UNDERSTANDING AND MODELLING SURFACE WATER-GROUNDWATER INTERACTIONS

Report to the

WATER RESEARCH COMMISSION

by

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EXECUTIVE SUMMARY

The main objective of the total project was to contribute to the incorporation of uncertainty assessments in practical water resource decision making in South Africa. The companion report addresses more general issues of uncertainty and hydrological modelling, while this report concentrates on the uncertainties in both understanding and modelling the interactions between surface water and groundwater. Since groundwater routines were added into the widely used Pitman model in the early 2000s by both Prof Hughes and Mr Karim Sami, the approaches have come under a great deal of criticism mainly from the geohydrological community of specialists within South Africa. Arguably, a great deal of this criticism is based on misunderstandings of the intention of adding groundwater routines into an existing surface water model. It was stated quite clearly at the time that this approach was not seen as a replacement for existing detailed numerical approaches to groundwater modelling. The intention was to create a scientific and practical tool that could be used to simulate the complete hydrological cycle at the catchment scale so that integrated water resources decision making could be better supported. Further criticism was levelled at the model for being too simplistic and 'conceptual'. Once again there was a reason for doing this and that was related to the need for a degree of consistency within an existing model – i.e. the added groundwater routines should be aligned with the original design concepts of the model. Interestingly, the major criticism from the surface water modelling community was that the added routines were too complex! All of the criticisms of the model structure and its application (indeed, the application of any model of surface-groundwater interactions) have to be viewed in the light of our current understanding of the real world processes involved, as well as the available data to support this understanding and to establish any model. The first part of this report therefore reviews this understanding in the context of different hydrogeological environments and with respect to the different processes. The conclusion is that while we may understand the basic concepts very well, developing a site specific understanding is heavily constrained by an almost complete lack of direct observations. This is one of the critical issues that separates purely surface water studies (where we typically have some observations and can at least see the processes happening) from studies that involve the movement of water in highly heterogeneous sub-surface environments.

Chapters 2 to 4 highlight many of the uncertainties that are involved in understanding and quantifying the interactions between surface water and groundwater and makes it abundantly clear that, at the catchment scale, this is a worldwide phenomenon. All the processes, from recharge, through unsaturated zone water movements to interactions between groundwater storage and river channels, are complex and difficult to quantify. This makes constructing and simulating a water

balance extremely difficult, regardless of the type of model being used. Some model types have traditionally focussed on the detail, but suffer from the problem of a lack of data to define that detail. Others, such as the Pitman model, concentrate on conceptual representation of the main water balance components and their linkages at the catchment scale (which can cover a wide range of scales in practical water resources modelling). The emphasis in models such as the Pitman model is quantifying the water balance components correctly enough such that anthropogenic impacts on some components (e.g. groundwater abstractions) will be more-or-less correctly reflected in others (e.g. reductions in low flows in rivers).

Chapter 5 presents a brief description of the model and summarises the model applications and evaluations that have already been published after international peer review. This is an important point that has been neglected in several other local South African reports and presentations. The model concepts and their application have been subject to substantial amounts of scrutiny by the international community of hydrological scientists. Chapter 6 presents several additional examples of the application of the model in a range of different environmental settings. These were selected partly because they represent examples of different aspects of uncertainty (in understanding or quantification) and partly because they have been subject to previous studies by other groups and therefore comparisons can be made between our conclusions and previous ones. For example, previous studies on the Breede River concluded that the Pitman model (albeit the Sami version of the model) could not be applied to represent surface water-groundwater interactions in this complex environment. In contrast, this report concludes that the model can be used very successfully. The difference is that this report also accepts that there are substantial uncertainties in the way in which the model should be set up to represent the processes in the Breede River catchment. However, these are related to our lack of understanding and lack of real data, not with the structure of the model. If the understanding uncertainties can be reduced through the collection of additional information, the Pitman model uncertainties will also be reduced. This is the situation in any modelling study and is certainly not unique to the Pitman model. In almost all of the examples used, the model has been able to simulate the various water balance components. The fact that it can achieve this in different ways is a testament to good model design and reflects the real world situation. For the Crocodile River example we simply do not know how much recharge occurs and therefore we do not know how much groundwater is lost through riparian evapotranspiration. This example demonstrates that the Pitman model can simulate high recharge and high losses, or lower recharge and low losses and still get the right results based on the observed stream flow response. Resolving which is correct can only be achieved with more data, not with a different model.

The Gamagara example illustrates that when we have good groundwater data, we can understand the processes in a better way. It is unfortunate that we do not have any surface water data to support the establishment of the model. The way in which the model has simulated transmission losses under no de-watering of the aquifer and under de-watering conditions is supported by the observed borehole level response data. It would have been very useful to have access to stream flow data to validate the effects on losses from the channel. This is a very typical (and often frustrating) modelling issue – there are some data available to support and validate certain aspects of a simulation result, but not enough to validate the whole model. Once again this is not unique to the Pitman model and applies to all models.

From a modelling perspective, the overall conclusion of the study is that the Pitman model with the revised groundwater routines is a useful tool for exploring different hypotheses about some of the processes involved in SW/GW interactions at the catchment scale. However, in the absence of critical field information, the model is not always able to help resolve some of the uncertainties that exist. Unfortunately, such information is frequently not available and therefore these uncertainties remain regardless of the type of model that is used. Some of the uncertainties can be partially resolved using conceptual 'common-sense' plus limited field information to support such tentative conclusions. However, there is a need to clearly identify those uncertainties that remain unresolved and determine which ones are critical from a water management perspective and design appropriate data collection and research programmes. One of the advantages of an integrated model, such as the Pitman model, is that it is able to assist with the identification of these uncertainties.

It is hoped that this report will go some way towards counteracting the very negative attitude towards the revised Pitman model (either the Rhodes University version used here or the Sami version) that has prevailed within some hydrogeological communities in South Africa. Given that the whole motivation for developing the model was to improve the way in which surface water and groundwater simulations can be integrated, it is unfortunate that it seemed only to increase the divisions between surface and groundwater hydrologists. As mentioned many times in the past, there was never any intention to challenge the need for existing types of groundwater models. The purpose of these models is frequently different to that of the Pitman model and previous experiences suggested that existing groundwater models could not fulfil some of the needs that the modifications for the Pitman were made for. It is also interesting to note that, subsequent to the development of the modifications, other research projects that have been designed to investigate surface water-groundwater interactions have been initiated by some of the models fiercest critics. At least some of these have developed even simpler water balance routines that ignore some of the processes that have been included in the Pitman model. It is not possible to reference these studies as they have yet to be published. The authors of this report would contend that the processes that have been included in the Pitman model are the minimum that are necessary to deal with the variety of situations found in South Africa (and other parts of the sub-continent). This contention has been partly supported by the example studies presented in this report and within the PhD thesis of Dr Jane Tanner (2013). Further simplification is therefore not considered as a viable option.

Finally, while this report is completely focused on the Pitman model, it is not the intention to suggest that this is the only model that can be used to successfully simulate surface-groundwater interactions. It is clearly understood that there are other models that can achieve this objective. However, it was not the intention of this project to review a wide range of different models. The intention was to review our understanding of the processes involved in surface-groundwater interactions and to determine whether the Pitman model can be considered to be aligned with this understanding.

Table of Contents

List of Figures

List of Tables

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A very important part of this project was the participation of various individuals in the Reference Group meetings and a single workshop that we held to discuss some of the issues associated with the practical application of uncertainty analysis and decision making. The following table lists the names of the individuals and their organizations who attended the reference group meetings and therefore made significant contributions to the success of this project:

Name	Organisation
Alan Bailey	Stewart Scott
André Görgens	Aurecon
Bennie Haasbroek	Hydrosol
Herman Keuris	DWA, WRIW
Tendayi Makombe	DWA, NWRP
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Sabine Stuart-Hill	CWRR, UKZN
Isa Thompson	DWA
Peter van Niekerk	DWA
Geoff Pegram	UKZN/Pegram & Associates
Kornelius Riemann	Umvoto Africa
Pieter van Rooyen	WRP
John Wise	Umvoto Africa
Niel van Wyk	DWA

In addition, data from a number of consultancy projects and publications have been invaluable to this study. They include a report on the Gamagara River by GCS (Pty) Ltd (Bredenkamp and De Jager, 2011), the Berg Water availability and assessment study (DWAF, 2007a) undertaken by Umvoto Africa, and a conference paper by Koos Vivier (Vivier *et al*., 2007).

CAPACITY BUILDING

Apart from the students who were directly associated with the project, there are several other post-graduate students at Rhodes University who have directly benefited from the uncertainty and surface water-groundwater components of this project. At the start of 2009 the Institute for Water Research began a new programme (SSAWRN) supported by the Carnegie Foundation of New York. The Sub-Saharan Africa Water Resources Network (SSAWRN) is part of the Regional Initiative for Science Education (RISE) project managed by the Science Initiative Group (SIG) of Princeton University and is designed to build academic capacity in Africa. Two of these students worked directly on the project. Dr Jane Tanner (graduated in 2014) is the senior author on this report, while Mr Thabiso Mohobane is an author on the companion uncertainty report and has been focusing on the Caledon River basin. He was finalising his PhD study during 2014. Ms Madaka Tumbo, from the University of Dar es Salaam in Tanzania, is registered for a PhD at Rhodes University, is funded by Danish support (as part of a the CLIVET project) and has been using some of the developing uncertainty modelling techniques on the Great Ruaha River basin. She was also finalizing her PhD during the early part of 2014. Mr Sbongiseni Mazibuko is a RISE MSc student who has been investigating the value of MODIS data products for calibrating and validating the evapotranspiration component of hydrological models. Mr Gregory Pienaar started his MSc with the Institute for Water Research (and funded by the IWR) at the start of 2014 and he will be looking at water resources decision making with uncertain information.

1. INTRODUCTION

This report explores some of the principle issues and uncertainties associated with quantifying surface and groundwater interactions and the practical application of models, in particular the Pitman Model, in a data scarce region such as South Africa. While the hydrological cycle is well documented, the linkages between the interdependent components are less well understood. This is especially true of those regions that suffer problems of data scarcity and there remain urgent requirements for regional water resource assessments especially in arid and semi-arid regions where optimal management of all water resources is essential. The potential impact of the abstraction of groundwater on stream flow in rivers has prompted research toward improving the conceptual understanding and toward the development of tools to quantify the interactions between surface and groundwater. This integration is constrained by problems such as a lack of observed data (for surface and groundwater, as well as their integration), a lack of understanding of the processes of interaction and the fact that different methods (models) of assessment have been traditionally used for surface and groundwater. There remains a great deal of disagreement on the most appropriate methods to use to quantify surface and groundwater interactions. Part of this disagreement is founded on a lack of common understanding of the processes, part on disparate disciplinary traditions and part on disagreements about the purposes that quantification methods (models) are used for.

Hydrology (both surface and groundwater hydrology) is an uncertain science; it aims to represent highly variable and non-stationary processes which occur in catchment systems, many of which cannot be measured at the scales of interest (Beven, 2012). Model structural uncertainties are difficult to resolve due to the lack of relevant data. Incomplete and often flawed input data (such as rainfall and evaporation data) are used to force the models and generate quantitative information. Uncertain implementations (model structures and parameter sets), driven by uncertain input data will necessarily produce uncertain results (Beven, 2012). Model testing has limited power as it is difficult to differentiate between the uncertainties within different model structures, different sets of alternative parameter values and in the input data used to run the model. This has been termed equifinality by Beven (1993, 2012) which represents the possibility that many different parameter sets within a given model structure might give equally acceptable results when compared with observations. However, not all of them will be behavioural, in the sense that not all of them will produce the right results for the right reason (Kirchner, 2006). This issue is exacerbated when the processes of surface and groundwater interaction are poorly understood, in that we sometimes do not even know what the right reasons are. From a practical water resources management perspective, if we do not get the right model results for the right reasons, it is possible that we will make the wrong management decisions.

