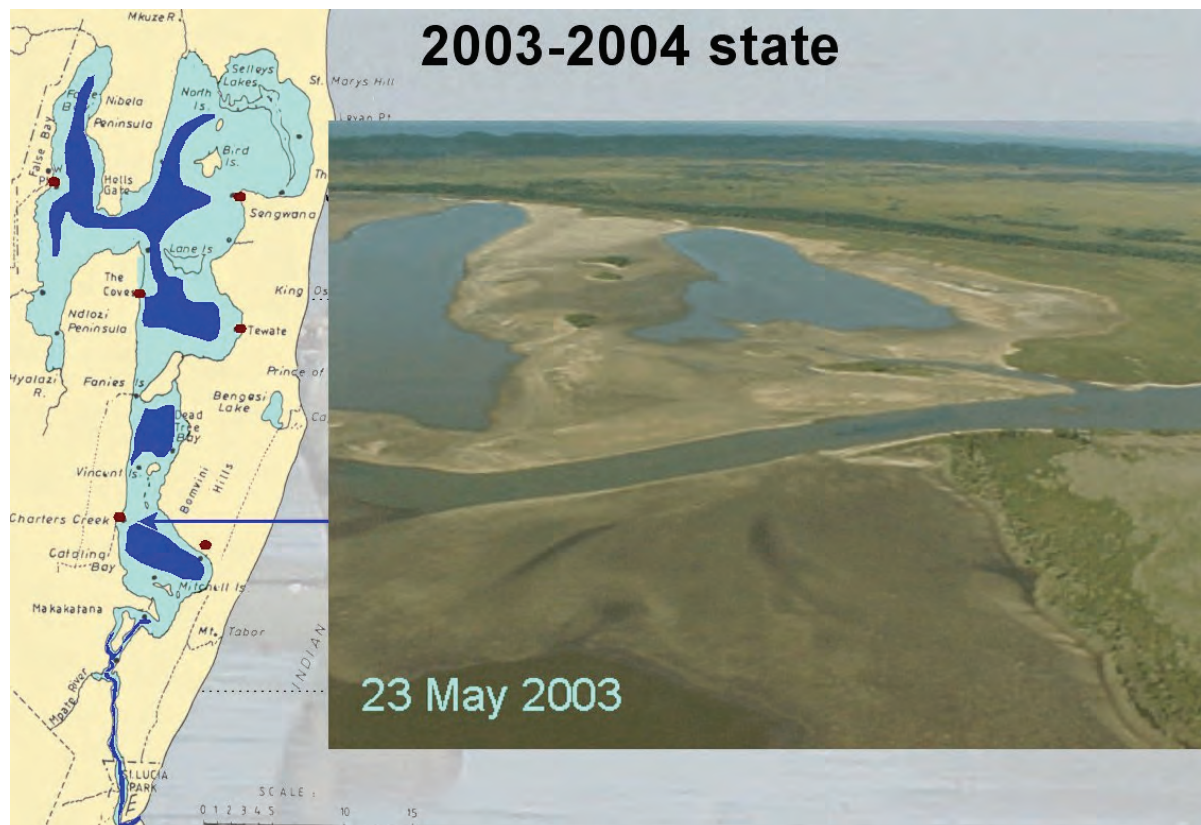


Figure 7.14: The outline of Lake St Lucia under normal conditions (light blue) and the remaining water bodies during the severe drought of 2004 after the mouth had remained closed for nearly 3 years. Inset photograph shows the remaining water in 2003

The importance of the diffuse seepage zones along the lake shoreline and the small streams and concentrated groundwater discharge into the lake have been described by Taylor (1995) and Taylor *et al.* (2006). The ecology of the system is influenced by the salinity, temperature, depth of the water, as well as the time that these conditions persist in the lake. These conditions are generally inter-correlated as reduced depth (volume) is associated with dry conditions that lead to higher salinities and temperatures. Consequently, only salinities are considered in this study. Since 1960, the lake has fluctuated on four occasions from brackish-freshwater state (with low salinities) to hypersaline states for extended periods of several years. The ecological response to these changing physical conditions is time dependant and resource limiting. The change from an ecological state with freshwater species to a state with marine species, or vice versa, requires sufficient time or initial populations of each species to reach a stage of abundance. Studies by Taylor (1995) have shown species recovery rate from a hypersaline to freshwater state is higher than traditional growth curves suggest. It is believed that this is due to a higher survival rate of freshwater species during hypersaline conditions than previously thought. Taylor (1995) has suggested the existence of micro habitat (refugia) sites where these species are able to survive. These sites are associated with groundwater seepage that provides sufficient freshwater for their survival during extended periods of hypersalinity in the lake. These refugia sites are associated with both diffuse seepage along the shoreline and through concentrated flows in rivulets and streams.