The approach adopted in this report is to define the critical surface and groundwater interaction processes which occur within 'typical' interaction environments in South Africa and assess whether an available model can represent them and to assess the uncertainties in our knowledge, our data and ultimately in our model results. The available knowledge about different components of surface groundwater interaction systems is assessed using both local (to southern Africa) and international literature. Part of this knowledge is associated with our conceptual understanding (qualitative knowledge) of the processes involved and part of it is associated with the quantitative data that are available to support the conceptual understanding and to parameterise appropriate models. The model that has been used within this study is the modified version of the widely used Pitman model (Pitman, 1973; Hughes, 2004; Hughes 2013). The authors acknowledge that the choice of this model will meet with some criticism from some sectors of the water resources community within South Africa. Part of the justification for the use of the model is that the authors are very familiar with it and very experienced in its use (unlike some other organisations and individuals who have compiled unpublished reports on the assessment of the model). Part of the justification is that the surface water components of the model have been used for many years in South Africa and elsewhere and applications of both the surface and groundwater components have been subject to international peer review and have been published (Hughes, 2013). However, the main justification is that the modified version (or the very similar counterpart version developed by Mr Karim Sami) of the Pitman model was specifically designed to address catchment-scale water resources management issues that are related to the interaction between surface and groundwater. It is also important to emphasise that this study is not about developing or testing models that are used for designing local borehole abstraction schemes. It has been emphasised on many occasions that the model is designed to complement more detailed finite-element or finite-difference type groundwater water models and not to compete with such models.

One of the main focus points of this report is the acknowledgement of different sources of uncertainty in the whole process of modelling both surface and groundwater resources. These uncertainties derive from our conceptual understanding, the data that are used to drive and test a model, the model structure and the model parameter values used for a specific catchment. It is very difficult in many cases to identify which are the main uncertainties, but this does not mean that that any of them should be ignored, or that they can be simply lumped together. It is important that the individual uncertainties are more clearly identified so that we can focus attention on the areas where more data are required or more research needs to be done. It is therefore equally important that the framework, model and data analysis tools used in this project are suitable for assessing different types of uncertainty. The Institute for Water Research has now spent more than 6 years developing an uncertain framework in which we can run the Pitman model and explore different options in terms of input data, model setup (dominance of certain components over others) and parameter value sets. It is therefore possible to explore a wide variety of possible concepts and to try and identify those (if any) that appear to be more behavioural (or consistent with both observations and/or conventional conceptual wisdom) than others. In attempting to determine the most behavioural model setup, the uncertainty framework can be useful in guiding the model user in formulating and streamlining conceptual ideas. The absolute validation of any model's output is not possible in a data scarce country such as South Africa, and it is rarely possible to confirm a single hypothesis in a catchment. However, the aim is to determine whether it is possible to use the model to reject or 'accept as possible' certain hypothesis in many situations.

While it will always be difficult to differentiate between input, parameter and structural uncertainties, it should be possible to at least partly identify the dominant sources of uncertainty by a careful examination of the evidence for specific processes compared with the conceptual structure of a specific model. While the lack of appropriate data means there will always be considerable uncertainty surrounding model validation, it can be argued that improved process understanding in an environment can be used to validate model outcomes to a degree, by assessing whether a model is getting the right results for the right reasons. The ultimate objectives of this study are to improve the conceptual understanding of surface and groundwater interactions in South Africa, and to assess whether it is possible to reliably quantify these interactions at the catchment scale. One of the more specific objectives is to assess whether the revised version of the Pitman model (Hughes, 2004) can be considered appropriate for achieving this quantification.

One of the key issues in surface and groundwater interactions is how much of the groundwater recharge contributes to stream flow in river channels. There continues to be a great deal of confusion surrounding this issue and the simple concept of the 'baseflow' component of total stream flow. This concept has been a part of environmental flow requirement (EWR) studies for a long time. This is justified because, to a large extent (ignoring potential water quality considerations), the source of the 'baseflow' is not the primary concern. The important issue in EWR studies is the maintenance of the low flow regime of a river to sustain ecological functioning. Unfortunately, the 'baseflow' component of total stream flow has been interpreted as being derived from groundwater outflows, which in many catchments is patently not the case. Parsons (2004) and Hughes (2010a) provide some detailed explanations of why this may not be the case and presented evidence in some catchments where baseflows are clearly not derived from groundwater (see Section 1.1 for more details). It is therefore interesting to note that this concept is still used by some and forms the basis of their estimation techniques. It is therefore not possible to analyse observed stream flow time series and subject them to a mathematical baseflow separation approach and determine how much groundwater contributes and by extension, how much groundwater should not be exploited if the low flows in the rivers are to be sustained. It is therefore essential that we develop better understanding and use that understanding in appropriate models that combine the main processes that link surface and groundwater systems. It is also important that these modelling approaches are sensitive to the key environmental and anthropogenic changes that we have already experienced and that we are likely to experience at an increasing rate into the future. These include land use changes, water abstraction changes and climate changes.

1.1 Defining key terms

The correct and consistent use of hydrological terms is considered essential for developing a better understanding of surface and groundwater interaction. It is assumed that basic terms and principles for groundwater systems are understood, for example, *confined* and *unconfined aquifer systems*, *hydraulic head, transmissivity* and *storativity*. The term *fractures* refers to all cracks, fissures, joints and faults that may be present in a formation. These terms are defined in basic texts such as Davis and DeWiest (1966) and the essential terms are included in the glossary of this report. There are however, terms which can be ambiguous and numerous authors have called for a more consistent use of the terminology amongst all hydrologists and groundwater hydrologists (DWAF, 2004a; Parsons, 2004; Hughes, 2010a). Perhaps the most pertinent is the call for clearer definitions of the terminology used to refer to the low flow components of stream flow. This is particularly relevant in the frequent use of the term *baseflow* without any reference to the source of this water. This leads to potential confusion which can be largely attributed to different perceptions of the dominant stream flow generation processes (Hughes, 2010a). Parsons (2004) proposed *baseflow* be used as non-process related term to signify low amplitude, high frequency flow in a river during dry or fair weather periods. This approach would be compatible with hydrograph separation methods that do not attempt to identify the source of the *baseflow* component (Hughes, 2010a). *Baseflow* can originate from a variety of sources including the regional groundwater body, seepage of percolating water from outcropping fractures, springs draining perched water tables, interflow through the soil and weathered zone, artesian springs, high lying springs above the regional valley bottom aquifer or through the attenuation of storages such as wetlands. When referring to a potential source of *baseflow*, this should be explicitly stated.

Another term which can cause confusion is recharge. The quantification of recharge based on either surface or sub-surface methods can be different due to different interpretations of the recharge process. Recharge is often quantified based on the infiltration of water into the sub-surface, but this method ignores the numerous processes which can occur in the unsaturated zone before the infiltrating water reaches the regional water table. Only that portion of the infiltrating water which reaches the regional groundwater table should be deemed recharge. The *unsaturated zone* is defined in this report as the area above the regional groundwater table. Perched aquifers are included in the unsaturated zone as they are situated above the regional groundwater table. Parsons (2004) provides comprehensive explanations of the relevant terminology for the study of surface and groundwater interactions and the terminology utilised here will follow the same guidelines.

2. BASIC CONCEPTS OF SURFACE WATER AND GROUNDWATER INTERACTION

The interactions between surface and groundwater systems remain poorly understood in many catchments throughout the world and yet they are essential to effectively managing the quantity and quality of water resources. It has been well documented that these systems interact in a range of geological, topographical and climatic settings and that many surface water features, such as rivers, lakes, dams and wetlands will have varying degrees of connection with groundwater systems (Fleckenstein *et al*., 2010; Haria *et al*., 2013 and Ivkovic, 2009). A number of comprehensive contributions to the understanding of the physical interactions that occur between groundwater and surface water systems have been written internationally (Brunner *et al*., 2009; Lamontagne *et al*., 2014; Sophocleous, 2002; Unland *et al*., 2013 and Winter, 1999). Locally, papers have tended to focus more on management issues surrounding the interactions between surface and groundwater (Le Maitre and Colvin, 2008; Levy and Xu, 2012; Parsons, 2004 and Xu *et al*., 2002). South African research that has focused on the physical processes behind the interactions has tended to focus on relatively small scale studies (Hughes and Sami, 1992; Kelbe and Germishuyse, 2010; Lorentz *et al*., 2004; Roets *et al*. 2008 and Wenningger *et al*., 2008). Both local and international contributions have classified the different types of interactions that occur between surface and groundwater systems. Some of the literature focusses on the processes occurring at detailed scales (Frei *et al*., 2009; Sophocleous, 2002; Ward et al., 2010 and Winter *et al*., 1998), while others focus on larger scales for the purpose of water resource estimation (Gilfedder et al., 2012; Smakhtin, 2001 and Vegter and Pitman, 2003). One of the most striking differences between international and local contributions is the amount of available data used within the studies. Most of the international case studies are based on channel reaches or catchments with extensive data sets, in stark contrast to many of the South African case studies that are based on scarce data.

2.1 Hydrogeological framework

The occurrence of groundwater within the Earth's crust and its emergence at the ground surface are determined by the lithology of geological materials, regional geological structure, geomorphology of landforms and the availability of recharge sources (Hiscock, 2005). The qualitative understanding of the processes occurring in a catchment, such as the geological framework, the location, types and characteristics of the aquifers in the study area, and the runoff characteristics can be termed a perceptual model (or conceptual understanding). It is difficult to produce generic perceptual models of interaction types and while this has been undertaken in many parts of the world (Winter *et al*., 1998; Roets *et al*., 2008; Banks *et al*., 2009), many of the classical ideas incorporated into perceptual models from Europe or North America encompass very different processes to those found in South Africa. For example, in semi-arid environments the groundwater level is very deep and therefore the unsaturated zone is far larger than that found in temperate environments. This renders the numerous international perceptual models not applicable to many of the environments found in South Africa as far more focus on the processes occurring in the unsaturated zone is necessary. General perceptual models were produced, using information available in the literature, in reports, as well as expert opinion, for four common South African interaction environments. These include fractured rock aquifers (found over the majority of South Africa), karst aquifers (which cover only 2.7% of the country but are the most productive aquifer types), primary aquifers (important over many of the coastal areas of South Africa) and alluvial aquifers (found largely in semi-arid basins in the interior of the country).

Hard Rock aquifers

Many authors have contributed to the improved understanding of hard rock (also termed fractured rock and secondary) aquifers from a variety of perspectives (Nastev *et al*., 2004; Rodhe and Bockgard, 2006). However, there is still a large amount of uncertainty associated with the characterisation of hard rock aquifers. Part of the problem lies in defining the type of information required. The inherent structural and hydraulic complexity of fractured rock severely limits the type and quality of data that can be obtained from field measurements. An important consideration when beginning an investigation is that the problem of interest, as well as the required data and their methods of interpretation, are a function of scale. Therefore before conceptualisation, data collection and modelling, it is essential to consider the relevant scale of the problem of interest, the end use of the data and the required level of detail (Berkowitz, 2002). Voss (2003) suggested that *a priori* characterisation of fracture systems might be impossible. It has been suggested (Krásný and Sharp, Jr, 2007) that at the larger scale, the average permeability remains roughly constant, irrespective of the position of the field investigation within the entire environment. If so, this represents a regional transmissivity value that corresponds to a representative storage. At larger scales no change in mean values would be expected, although it is often necessary to account for higher permeability zones such as dolorite dyke intrusions and regional fault zone structures (Botha and van Rooy, 2001; Sami, 1996). This is an important concept for models (both conceptual and numerical) as it suggests to what extent hydrogeological conditions can be represented. Many investigations into fractured rock environments (using both numerical and conceptual models) assume a quasi-homogeneous flow through the bedrock by taking into account conductivity ranges as determined from hydraulic aquifer tests (Illman and Neuman, 2000) and the effects of structures such as faults and fractures are averaged.

Hard rock environments typically consist of three vertical zones, upper weathered, middle fractured and deep massive. These zones often correspond to the three vertical zones of flow which form part of general hydrogeological principles, schematised by several authors (Toth, 1963; Winter, 1999). The zones have been designated from the land surface downwards, as local (intensive, shallow), intermediate (retarded) and regional (slow or negligible, deep, stagnant) (Krásný and Sharp, Jr., 2003). The weathered zone can be formed by saprolite, colluvium, talus etc. and is often present along with alluvial, fluvial and lacustrine deposits. Many headwater catchments in South Africa are frequently important sources of recharge in fractured rock aquifers due to favourable climatic and geomorphological conditions. High precipitation and low evapotranspiration promote high and relatively uniform recharge. In addition, these steep areas often have little or no soil cover which enables concentrated and rapid recharge through fracture zones in outcropping hard rock. High hydraulic gradients in mountainous headwater catchments can result in intensive groundwater flow in spite of the prevailing low transmissivity of rocks.

Alluvial aquifers

Several authors have documented diverse alluvial interaction scenarios that occur in several parts of the world. Some authors have investigated the effects of transmission losses on stream flows and on the recharge of the alluvial aquifers (Hughes and Sami, 1992; Subyani, 2004) while others have investigated scenarios where the alluvial aquifers recharge the stream (Negral *et al*., 2003; DWAF, 2008a). Studies documenting complex interaction scenarios within alluvium (Osman and Bruen, 2002; Tooth *et al*., 2002) have highlighted the importance of a sound perceptual model as well as the danger of over-simplifying the interaction processes. Although the significance of transmission losses has been known for many years (Dubief, 1953; Schick, 1988), relatively little is known about the processes involved, as gauging and monitoring of surface water flow in arid environments is often scarce (Lange, 2005). One of the problems of obtaining this understanding in many semi-arid areas is the meagre number of flow events that occur (Hughes and Sami, 1992). However, the attenuation of flow is expected to occur through recharge of the aquifer, and when the river is flowing permanently it is assumed that the alluvial aquifer is full and no attenuation will take place, unless artificial abstraction disturbs the water balance. Riparian evapotranspiration is also likely to affect the water balance of alluvial aquifers in a major way and yet there is very little information published on this process.

Stream alluvium is usually relatively coarse grained and can transmit and store large quantities of water. Fine grained alluvial deposits (commonly found on flood plains) often has a large porosity and can store large quantities of water, however the hydraulic conductivity is low and water is not as readily released from storage. Perched water zones can overlie parts of finer grained facies and interbedded lenses of fine grained sediment. The volume of transmission losses can vary according to characteristics such as the time interval between successive flood events, intensity of flood events and length of flood events. During a flood event modifying processes (e.g. air entrapment, scour and fill) are active in varying transmission losses. For example, silt carried by flood waters can effectively seal the channel bed or alluvial surface, even during high velocity events (Tooth *et al*., 2002). The interplay between scour and fill is complex and not fully understood, which makes reliable quantification of infiltration rates and loss volumes during real flood events difficult to achieve, especially in large scale systems (Lange, 2005). Dahan *et al*. (2008) concluded that it is the limits on storage capacity combined with the regulated minimum flux rate through the riverbed that control recharge. Most of the infiltration will take place in the main active channel at similar rates, independent of the flood stage. While infiltration in flood plain areas might not contribute as significantly to groundwater recharge of the surrounding aquifer, flood plain water losses are important in that they can significantly decrease flood discharge downstream (Knighton and Nanson ,1994; Lange ,2005.

In southern Africa, groundwater is frequently located in a strip of alluvium along main river stems surrounded by weathered and fractured rocks in tributary catchments. In South Africa alluvial aquifers are found in several regions in the interior where alluvial sediments have accumulated due to the low relief, or in extensive basins such as the Kalahari. In the Limpopo Basin, the northernmost water management area in South Africa, alluvial deposits have a patchy distribution along the major rivers. At the northern fringes of the area, alluvial strips along the Limpopo and Sashi Rivers are limited to a maximum of 1.5 km in width and 25 m in thickness (Busari, 2008). Most of the rivers in this semi-arid area are ephemeral and rely on infrequent flow events to recharge the alluvium and surrounding fractured rock aquifers. However, the main Limpopo River is predominantly a perennial system (under natural conditions), as are some of its larger tributaries such as the Nyl River. Due to variable topography in this region, related to complex tectonic deformation, extensive flood plain vleis, such as the Nylsvlei are formed where the river emerges from high relief areas onto low flood plains which can form extensive wetland areas (Frost, 1987). In the Cape Fold Belt, alluvial fans are formed from fast flowing perennial rivers in narrow valleys emerging from mountainous areas onto flatter valley floors. Evapotranspiration is a large component of the water balance in many alluvial environments in South Africa as most environments fall within semi-arid areas with high potential evaporation rates. Groundwater stored in the flood plain aquifers is accessible to floodplain plant communities such as gallery forests (Le Maitre and Colvin, 2008). The riparian margins of streams can remove large volumes of water temporarily stored in the channel margins after recharge events, and directly from groundwater in some environments.

Karst aquifers

Karst systems are often characterised by substantial surface and groundwater interactions via processes such as aquifer recharge by losing streams (Baffaut and Benson, 2009), fracture and conduit connections between surface and groundwater and spring flow contributions to surface water (Swart *et al*., 2003). Temporal variations in geochemical parameters such as strontium isotopes, stable isotopes and anthopogenic contaminants have been observed in karst systems in response to variations in flow and recharge (Boyer and Kuczynska, 2003; Barbieri *et al*., 2005). Musgrove *et al*. (2010) asserts that an understanding of temporal variability in karst systems provides insight into hydrologic processes and aquifer structure and is required when managing the system in an integrated fashion. It is not straightforward to delineate catchment boundaries which encompass both the surface water and groundwater system. Realistic water balances can only be carried out once the groundwater catchment is known in relation to the surface water catchment. If the groundwater catchment size is unknown, normalised baseflow estimates from springs have been used to provide reasonably reliable estimates of the catchment area (Bailly-Comte *et al*., 2009). Normalised baseflow is defined as baseflow discharge per unit area (litres/second/km^2) and can be used to estimate the approximate recharge area of springs. If the groundwater catchment size is known, a relatively reliable characterisation of the aquifer can be undertaken. Due to the large variability and heterogeneity in karst systems (less developed diffuse flow aquifers to well-developed conduit flow aquifers), which causes their particularly non-linear and non-stationary hydrological behaviour, modelling the rainfall-runoff relationship is not straightforward. For these reasons, surface and groundwater interaction studies are mostly based on hydrochemistry (Moore et al., 2009).

Karst terrains are characterised by (1) closed surface depressions of variable sizes and shapes known as sinkholes, (2) an underground drainage network that consists of solution openings that range in size from enlarged cracks in the rock to large caves and (3) highly disrupted surface drainage systems, which relate directly to the unique character of the underground drainage system (Winter *et al*., 1999). The hydraulic conductivity of a karst aquifer is largely dependent on how well developed a karst system is. An advanced karst system will have a large carrying capacity which can accommodate even extreme recharge events from large storms. In these systems there is no surface flow across the karst. In karst systems with medium development, there might be no baseflow as the carrying capacity of the aquifer is large enough to accommodate all baseflow but during extreme storm events, some surface flow would occur. In karst systems with a small carrying capacity, there is likely to be a perennial stream, although perhaps with less flow than would be expected given the rainfall and size of a catchment (White, 2002). Similar to fractured rock systems, the measurement of effective permeability is completely scale dependant (Scanlon et al., 2003). While karst aquifers often have a high storage capacity, their capacity for water retention is low. Seeps and springs of all sizes are characteristic features of karst terrains. This spring discharge accounts for the runoff from the entire karst groundwater basin including allogenic inputs. This is an important attribute of karst systems as the discharge water carries an imprint of all water sources upstream of the spring (Le Moine *et al*., 2008). Most of the dolomitic eyes in South Africa are dammed springs that are formed when vertical dyke intrusions dam the groundwater in the aquifer resulting in rising groundwater levels (Swart *et al*., 2003), however dolomitic eyes can also be morphology controlled where the relief drops below the upstream groundwater table (Winde and Erasmus, 2011).

Primary aquifers

Primary aquifer types include sand aquifers (coastal, dune, lowland floodplains etc.) (Trojan *et al*., 2003; Kelbe and Germishuyse, 2010), chalk aquifers (Lee et al., 2006) and glacial deposits (Delin *et al*., 2006). Also included in this category are regolith or saprolite aquifers (Banks *et al*., 2009). While these types of aquifers are associated with the upper weathered zone of hard rock, they display many of the characteristics of primary aquifers and are therefore included in this classification. Contributions have highlighted the relative homogeneity of some primary aquifers (Kalbus *et al*., 2009), but also the potential heterogeneity of many of these aquifers (Diem et al., 2010). While primary aquifers are the most homogeneous of aquifer types, this type of system does display spatial variability in flow on local and macroscopic scales. In largely homogeneous primary aquifers, it is more straightforward to characterise the aquifer properties, although a relatively small degree of variability does exist. This type of aquifer is unlikely to have much of an interflow or lateral flow component (Kelbe and Germishuyse, 2010). Primary aquifers can also be extremely variable in terms of hydraulic characteristics like porosity and permeability. Hydraulic conductivity of heterogeneous porous aquifers has been found to vary over several orders of magnitude (Diem et al., 2010). Heterogeneous porous aquifers often have significant amounts of clay and silt, which can lead to perched aquifer conditions and confined conditions. Evapotranspiration directly from groundwater is prevalent in coastal and dune terrain with many plants having root systems deep enough to transpire a high volume of groundwater (Winter *et al*., 1998).

In South Africa only a few coastal primary aquifers are of significance. While unconsolidated sands often tens of metres thick are found in the Kalahari Basin, these form poor aquifers and most groundwater bodies lie within the underlying Paleozoic and Mesozoic rocks and basalt of the Karoo Supergroup. Examples of significant primary aquifers are found at Atlantis and on the Cape Flats of Cape Town and along the Zululand coast north of Durban (Parsons, 2004). These coastal aquifers are frequently shallow unconfined aquifers where the water table varies in undulating form and slope, depending on areas of recharge and discharge and hydraulic properties of the porous medium. The groundwater direction is toward the coast and, in general, groundwater levels mimic the topography, although Conrad *et al*. (2004) concluded that the surface and groundwater catchment divides of the West Coast Sandveld aquifer do not coincide. The aquifers along the arid west coast of the country are underlain by fractured rock aquifers which are recharged inland in areas of higher relief and rainfall and discharge at the coast into the sand aquifer (Conrad *et al*., 2004). It is important to include the connections between the fractured rock and coastal sand aquifer to obtain a reliable water balance for the region as a whole (Munch *et al*., 2013).