Water Quality (salinity)

A major factor contributing to the ecological state of the lake system is related to water quality, in particular the lake salinity levels. In a freshwater state, most freshwater species multiply and invade the lake system until some limiting factor restricts their further production. Similarly, when the lake is in a saline state, marine and saltwater tolerant species propagate until the system limits their growth. In periods of above average rainfall, the lake experiences a net inflow of freshwater. As the climate changes to a drier state, the inflow of sea water, coupled to high evaporation rates, creates an initial brackish state that may eventually lead to saline conditions in the lake. In exceptional dry periods, the lake may experience hyper-saline conditions when the salinity levels greatly exceed the salt concentration of sea water. These states may be exasperated by the estuary mouth closure which further limits water inflow into the lake to compensate for evaporation leading and may lead to even higher salinities and reduced lake area/levels. Such a situation occurred between 2002 and 2005 when the lake area diminished to the areas shown in Fig. 7.14 and the salinities rose to over 90 ppt for Catalina Bay (Fig. 7.15) while the Narrows' salinities remained close to 30 ppt (Fig. 7.15). The salinity levels dropped drastically during this period when the Umfolozi River overflowed into the estuary.

Groundwater Impact

In an attempt to understand the role of the groundwater in the lake during these extreme periods a mass balance model was developed for the lower compartments of the lake during these periods when they became isolated from the northern sections of St Lucia. Consequently there was no river inflow from the north and only limited inflow from the Mpate River, some overtopping events from the Umfolozi and direct rainfall. The conceptual model is shown in Fig. 7.16. The model was calibrated against the water level measurements and the salinity fluctuations shown in Fig. 7.15.

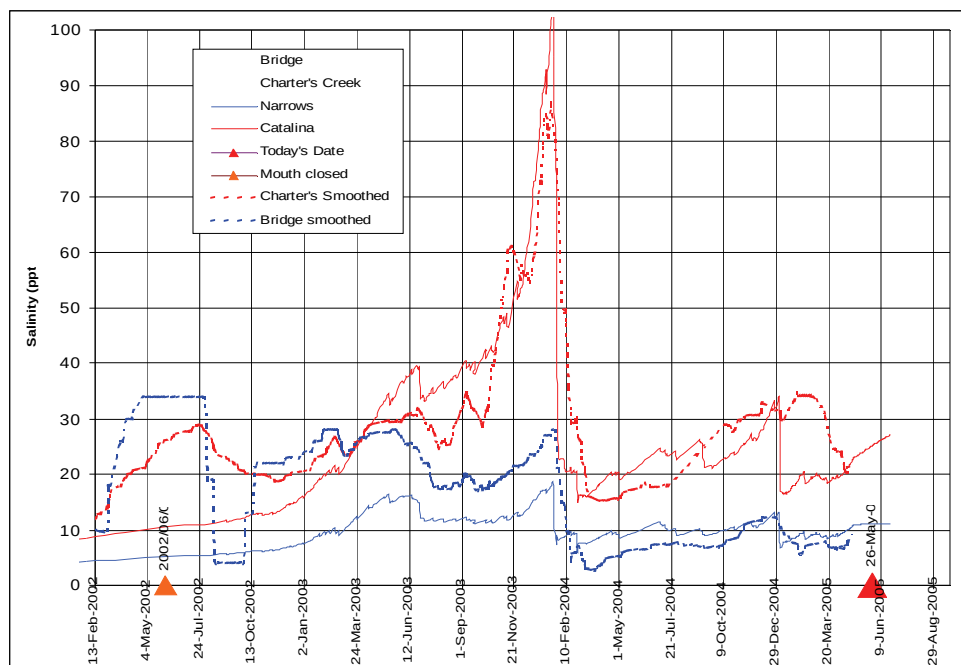


Figure 7.15: The measured and simulated salinity levels for the estuary and Catalina Bay during the drought of 2004