2.2 Groundwater and topography

Although groundwater regions and aquifer types have been classified, the geologic complexity of South Africa and many countries means that much uncertainty still exists. Currently water resource estimation is carried out on a surface water catchment scale and the existing groundwater database, GRA II (DWAF, 2005) reports aquifer parameters like recharge, transmissivity and storativity at this scale. There has been criticism, however, that surface and groundwater divides do not coincide and that groundwater in South Africa does not follow the topography, therefore it is not sensible to report groundwater parameters on a surface water catchment scale. Much of the literature considers that the water table in unconfined aquifers is a subdued replica of the topography or land surface (Gardener, 1999; Wright and Xu, 2000; Parsons, 2004). However, there are authors who do not agree with this assumption (Haitjema and Mitchell-Bruker, 2005; Devito *et al*., 2005; Worman *et al*., 2007). Haitjema and Mitchell-Bruker (2005) maintain that only under certain conditions does the water table follow the topography. They argue (using numerical models) that the most important indicator is the recharge to hydraulic conductivity ratio, only with a high recharge and low hydraulic conductivity (as a dimensionless ratio) could the water table possibly follow the topography (Figure

2-1). Worman *et al*. (2007), however, examined fractal distributions of recharge, discharge and associated sub-surface flow patterns caused by the land surface and maintain that topography remains a primary control even for deep groundwater circulation.

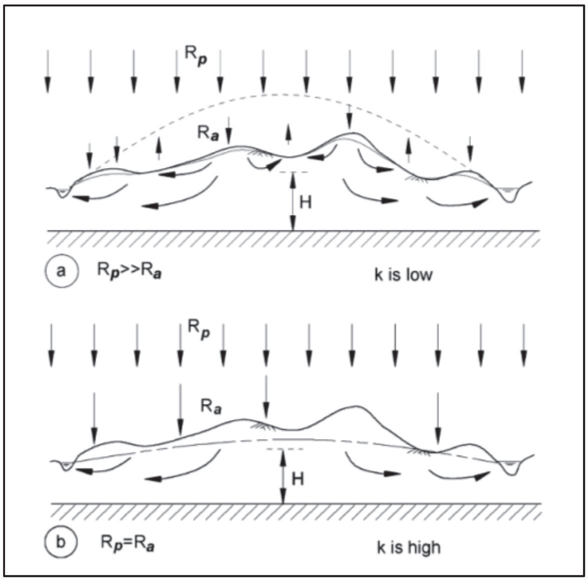

Figure 2.1 (a) Low permeable aquifer with topography controlled water table; (b) highly permeable aquifer with a recharge controlled water table (Haitjema and Mitchell-Bruker, 2005).

There are, therefore, two possibilities for establishing groundwater management units. The first is to use the surface water catchments, which is in line with the hydrological approach. This can be done if groundwater head elevations are highly correlated with surface elevations. Vivier (2009) carried out a comparison between topography and groundwater levels in the Outeniqua catchment, and the results showed a good correlation with R^2=0.9 for the data (272 boreholes). The author also maintained that the majority of groundwater assessments show an acceptable correlation between groundwater head elevation and topography. Deviations occur where permeable aquifers are stressed by over abstraction on a local scale. Wright and Xu (2000) concur and argue that it is logical to attempt an initial groundwater volume balance exercise at surface water catchment scales.

The second view is to use groundwater resource units (GRU) where the geological units are grouped into hydrogeological units and characterised across quaternary catchment boundaries. Groundwater flow systems are defined by the boundary conditions imposed by their physiographic framework and by the distribution of recharge. Generally natural groundwater systems do not have simple boundary conditions and are not composed of isotropic and homogeneous porous media (Winter, 1999). Vivier (2009) recommend that when this approach is required, the GRU be defined as a secondary management unit and that the management of groundwater on a surface catchment scale be followed for primary consideration. Wright and Xu (2000) suggest that in the event of it not being possible to account for all the water entering or leaving the unit, it may be necessary to include adjacent catchments until a large portion of the groundwater can be accounted for. This scenario could also include having to zoom in on a portion of the surface water catchment to achieve a water balance. The larger the scale of the assessment the more groundwater head elevation is expected to follow topography. In South Africa, hydrological data are available on a surface catchment scale for the entire country. It would complicate groundwater quantification for management purposes further if GRUs would be used as the primary management unit. It is also impossible to reliably define all the inflows and outflows from a GRU. In any assessment, use of the method with the least unknown factors would be recommended. However, both methods discussed are plagued by data problems.

3. CONSTRUCTING A WATER BALANCE

Realistically models can only be evaluated if there is an adequate conceptual understanding of the processes involved in surface water and groundwater interactions. A simplified representation of the main processes is shown in Figure 3.1 along with key questions associated with these processes. Resolving the uncertainty surrounding these questions is vital to ensuring that models are sensibly representing the dominant processes and their interactions. However, without the right type of data, many of these issues could be irresolvable.

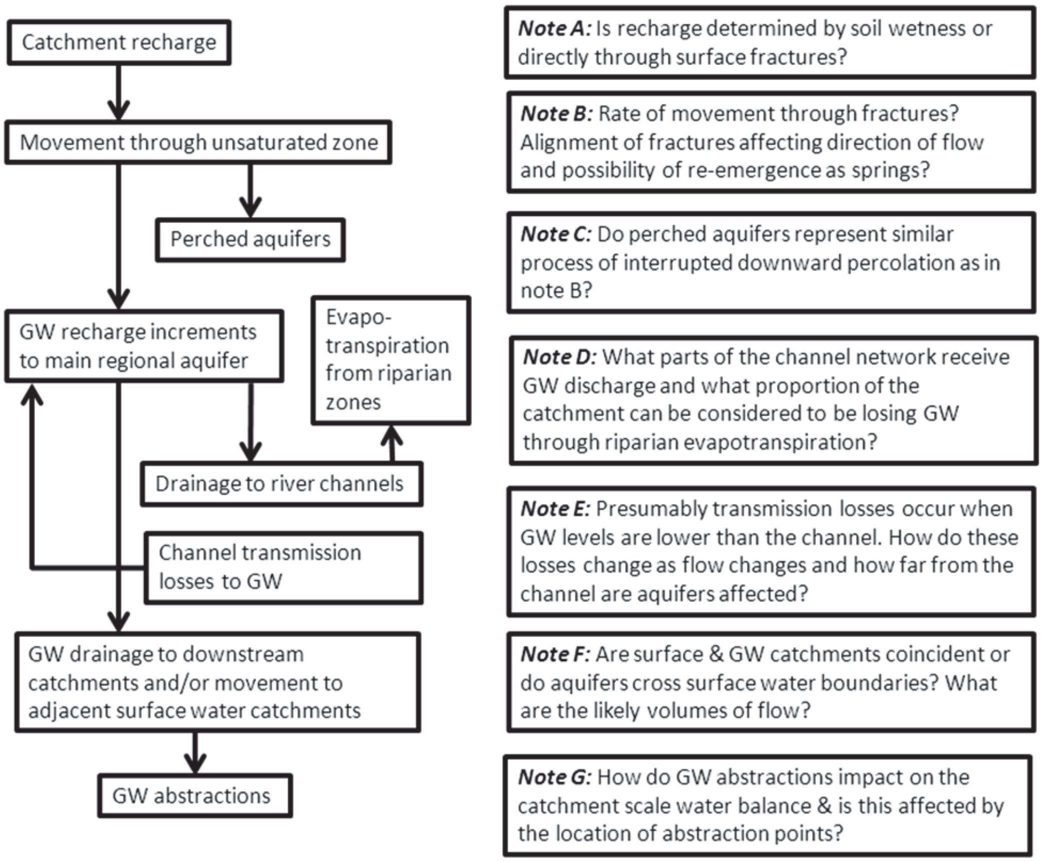

Figure 3.1 Large scale processes associated with surface water and groundwater interactions and key questions associated with the various processes.

Recharge to groundwater is generally assumed to either leave the immediate surface water catchment as sub-surface transfers to other surface water catchments, contribute to stream flow within the surface water catchment, be lost to evapotranspiration (in the riparian margins of the channel), move downstream as groundwater flow or be abstracted. The basic water balance can be represented by the following equation:

$$\mathbf{R = ET - BF_{GW} - DS_{GW} - ABS + CL + US^{GW} + \Delta S} \qquad \text{Equation 3.1}$$

Where:

R	=	recharge
ET	=	riparian evapotranspiration
BF_{GW}	=	groundwater baseflow
DS_{GW}	=	downstream groundwater flow
ABS	=	abstractions
CL	=	channel or transmission losses
US_{GW}	=	upstream groundwater inflow
ΔS	=	the change in groundwater storage

Most of the processes that form part of the surface-groundwater interaction water balance equation are almost impossible to measure directly at all but very small scales. Therefore they are generally estimated using indirect methods or based on extrapolations of local scale observations. Some idea of the relative importance of many of these outflow components can be gained by examining information about the topography, regional groundwater gradient and hydraulic conductivity of the environment. However, much of this information is obtained from borehole records that are inherently biased, especially in highly heterogeneous fractured rock aquifers, as they are typically located within the highest yielding components of the aquifer. Even records of groundwater use are frequently uncertain or poorly quantified. In areas with little data, groundwater abstraction and return flows can be approximately derived from estimates of area under irrigation, crop types, irrigation requirements, population or livestock supported, pumping capacity and pumping hours. The change in groundwater storage is often estimated as a residual of the water balance, and is never straightforward to quantify in fractured rock or karst environments. Data from aquifer pumping tests combined with aquifer depths from borehole logs can be useful to approximately estimate storage in these heterogeneous environments. These approximations however, result in highly uncertain estimates of interaction volumes at the scale of quaternary catchments.

3.1 Catchment recharge and infiltration into the unsaturated zone

Infiltration of water into the subsurface varies significantly over a catchment both spatially and temporally, especially in arid and semi-arid zones and it is not straightforward to trace the path of this water on a scale that might be useful for water resource management. Although the processes behind catchment infiltration and subsequent recharge of groundwater can be complex, a number of authors (Dawes et al., 2012; McCallum et al., 2010) found that the most important factor is annual rainfall. However, other parameters such as rainfall seasonality, rainfall intensity (Barron et al., 2012), air temperature, humidity, surface slope, interception, the presence of macropores and fracturing, moisture retention capacity of the unsaturated zone and the concentration of atmospheric CO_2 also exert an influence, emphasising the importance of local studies (Dawes et al., 2012; DWAF, 2004b). Reviews of field studies of groundwater recharge have attempted to investigate how climate characteristics control recharge, but cite a lack of data as a major hindrance to drawing any strong conclusions beyond that rainfall is the major determinant. There have been studies (Barron et al., 2012) that identified rainfall intensity and seasonality as an important determinant of catchment infiltration and groundwater recharge. Anthropogenic influences can also exert a major control on catchment infiltration and groundwater recharge and Dawes et al. (2012) identified areas in Australia where the natural water balance had been significantly altered due to anthropogenic activities. These included pine plantations with significant catchment infiltration which did not translate into expected volumes of groundwater recharge and areas with extensive groundwater use resulting in increased recharge due to larger storage potential from lowered groundwater levels. Dawes et al. (2012) found that reduced rainfall affected areas with low conductivity soils more than areas with highly conductive soils with the former experiencing a greater reduction in catchment infiltration.

In fractured rock environments, traditional catchment infiltration and unsaturated zone theory does not always apply. These traditional theories focus on diffuse and equilibrium modes of flow, and preferential flow through fractured rock does not clearly fit into the theories. Preferential flow is common in certain areas and occurs when infiltrating water bypasses much of the unsaturated medium via paths such as worm holes, fractures and fingers of enhanced wetness. Therefore recharge occurs in the absence of a wetting front, or occurs within such a short time frame that does not allow water to first reduce soil moisture deficits or be evaporated. Resulting model estimates (from models which do not allow recharge to occur in the presence of a soil moisture deficit) could be in error (Cuthbert et al., 2013). Preferential flow can depend on threshold effects, possibly related to input fluxes or antecedent soil moisture (Hardie et al., 2011; Cuthbert et al., 2013). Ireson et al. (2009) identified three modes of flow in a fractured system, slow matrix flow, non-preferential fracture flow (recharge responses which lag effective rainfall by 10s of days) and preferential fracture flow (large water table responses within 1 day). They propose that the magnitude of the rainfall event is a good predictor of the occurrence of preferential recharge responses. They found that in moderate rainfall events, it was necessary to take the antecedent soil moisture into account when predicting the groundwater response. However, it was noted that it is likely that the thresholds associated with preferential recharge, in both the rainfall event magnitude and antecedent soil moisture will be site specific. On a regional scale, infiltration into fractured aquifers (through exposed surface fractures) may be highly dependent on the orientation of the main permeable structures, fracture properties (length, aperture, filling) and fracture organisation (Jimenez-Martinez et al., 2013).

3.2 Movement through the unsaturated zone

Unsaturated zone water is defined in this report as any percolating or resident water in the unsaturated sub-surface above the regional groundwater table (which may consist of the soil and weathered zone, fracture zones, sandy zones etc.). Included in this definition is water from perched water tables. Interflow is defined as lateral flow from the zone, while vertical flow will become recharge to the regional groundwater table. The processes of interflow can include:

- Lateral drainage from near channel storages such as channel bank soils, alluvium and wetland areas (Smakhtin, 2001).
- Drainage from a perched water table occurring above a zone of reduced percolation (Figure 3.2A).
- Drainage from fracture zones that have a favourable geometric orientation that allows for lateral movement of water (Figure 3.2B). This type of interflow will only contribute to springs or seepages if the topography is steep enough to allow re-emergence of laterally moving water (Hughes, 2010a).
- Flow from a perched water table that occurs as a result of a zone of localised increased lateral conductivity.

Springs resulting from water movement within the unsaturated zone occur most frequently in terrain that is sufficiently steep to intercept the lateral movement component of water in that zone. The length of time these storage zones will take before discharging into the channel depends on the volume of available storage as well as the transmissive properties of the storage zone. In fractured rock environments rates of outflow will depend upon the head of water within the fracture system, the fracture size and density as well as the relative importance of the lateral drainage component compared to the vertical component, which recharges the 'true' groundwater storage (Smakhtin, 2001). While vertical flow will dominate where gravity is the main force of movement, if some of the major fractures are not vertical there is expected to be a lateral component to the flow (Figure 3.2B). A lateral component may also occur when the rate of recharge at the surface is more rapid than the rate of vertical percolation within the fractures. In fractured unsaturated zones the process of vertical percolation may decrease as fracture density reduces with depth, further promoting lateral flow (Hughes, 2010a).

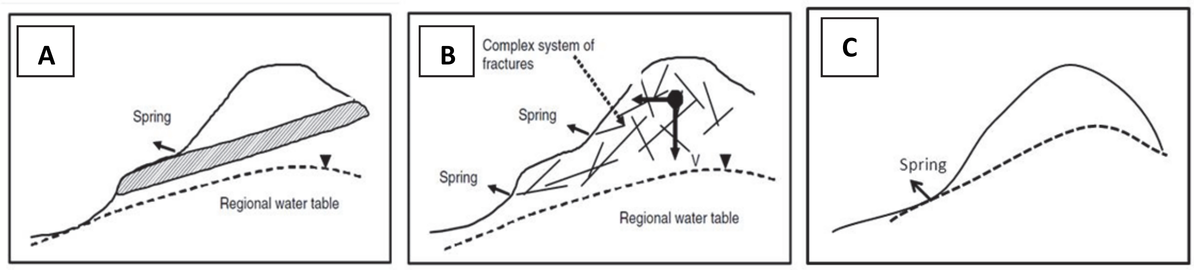

Figure 3.2 A shows a spring resulting from a confining layer; B illustrates the scenario where springs result from a lateral component to otherwise percolating recharge water; C illustrates the scenario where springs result from the intersection of the regional groundwater table with the surface (adapted from Hughes, 2010a).

There are methods that have been utilised to carry out process studies in unsaturated zones. However, differences in the sources and pathways of the water collected using different methods can have a major effect on the interpretation of the results (Landon *et al*., 1999). In many examples, macropore development is common and preferential flow is likely the dominant flow mechanism. Differences in recharge rates based on chloride data in the unsaturated zone (averaged 3 mm year^{-1}; range 1-10 mm year^{-1}) and saturated zone (averaged 7 mm year^{-1}) in the eastern Kalahari Desert were attributed to focused flow and preferential flow (de Vries *et al*., 2000). Similarly evidence of preferential flow in the southern Kalahari in South Africa is indicated by higher recharge rates based on tritium distribution (13 mm year^{-1}) relative to those based on chloride in the unsaturated zone (1.8 and 5 mm year^{-1}) (Butler and Verhagen, 2001). Although preferential flow in sandy soils can be a dominant flow mechanism, it is still relatively homogeneous when compared to the preferential flow encountered in fractured rock environments due to a smaller contrast in hydraulic properties between the matrix and preferential pathways (Landon *et al*., 1999). Environmental tracer concentrations have been used to determine the importance of preferential flows in fractured rock aquifers but significant challenges in interpreting tracer concentrations arise because of geologic complexities that result in abrupt spatial changes in hydraulic properties. Fractures, joints, conduits and vugs are the principle pathways for flow and have hydraulic properties that vary over orders of magnitude. The complex connectivity of these features yields highly convoluted flow paths over distances from metres to kilometres, where velocities can also vary over orders of magnitude (Shapiro, 2011). Flow within these environments yields preferential flow paths and areas subject to slow advection or stagnant water. The volume of fractures that contribute to preferential flow can be significantly less than the total fracture volume (Shapiro, 2011). In addition, fractured rock is characterised by flow regimes where a few preferential flow paths can control the majority of groundwater flow.

Field observations of rapid water table rises indicate that vertical unsaturated zone preferential flow can contribute substantially to groundwater recharge (Gleeson et al., 2009; Cuthbert and Tindimugaya, 2010). These empirical water table fluctuation methods incorporate the impact of both preferential flow and diffuse flow processes into an area-averaged estimate of unsaturated fluxes. Mirus and Nimmo (2013) maintain that preferential flow needs to be explicitly represented in models as numerous observations document preferential flow under far from saturated conditions and as a result, simulations often underestimate unsaturated fluxes of water. While the evidence for preferential flow is considerable, further studies are needed to improve the quantitative understanding of the process (Nimmo, 2012). Ireson and Butler (2011), however, maintain that groundwater resource assessment is generally less sensitive to recharge timing, so simple models may still be suitable as long as they predict the long term volume of recharge with some degree of accuracy Recent work on improving the quantification of preferential flow has focused on the development of physically based models (Mathias et al., 2006), source responsive models (Mirus and Nimmo, 2013) and improved process understanding through field investigations (van den Daele et al., 2007). Ireson et al. (2009) examined the use of physically based models to predict preferential recharge and found limitations in terms of lag times between the matrix and fractures, as well as the difficulty of parameterising the models. They did identify the dual continua approach of Gerke and van Genuchten, 1993) as the best physically based model examined. Shapiro (2011) examined the results of quantitative tracer tests in carbonate and fractured rock environments and identified an initial rapid breakthrough of the tracer and a peak concentration (in hours to days), followed by an elongated declining limb of the breakthrough curve over weeks to months. The dilution experienced from recharge to discharge locations indicates contributions from other recharge locations and less permeable fractures. The elongated declining limb indicates that the tracer is migrating through flow paths having slower velocity than the most permeable features. Although interpreting results from environmental tracers and controlled tracer tests in fractured rock environments is challenging, it is clear that there is tremendous variability in velocities encountered in complex geologic settings. The connectivity of fractures is a critical feature controlling water movement in the unsaturated zone

(Berkowitz, 2002). Cuthbert et al. (2013) detail several investigations where preferential flow varied markedly depending on macropore connectivity, depth and conductance.

3.3 Groundwater recharge to the regional aquifer

Recharge is often the variable to which a model's simulated results are most sensitive (Delin *et al*., 2006) however, the measurement of recharge is not straightforward and estimates often vary considerably for the same region. Lerner *et al*. (1990) simplified recharge into the following categories (combinations can occur):

- Direct recharge, (also termed diffuse recharge) which occurs by direct vertical percolation through the unsaturated zone, derived from precipitation or irrigation that occurs fairly uniformly over large areas.
- Localised recharge, (also termed focused recharge) which results from a concentration of water in the near surface zone from small depressions, joints or rivulets.
- Indirect recharge, due to transmission losses from surface water systems such as rivers and lakes.

Rushton and Ward (1997) also distinguish:

- Actual recharge, which reaches the water table, estimated from groundwater investigations.
- Potential recharge, defined as infiltrated water that may or may not reach the water table due to unsaturated zone processes or the ability of the saturated zone to accept recharge, estimated from surface water and unsaturated zone studies

While it has been established that rainfall is a dominant control on groundwater recharge, Barron et al. (2012) attempted to distinguish specific rainfall characteristics that affect recharge in different environments. They examined climatic controls on diffuse groundwater recharge across Australia for a range of Koppen-Geiger climate types (tropical, arid and temperate) and identified certain trends which are outlined below:

- Annual rainfall is a major factor influencing recharge. Although for most of the climate types, recharge shows a greater dependency on rainfall parameters which reflect higher rainfall intensity. The exceptions include tropical climate types with largely high intensity rainfall and arid areas with largely low intensity rainfall.
- Annual recharge is more sensitive to daily rainfall intensity in regions with winter dominated rainfall, where it is also less sensitive to absolute changes in annual rainfall.

- A comparison of winter and summer dominated rainfall regions with perennial vegetation and trees under the same annual rainfall and soil conditions, indicates less recharge in the winter dominated regions (this trend is not observed under annual vegetation).
- The relative importance of annual rainfall in recharge estimation reduces under lower rainfall conditions, and along with that an increase in the relative importance of other climate parameters (temperature, solar radiation and vapour pressure deficit). The effect of climate parameters other than rainfall is greater under summer dominated rainfall as well as cooler climate types.
- An increase in rainfall intensity leads to:
    - An increase in recharge.
    - A higher proportion of rainfall that becomes recharge.
    - An increase in the relative importance of rainfall.
    - A reduction in the relative importance of other climate parameters.
- There is a non-linear relationship between recharge and rainfall, which is likely due to the effect of rainfall intensity or duration of consecutive days with rainfall. Therefore the proportion of recharge to annual rainfall (R/P) is not likely to be a constant – even under the same land cover and soil type.
- Annual changes in recharge are largely proportional to annual changes in rainfall but are not equal. It has been demonstrated that changes in annual rainfall lead to 2- to 4-fold greater changes in recharge.

In arid and semi-arid regions, such as the western portion of South Africa, recharge events are scarce and erratic. Simmers (1998) states that as aridity increases, direct recharge is likely to become less important than localised and indirect recharge. Numerous studies have shown that very little flow percolates through the soil matrix to any significant depth, even with high rainfall (Sami, 1992). In these areas extreme local variability in recharge results from focused/indirect recharge beneath ephemeral streams and preferential flow mostly in fractured systems (Scanlon *et al*., 2006). This is because large storm thresholds are required to overcome large soil moisture deficits and initiate direct recharge through the soil matrix. Recharge from transmission losses seems to vary considerably between regions and in some areas only abnormally high rainfall events have an impact on groundwater recharge. Scanlon *et al*. (2006) synthesised the findings from around 140 recharge study sites in arid and semi-arid regions and found that average recharge rates estimated over large areas range from 0.2 to 35 mm year^{-1}, which represents 0.1 to 5% of long term annual precipitation.

Water balance methods are limited in these environments because of such small recharge components relative to errors in the measurement of the other components of the water balance.