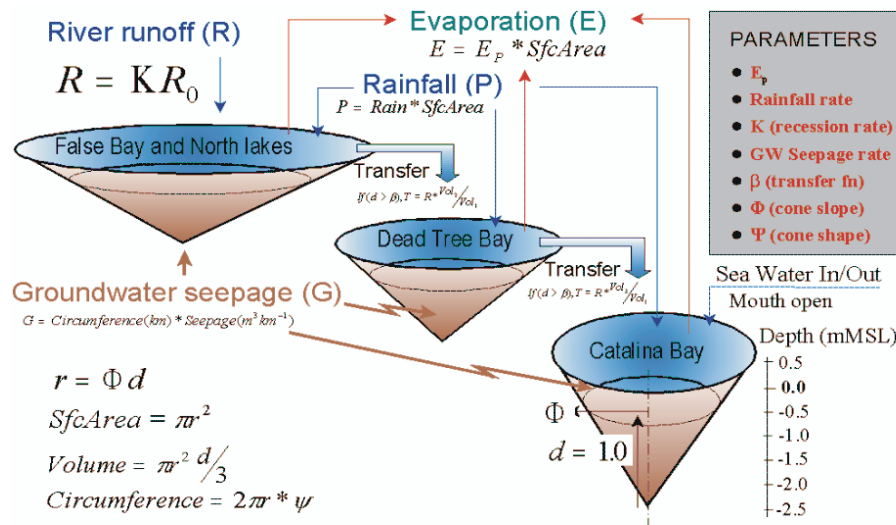


Figure 7.16: The conceptual water balance model for assessing the role of the main components that sustain the lake levels and salinity profiles during the drought

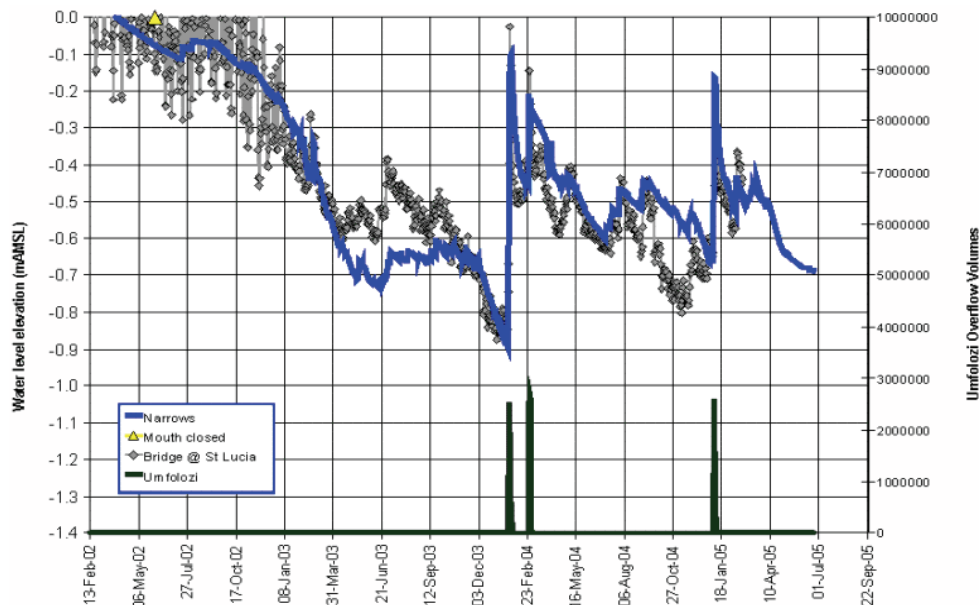


Figure 7.17: The measured and simulated water level elevation for the mass balance model of the Narrows and Catalina Bay

Summary and Conclusions

The freshwater inflow to Lake St. Lucia from the main rivers is highly variable, ranging from extreme floods to prolonged droughts. Subsurface seepage along the shoreline is derived from large groundwater mounds located mainly on the eastern shores region of the lake. This groundwater seepage is very consistent, even in extreme droughts when it constitutes the only freshwater inflow to the lake other than direct rainfall. This consistent flow, both diffuse and concentrated (streams), provides freshwater for the formation of refugia sites that support freshwater species during harsh environmental conditions.

The survival of these species provides a significant population for a rapid re-growth under more favourable conditions, when the lake returns to a freshwater state. Consequently, the groundwater seepage is considered an important component in the ecological resilience of Lake St. Lucia to large changes in the lake salinity that occur with the long-term climatic changes.

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