There are significant differences in local and regional scale estimates of recharge. Local scale estimates generally are not representative of an entire catchment and regional estimates may be too general to capture recharge variability within a catchment (Delin *et al*., 2006). Techniques which are regionally applicable include chloride measurements, isotope applications, water balance methods using conceptual modelling and rainfall analysis. Due to the difficulty of measuring recharge in fractured rock environments, most recharge figures are estimated by using a model to link recharge to rainfall (Kelbe and Germishuyse, 2010). While Bredenkamp *et al*. (1995) have advocated the use of rainfall-recharge relationships, there are reservations. These include the uncertainty introduced when transferring relationships between rainfall and recharge to areas other than those in which they were derived, the fact that the temporal distribution of rainfall is not accounted for and the fact that the accuracy of the relationship is dependent on the accuracy of the rainfall estimates from which the relationship was derived. However, rainfall-recharge relationships that have been determined provide a useful means of calculating groundwater recharge. Over sufficiently long periods of time, the recharge volume will equal the volume of infiltrated rainfall minus evapotranspiration, frequently termed effective rainfall. On shorter timescales (sub-annual) it is harder to quantify recharge due to the attenuation of effective rainfall by storage in the unsaturated zone, which becomes more significant for increasingly shorter time scales (Ireson and Butler, 2011). Jimenez-Martinez et al. (2013) used frequency domain analysis to investigate the hydraulic behaviour of a fractured aquifer over a range of time scales and identified large variability of the recharge processes in these complex systems. They identified that some wells (where the transfer functions follow a linear scaling) indicated that the aquifer may behave like an equivalent homogeneous one at large scale. On the contrary, other wells indicated the presence of preferential flow paths.

3.4 Groundwater drainage to downstream or adjacent surface water catchments

The works of Tóth (1999) and Sophocleous (2002), as well as others have shown that groundwater discharge is not only confined along the stream channel but also extends throughout the discharge area down gradient from the basin hinge line or surface water catchment boundary. Although this groundwater discharge is difficult to quantify, it is important to include in a catchment water balance to avoid incorrect accounting of the remaining variables. Hence, spatial flow characteristics must be examined in a water balance study. Sophocleous (2002) describes the general characteristics of subsurface flow in a region with irregular topography and identifies multiple flow systems of different orders of magnitude and relative, nested hierarchical order. Based on their relative position in space, three distinct types of flow systems – local, intermediate, and regional – are recognised which could be superimposed on one another within a groundwater basin.

Marklund and Wörman (2007) maintain that landscape topography is the most important driving force for groundwater flow and all scales of topography contribute to groundwater movement. They examined different scales of topography and how they affect groundwater flow at different depths using a spectral analysis of the topography and a couple of exact 3-D solutions of the groundwater flow. While the distribution of hydraulic conductivity in the rocks impacts groundwater flow patterns, Marklund and Wörman (2007) also identified depth dependent hydraulic conductivity as an important factor for groundwater movement. Depth dependant hydraulic conductivity tends to counteract the effect of the large scale topography on the groundwater flow more effectively than the smaller landscape scales. By controlling how deep the groundwater flow cells become, the topography also determines the residence time for groundwater.

Gilfedder et al. (2009) examined groundwater movement in terms of observed time lags between groundwater abstractions and discharge using simplified analytical groundwater theory. They produced a conceptual model which included three different processes that contribute to the overall groundwater response. The first is related to the vertical filling of an aquifer following an increase in recharge and the remaining two time scales are related to the lateral flow of groundwater. One relates to lateral pressure transmission, the other a gradient driven time scale. Aquifers with different characteristics will have varying dominant time scales. Walker et al. (2005) showed that as the catchment parameters (such as slope, hydraulic conductivity, flow length) change between the two lateral movement types, the time response associated with the groundwater discharge function moves smoothly between those two time scales. Without detailed site specific measurements, the precise shape of the groundwater response curve for the aquifer (e.g., exponential decay, log-normal function or a logistic function) will not be known (Sophocleous, 2012). However, the implication is that should a model's parameters adequately represent the physical basin characteristics (and assuming the model represents all of the dominant processes); the estimation of groundwater downstream movement will be functional.

Larkin and Sharp (1992) classify stream-aquifer systems (based on the predominant regional groundwater flow component) as (1) underflow-component dominated (the groundwater flux moves parallel to the river and in the same direction as the stream flow); (2) baseflow-component dominated (the groundwater flux moves perpendicular to or from the river depending on whether the river is effluent or influent; or (3) mixed. However, all three directions of movement coincide with the topography, and therefore follow the downstream direction of surface water systems. There are situations, however, where the groundwater does not conform to the topography and either moves out of a surface water basin through deep regional groundwater flow (to discharge in an alternative surface water basin), or moves into an adjacent surface water basin which is not located downstream. Identifying surface water basins where the groundwater movement is not downstream and estimating the volumes of groundwater flow is very difficult and indications are that the processes need to be better understood before reliable quantification of the fluxes can occur.

3.5 Drainage to river channels and contributions to baseflow

Smakhtin (2001) lists conditions that must be met in order for the groundwater contributing to baseflow to be sustainable: (a) the draining aquifer must be recharged seasonally with adequate amounts of water; (b) the water table must be shallow enough to be intersected by the stream; and (c) the aquifer's size and hydraulic properties must be sufficient to maintain flows throughout the dry season. Similarly Le Maitre and Colvin (2008) argue that sustained aquifer discharge to river systems depends on significant aquifer storativity and transmissivity, maintenance of high water tables and a hydraulic gradient towards the discharge point or zone, and hydraulic connectivity with the river. Surface water and groundwater systems have different response times, which pose significant water management challenges. While the evaporation and runoff hydrologic components may respond relatively quickly to land use change, a reduction in recharge may not express itself as a corresponding reduction in surface discharge for many years. Additional improvements are still needed for models to represent actual landscape processes to provide more realistic predictions of system response times. Methods such as groundwater age dating techniques have been developed and are being used to constrain model parameters (Sophocleous, 2012).

Characterization of baseflow has focused on quantifying groundwater discharge rates and locations using a variety of hydrograph analyses or chemical and isotopic separation techniques, often at the outlet of a catchment (Smakhtin, 2001). Many of these techniques cannot adequately identify the various sources of baseflow. Several in-river methods, including measurement of water chemistry (Tetzlaff and Soulsby, 2008; McCallum et al.,2013; Frisbee et al., 2012) and synoptic sampling of isotopic tracers (Cook, 2003; Gardner et al., 2011) have begun to unravel the internal dynamics of a catchment that will lead to more comprehensive characterisation. Different methods for characterisation must be combined to expand the knowledge of groundwater discharge locations, rates, and residence times (McDonnell et al., 2010). For example, Gardner et al. (2011) described a novel technique to separate the fractions of local-scale and regional-scale groundwater discharge to rivers by sampling a river for naturally occurring tracers (222Rn and 4He). These measured concentrations of tracers in the river were then modelled to estimate the total groundwater discharge and to separate local-scale and regional-scale discharge fractions. When combined with other methods to characterise groundwater discharge, an ability to identify the sources of baseflow in a particular river can be achieved. Such analyses are needed to refine conceptual models in many catchments and in order to facilitate improved water resource management.

Kalbus et al. (2009) examined groundwater fluxes between aquifers and streams to determine whether the streambed or aquifer permeability exerted the dominant control on infiltration. The results indicated that the aquifer has a stronger influence on the distribution of groundwater fluxes through the streambed than the streambed itself. However, the simulation of a homogeneous low-K streambed within a heterogeneous aquifer caused a significant homogenisation of the fluxes in contrast to the simulation of a heterogeneous low-K streambed within a homogeneous aquifer which was not sufficient to cause strong flux variations.

3.6 Channel transmission losses

Transmission losses occur within two main environments in South Africa. These include hard rock areas where river channels often follow lines of structural weakness and surface fracturing, and alluvial environments, where unconsolidated alluvial material underlies the river channel. Transmission losses in alluvial environments can be substantial during both low flows and during the early phases of flood events (Smakhtin, 2001). While transmission losses in semi-arid areas are the dominant source of recharge for many underlying aquifers, Vegter and Pitman (2003) maintain that conditions over much of South Africa prevent rivers from being all but minor localised sources of recharge. They assert that the lateral expansion of the recharge mound built up below the river by infiltrating water is limited and that rocky riverbeds and silty channels limit infiltration. While, studies have shown that recharge depends on the permeability of the aquifer, the distance from the channel and the duration of river flow, the average extent of 'recharge mounds' in South Africa has not been extensively investigated. Elsewhere the degree of indirect recharge from transmission losses varies significantly over different areas.

The relationship between surface runoff characteristics and volumes of infiltration into a channel bed are not well understood and vary significantly. Lange (2005) examined the temporal dynamics of transmission losses in the Kuiseb River in Namibia and suggested that volumes of transmission loss may be related to flow and channel characteristics by means of regression analysis, although this approach can be complicated by unknown lateral inflows. The results indicated that single high magnitude flows are more important than frequent small to medium flows for recharging the underlying aquifer. The author attributed this to enhanced water losses in flooded overbank areas. Morin *et al*. (2009) came to a similar conclusion when studying the same river using a new flood routing model. They hypothesised that with increasing infiltration rate, channel length or active channel width, the relative contribution of high magnitude floods to recharge also increases, whereas medium and small floods contribute less. Hughes and Sami (1992) monitored the soil moisture dynamics of an alluvial valley bottom in South Africa and found that transmission losses were higher for small flow events (75% of runoff) when compared to larger flow events (22% of runoff). Clearly more data of this type over other parts of South Africa would increase the understanding of the local relationships between surface runoff characteristics and transmission losses and clarify the degree to which transmission losses recharge underlying aquifers.

3.7 Evapotranspiration from riparian zone

In a predominantly dry country such as South Africa, evapotranspiration is the second largest component of the water balance after rainfall and accounts for the greatest loss of water from catchments. In arid and semi-arid environments, much of the water that infiltrates the unsaturated zone is removed by evapotranspiration. Izbicki *et al*. (2000) investigated a thick unsaturated zone (130 to 200 m) underlying an intermittent stream in southern California, USA and found that infiltration directly from rainfall outside the floodplain area did not reach the regional groundwater table.

Transpiration ranges from as low as 5% of annual rainfall to 100% (or more in situations where plants are tapping stored water) but generally ranges between 45 and 80% (Le Maitre *et al*., 1999). Evapotranspiration directly from groundwater can cause cones of depression similar to those caused by pumping wells. In some cases this process can remove most or all of the groundwater flowing toward the stream and in extreme cases can draw water directly from the stream into the sub-surface (Winter *et al.*, 1998). Vegetation of different types can transpire water from the soil profile, regolith, saprolite and rock fractures (Stone and Kalisz, 1991). Numerous studies (Ahring and Steward, 2012; Shah et al., 2007) have demonstrated that changes in vegetation alter both recharge rates and water table depths. Deep rooted plants are able to dry out soil profiles and weathered rock, and also tap groundwater directly to considerable depths (Eucalyptus has been recorded at up to 60 m depth) (Stone and Kalisz, 1991). Long term groundwater level rises in cleared areas can result from reduced evapotranspiration volumes (Leduc *et al*., 2001). This process has resulted in a major increase in salinity in dryland areas of Australia as the groundwater level rise is flushing out salts that accumulated in the unsaturated zone over thousands of years as well as increasing the hydraulic gradients toward streams (Cook *et al*., 2001; Herczeg *et al*., 2001).

While investigations into the vegetation-groundwater interactions often focus on alluvial aquifers or primary aquifers, research into the relationships in fractured rock environments is lacking. Afforestation of the whole of the grassed Mokobulaan research catchment in South Africa, with Eucalyptus grandis, led to the stream drying up after 9 years. At 16 years of age the trees were clearfelled although stream flow did not stabilise until 5 years later. While the soils in the catchment are very shallow, the roots of the eucaluptus penetrated more than 10 m into the fractured shale bedrock (Scott and Lesch, 1997).

Ahring and Steward (2012) examined the relationships between phreatophytes and groundwater along riparian margins in Western Kansas. They tracked the changes in vegetation with changing water use (and therefore changing groundwater levels) over a number of years (using past and present aerial images) They found that in areas with a dense tree population (>10% tree cover), the average depth to water ranged from 0.24-1.4 m. In areas with moderate tree density (5-10% tree cover), the average depth to water ranged from 2.1-19 m. In areas with low tree density (<5% tree cover), the average depth to water ranged from 11-28 m. Areas where phreatophyte trees were present along with deep groundwater tables were found to have local perched water sources in the alluvium. The results indicated that as the water table declines, trees will be distributed nearer to the river (assuming the groundwater is shallow enough for phreatophytes growth). Mapping phreatophytes could potentially provide insights into the channel densities which receive groundwater in a basin.

3.8 Groundwater abstractions

The effective management of water resources where exploited aquifers are in hydraulic connection with river systems requires an understanding of the response of hydrological systems to groundwater abstraction. The sustainable development of groundwater depends in part on determining the potential impact of that development on the groundwater contribution to rivers, lakes, wetlands and estuaries. Development of groundwater extraction boreholes introduces a new form of discharge and the system must respond by moving to a new equilibrium. This state is accomplished, either by a decrease in natural groundwater discharge, an increase in groundwater recharge, or a combination of the two (Sophocleous, 2002), until the change in these fluxes balances the pumping losses. Where groundwater levels are lowered (through groundwater abstractions) in the vicinity of a river, then stream flow losses will occur through either reduced baseflow (captured discharge) or induced seepage (induced recharge) (Braaten and Gates, 2003). As the cone of depression must expand and intersect the stream before induced seepage occurs, there is a time lag between groundwater pumping and the maximisation of stream flow depletion.

The setting of a sustainably available quantity of groundwater is a contentious issue. Cook (2003) used an analytical solution to simulate changes in aquifer discharge arising from changes in groundwater recharge. They compared situations where the groundwater extraction is evenly distributed across the aquifer, where all groundwater extraction occurs within 1 km of the river and where groundwater extraction is excluded from a 2 km buffer zone adjacent to the river. While all of the scenarios ultimately have the same impact on the river, the early time impact could be controlled by regulation of well location. In this case, the use of a 2 km buffer zone results in a reduced impact on the river of more than 100 years. Scaling analysis shows that large response times are favoured by large systems with high storativity and low hydraulic conductivity or transmissivity (Sophocleous, 2012).

4. SIMULATING THE WATER BALANCE

Computer based models attempt to represent conceptual understanding in a catchment with appropriate equations and with parameters which allow for flexibility in adapting the model to different but conceptually similar catchments (Wagener and Gupta, 2005). Various types of models represent these processes over a range of different spatial, temporal and conceptual scales. Modelling approaches in surface and groundwater interaction studies vary with each model differing in terms of the degree to which physical processes are represented, the data requirements and associated data/computational costs, the model capabilities and the form of model outputs (Ivkovic, 2009). Available models range from parsimonious-lumped to complex distributed physically based representations (Yadav *et al*., 2007). Much discussion surrounds the most appropriate type of model to use in an investigation, with each type of modelling approach having strengths and limitations. Often there is no 'best' model for all applications and the most appropriate model will depend on the intended use and data availability. More complex models aim to represent hydrological processes more explicitly but this introduces further complications as there seems to come a point at which added complexity in a model's structure is not matched by the ability to quantify the model parameters realistically given typically available data resources (Hughes, 2010b). Modelling studies can contribute to the understanding of hydrological processes at different scales, but only if uncertainties related to the quality of the model input information can be overcome. The availability and accuracy of the data utilised by models, has not kept pace with recent model developments. The choice, therefore, lies between using either a sophisticated model with inadequate input data or a less complex model, based upon a simpler conceptualisation of 'known reality', for which there are less data requirements (Xu and Singh, 1998).

Beven (2000) gives two examples in the United Kingdom of situations where planned developments were rejected because the simulations of groundwater flows differed drastically between modellers and highlights the fact that model results cannot be validated. The author predicted that the assumptions and predictions of models will come under increasing scrutiny, with modellers increasingly having to defend their predictions. While new philosophies and theories on modelling and results validation have been published (Beven, 2002; Gupta *et al*., 2008), there is still much controversy over the most appropriate type of model available and in many countries no real utilisation of new validation theories and philosophies. In many cases, models are still being validated and compared using sparse and uncertain datasets. Beven (2002) highlighted the need for a better philosophy toward modelling than just a more explicit representation of reality and argues that the true level of uncertainty in model predictions is not widely appreciated. This is because a model will only ever be an approximation of complex processes, that each place is unique and this uniqueness is essentially unknowable. Therefore there is always the problem of equifinality. Beven (2002) suggested that the range of behavioural models would be best represented in terms of an uncertain landscape to model space mapping. In this way the range of predictive uncertainty associated with the set of behavioural models would be explored. This set of behavioural models would then be constrained by observations which are clearly essential for refining understanding of the response of particular places. This type of philosophy is shared by many authors (Pappenberger and Beven 2006, Gupta *et al*., 2008; Demargne *et al*., 2010) and was a key focus of the International Association of Hydrological Science's (IAHS) Prediction in Ungauged Basins (PUB) initiative (Sivapalan *et al*., 2003). The hydrological decade focused research towards fundamentally changing the field of hydrological science from calibration based modelling to new approaches focused on fundamental understanding of hydrological systems.

The development or modification of models for the purpose of simulating surface and groundwater interactions has produced diverse models, each with their own interpretation and representation of the processes occurring within an interaction environment. As many of these processes are poorly understood, they have been represented in various ways. There are, however, fundamental processes which constitute the basic conceptual understanding of most interaction environments, which need to be incorporated into any integrated model hoping to realistically simulate surface and groundwater interactions. These processes include:

- Recharge
- Aquifer storage
- Aquifer drainage to rivers (groundwater baseflow)
- River drainage to aquifers (transmission losses)
- Evapotranspiration losses from aquifers
- Groundwater flow through an aquifer

In addition to the incorporation of these essential processes, DWAF (2004a) describe the essential components of any technique developed for application in integrated water resource management, these include:

- Practical and operational
- Simple enough to allow large scale basin-wide applications
- Take into account readily available data

- Acceptable and understandable by the groundwater and surface water communities
- Produce reliable outputs for certain specified conditions
- Be able to simulate conceptual processes at an adequate scale

4.1 Prediction across scales

Understanding hydrological systems and the effects of spatial and temporal heterogeneity on surface and groundwater requires integrating across units that differ in size, shape and arrangement. As data never completely represent the hydrological environment, heterogeneity and scale effects are always significant. It remains the case that surface and groundwater interaction data at large scales are essential for water resource management. There are still relatively few studies at larger spatial scales which investigate the nature of interacting controls on baseflow generation (Shaman *et al*., 2004). Studies that are carried out at large scales (Tetzlaff and Soulsby, 2008) suffer from a lack of tools which allow processes to be extrapolated from point scales to larger catchment scales and that can account for increasing anthropogenic impacts that mask natural variability. The development of guiding principles for combining data and models at different spatial and temporal scales and extrapolating information between scales remains a challenge. Blöschl *et al*. (2007) examined spatial scale issues in terms of assessing the impacts of land use and climate change. The authors identified two main approaches (detailed in Sivapalan *et al*., 2003). The first is an upward approach which consists of model cascades with each model representing a sub-process (rainfall, evapotranspiration, groundwater discharge etc.). While this approach can identify causal controls, the results are a reflection of the assumptions involved. The second method is a downward approach which includes trend analyses of long runoff data series and paired catchment studies. While these can capture the summary effects of all the controls, it is difficult to identify the causality.

The temporal non-linearities involved in modelling a collection of processes are significant, as different processes operate at different time scales. Integrated models have to represent many different time scale levels so there is no optimal spatial or temporal scale at which a model should be applied, rather this should be dictated by the purpose of the modelling exercise. Rodriguez-Iturbe and Gupta (1983) attempted to determine the optimum scale at which to carry out an integrated investigation and define the problem in terms of an appropriate level of conceptualisation of the hydrological processes, which is compatible with the phenomena observed over the catchment as a whole.

Complications like less detailed large scale information, variability at all scales (e.g. fractured rocks) and unknown governing equations at some of the scales, further complicate the process. The difficulty in carrying out broad scale assessments and of upscaling small scale data hinders the ability to address regional processes. Remote sensing and GIS has aided in providing information at regional scales but there remains significant uncertainty associated with these methods. Tetzlaff and Soulsby (2008) endorse the use of tracers in conjunction with hydrometric analysis in larger scale investigations. They claim that geochemical and isotopic tracers can enhance insights into hydrological functioning, reflecting the integration of smaller-scale hydrological processes that underpin emergent properties of catchment response at larger scales. Although there have been publications that address scale issues (Turner *et al*., 1989; Sivapalan, 2003; Blöschl *et al*., 2007), many questions remain. Two major issues include how one can upscale local information on soils, vegetation, groundwater and surface and groundwater interactions to the large scale, and whether catchment scale modelling of surface and groundwater interactions can yield reliable results.

4.2 Conceptual rainfall-runoff modelling approaches

Conceptual model equations are based upon a simple interpretation of the physical processes acting upon the inputs and outputs of a system. Conceptual models often, but not always, employ a lumped approach where the individual storage components within the catchment are represented as a single unit with parameters and model outputs representing average values over the catchment area. Each of these storage compartments generally consist of empirical models, so that many of the issues associated with empirical models (lack of explanatory depth) can translate to conceptual models. However, the configuration and relationships between the storages can provide additional insights into the physical processes governing the system behavior (Ivkovic, 2006). In some cases, the models are applied in a semi-distributed manner by splitting a catchment into linked sub-catchments (Schumann, 1993; Li *et al*., 2010; Hughes *et al*., 2013). Reliable integrated models can only be developed with appreciable understanding of the conceptual and physical mechanism of surface and groundwater interaction in space and time.

Conceptual models have been used for many years (e.g. Hornberger *et al*., 1985; Le Moine *et al*., 2008; Hughes, 2013) to successfully improve understanding of the real world and make predictions based on this understanding. Less complicated, parsimonious model structures were investigated as a result of the limitations of more complex models in terms of the uncertainty of model parameters. These ‘simpler’ models represent only those response modes that are identifiable from the available data, although care should be taken to ensure the model does not omit hydrological processes essential for a particular problem (Wagener *et al*., 2001). Proponents of simple or parsimonious modelling approaches (Perrin *et al*., 2001; Ivkovic *et al*., 2009) argue that they are relatively easy to use at larger scales, there are lower constraints on data and time requirements to parameterise and there is a relative reduction in the uncertainties associated with model validation when compared with those of over-parameterised models. Hughes (2010b), however, argues that the inevitable lumping of processes in simple models means that parameters have little physical meaning, are just mathematical constants and are difficult to extrapolate to ungauged catchments. In addition, the author questioned whether it was possible to adequately assess whether highly simplified models are simulating processes 'for the right reasons' (Kirchner, 2006). In other words, are the model outputs being attained through the simulation of the processes in a realistic manner? In the past, parameter values were typically obtained through calibration against observed data (where it existed), such as stream discharge. With the focus of hydrological research changing toward improved fundamental understanding, issues such as parameter identifiability and non-uniqueness are being addressed (Hughes, 2010b)

Although simple, non-physically based models offer many advantages, they are not detailed estimation tools due to their broad-brush, largely volumetric assessments. If a high level of spatial detail is required from a prediction, a physically based model is better suited for the problem. Simple, non-physically based models are not designed to determine details like optimum borehole location. However, when simpler predictions such as the effects of large scale groundwater abstraction on stream flows are required, less complex and conceptual lumped models have been shown to be as equally reliable as physically based models (Yadav *et al*., 2007). Other constraints of simple models include their inability to include highly variable (temporally and/or spatially) recharge (Wright and Xu, 2000; DWAF, 2004b). Further criticisms were outlined by DWAF (2006) and include many of the common assumptions within many of the model structures such as; isotropic aquifers, aquifers of uniform thickness, transmissivity being independent of head, inability to simulate perched aquifers, water is taken immediately from storage, no well losses, horizontal groundwater flow and stream levels unaffected by pumping. The question is then, how important are these processes in a large scale model (developed for the purposes of large scale IWRM), and what level of detail is necessary?

4.3 Numerical groundwater modelling approaches

These physically based models are based upon mathematical equations that describe fundamental physical processes. They are distributed models which operate over a large number of elements or grid squares (Ivkovic, 2006). Physically based models have a logical structure similar to the real world system (Xu and Singh, 1998). They can incorporate known aquifer geometry, parameter values and boundaries along with climatological, topographic and hydrological data into a well-developed model to make the best estimations possible regarding all aspects of groundwater flow. The advantage of numerical groundwater models is that they can model groundwater flow in two or three dimensions which enables them to readily incorporate spatially heterogeneous and temporally variable information and help predict the effects of local impacts on ecologically sensitive areas (Levy and Xu, 2012). Numerical models of surface-groundwater interaction combine numerical solutions of equations for surface water and groundwater flow mostly using Darcy's Law to model vertical exchange between the riverbed and the aquifer. The models are often used to evaluate the accuracy of simplified analytical solutions (DWAF, 2006).

The most common groundwater numerical methods are the Finite Element (FE) and Finite Difference (FD) methods. The FD approach is based on a rectilinear mesh whereas the FE approach is more flexible in allowing a spatial discretisation that can fit the geometry of the flow problem (Hiscock, 2005). Finite element models describe the distribution of heads, hydraulic conductivities and storage properties throughout the system using the Boussinesq Equation for unconfined aquifers. A common example of a FE model is the Finite Element Sub-surface Flow and Transport Simulation System (FEFLOW: Diersch, 2005). The finite difference method defines the basic Boussinesq Equation in finite difference form and then solves the resultant matrix using iteration techniques. The most common finite difference model is the USGS code called MODFLOW (MacDonald and Harbaugh, 1988). The Systeme Hydrologique Europeen (SHE: Abbot *et al*., 1986) model is another popular finite difference model.

Numerical groundwater models are becoming ever more multifaceted with the desire to develop more physically realistic representations of the dynamics of natural systems (Gupta *et al*., 2008). While these models can be very useful in representing the physical processes within a catchment (given data limitations and the validity of the structural assumptions), the risk is that they become over-parameterised (Beven, 2001). This increases the chance that there may be more than one set of parameter values that can give equally acceptable predictions of the observed data available. This equifinality or non-uniqueness is largely due to a models ability to simulate different processes, all with the same resulting output and because we do not yet have adequate measurement techniques to reliably define sub-surface processes (Beven, 2000). Hughes (2010b) argues that the concept of equifinality is naturally present in hydrological systems and its presence in a model should be seen as a benefit, as long as there is enough knowledge of a specific catchment to resolve the equifinality. For example, baseflow can be generated via different processes in isolation or combination and this will vary from catchment to catchment. Therefore it is important that a model has an ability to represent this variation. However, if we don't have enough information about a particular catchment, which is a common situation, the equifinality in a model becomes difficult to resolve and other validation methods such as uncertainty estimation must be relied upon. Physically based models which are simulating a large number of processes together with the associated parameters, risk having a large amount of uncertainty associated with the model inputs, which can be translated through to the model outputs (Ivcovic, 2006).

Beven (1989) argued that there are flaws in the application of physically based models and the frequent confidence in and lack of critical thinking regarding the outputs from the models is misplaced. He argued that the problems result from limitations of the model equations relative to heterogeneous reality, the lack of a theory of sub-grid scale integration, practical constraints on solution methodologies and problems of dimensionality in parameter calibration. He suggests that applications of physically based models can be compared to lumped conceptual models at the grid scale. Although he agrees that physically based models can be very useful tools, their limitations and assumptions must be clearly understood before application. Other common issues with these complex models include their extensive data requirements which are not readily available and expensive to gather. Where data are available, this is most often point data from localised areas which is upscaled and aggregated to the scale at which the model algorithms apply. This introduces additional uncertainty in the model outcomes. In addition, most physically based models are not suitable to model interactions at anything larger than channel reach scale since they are overly reliant on highly heterogeneous parameters of recharge and hydraulic conductivity that are difficult to quantify, as well as selected cell size. The MIKE SHE (DHI, 2001) model however, seems to have resolved the scale issue and is applicable for any catchment where there is a good quality data set irrespective of catchment size. The trade-offs between the modelling approaches tend to be that of parsimony versus complexity, the associated predictive versus explanatory powers, and the data/computational requirements versus the costs (Ivkovic, 2006).

4.4 Selection of a model

Because of the complexity of fully distributed, physics-based catchment models, most hydrological models utilised in real world applications are of the conceptual type. By comparison, most hydrogeological models are of the distributed and physically based type (Levy and Xu, 2012). The selection of a model primarily resolves around the purpose of the modelling exercise. Other considerations include the availability of time, availability of data and human resources (training and experience) (Ivkovic *et al*., 2009). While there is a large amount of controversy associated with the use of different approaches to modelling, these are often resolved when the primary purpose of the exercise is stated explicitly (Xu and Singh, 1998). Whilst each type of model has its advantages and disadvantages, it is important to see the different model approaches as complementary, and not competing, with each approach providing different insights into a system. There are many types of models as well as combinations of model types available.

During the model selection process, it is important to ensure that the major processes that occur in the region of application are included in the model functionality; other considerations include the availability of data and the scale of application. For the purposes of water resource assessment a model needs to be able to handle large scale applications which can then be used to identify regions where spatial and temporal scale differences require more detailed modelling or data collection. In South Africa the major concerns include a lack of data, highly heterogeneous environments and practitioners and scientists who are unable to agree on the way forward in terms of integrated water resources estimation and management. While ideally in such a heterogeneous environment, a more physically based model should be utilised, the data deficiency renders the results from detailed models too uncertain. In addition the large size of the country does not lend itself to detailed national characterisations. Rather a “compromise” could be in the form of a moderately detailed conceptual model which encompasses most of the major processes that occur in South Africa which could be used to characterise the surface and groundwater interactions nationally, while assisting in identifying areas which require more detailed modelling. The difference in model structures and scales between surface and groundwater models means that integration for regional scale modelling will always be very difficult (Hughes *et al.*, 2010a). Unfortunately, despite the urgent need for integrated surface water and groundwater resource management, integrated models applicable to the real world are scarce (DWAF, 2004a).

5. MODELLING SURFACE WATER AND GROUNDWATER INTERACTIONS

The objectives of the modelling component of this study were focused on tests of the application of the SPATSIM version of the Pitman model that includes surface-groundwater interaction components added by Hughes (2004). The motivation for this approach was based on the perceived need to integrate surface and groundwater resource assessments and on the understanding that many (if not most) of these assessments are likely to continue to operate at the quaternary catchment scale for a long time into the future. This type of modelling approach (which includes a similar version of the Pitman model groundwater components developed by Sami) has met with a large amount of criticism from the groundwater community within South Africa and yet it has not been thoroughly tested by any other group. The Berg WAAS (DWA, 2008) study included a very brief assessment of the Sami version of the model and concluded that it did not work. However, no real details of the assessment were provided and therefore it was difficult to see how the conclusion was reached. Other verbal assessments (at meetings and conferences) have reached similar conclusions but have failed to provide in-depth material and explanations. As far as the authors of this report are aware, there is no peer-reviewed assessment available.

The objectives of this section of the report are to provide examples of the application of the model in different hydrogeological settings and to try and answer what the project team consider to be the most critical questions about the value and usefulness of the model. However, before these questions are raised it is important to re-iterate the purpose of the model, especially given that some of the criticisms appear to be based on using the model for purposes for which it was not designed.

- The model is an extension of the Pitman monthly time-step model that was originally designed for simulating catchment scale (~100 to 10 000 km^2) hydrological responses for water resources assessment purposes.
- It was therefore always important that the level of detail of the new components of the model should be reasonably consistent with the level of detail of the original model.
- The model additions were designed to assist with understanding and estimating catchment scale processes of surface and groundwater interactions that include:
    - Recharge.
    - Unsaturated zone interflow processes.
    - Fluctuations in groundwater storage and the effects on outflow dynamics.
    - Groundwater drainage to river channels and contributions to stream flow.

    - Groundwater losses to evapotranspiration, specifically in channel riparian areas.
    - Transmission losses from stream flow to groundwater when groundwater levels are below the river channel.
    - Groundwater abstraction impacts at the catchment scale.
- The model was never designed for, or intended for use in, small-scale localised studies of groundwater abstraction, nor for the design of borehole locations or abstraction management.
- It was also never designed to solve other localised surface-groundwater interaction problems such as contaminant transport assessments.

The most critical questions that have been identified by the project team over several years of application of the model are:

- Is the model structure generally appropriate for the purpose for which it was designed?
- Are there certain hydrogeological settings for which the current model structure is not appropriate? An example might be situations where the surface and groundwater catchments for not coincide and where a substantial proportion of the recharge over one catchment may reach the surface in an adjacent, but not downstream, catchment.
- Are the parameters conceptually identifiable and meaningful in terms of known hydrogeological principles (at the appropriate scale)?
- Can the parameters be realistically quantified either directly (*a priori*, without calibration) or after calibration against some observed data or other available estimates?
- How sensitive are the model outputs to different parameter values and therefore which should be concentrated on either in a calibration approach or through *a priori* estimation?
- How can the inevitable equifinalities in a relatively highly parameterised model be resolved and is this an important issue from a water resources (including both surface and groundwater) assessment perspective?
- Finally, how much confidence (or uncertainty) can we express in the results obtained under typical situations of data availability?

5.1 Brief description of the surface-groundwater components of the Pitman model

The model has been described in the literature many times (Hughes, 2004; Hughes et al., 2006) but the relevant parts of the model are briefly described here to avoid readers having to access additional sources.

The ***interflow and groundwater recharge functions*** in the model are very similar and depend upon a non-linear relationship with storage in the unsaturated zone (also referred to as soil moisture in some descriptions of the model):

$$Qint = FT * (S/ST)^{POW} \qquad \text{Equation 5.1}$$

$$GWR = GW * ((S - SL)/(ST-SL))^{GPOW} \qquad \text{Equation 5.2}$$

Where ST is the maximum unsaturated zone storage (mm), S is the current level of storage (mm), SL is the minimum storage (mm) for recharge, FT and GW are the maximum rates (mm month^{-1}) of interflow and recharge, respectively, and POW and GPOW are the two power parameters. The recharge depth (GWR mm) is added directly to the groundwater storage in the model. The implication is that if the model is to represent a lateral component of flow for water moving down through the total unsaturated zone then it must be included as part of Qint (interflow depth in mm). This further implies that ST and S represent not only the near surface soil moisture storage, but also the deeper unsaturated zone storage in the rocks above the regional water table (Kapangaziwiri and Hughes, 2008). The interpretation of FT and ST in this version of the model can therefore be different to the original interpretation in areas where substantial lateral flows in the deeper unsaturated zone are expected.

The groundwater storage is represented by simple geometry based on a number of representative slope elements. The number of slope elements is determined from the catchment area and a drainage density parameter (DDENS). This approach allows the total sub-basin to be divided up into an even number of equally sized slope elements (as shown in Figure 5.1) for which the slope width and channel length can be defined. The groundwater table is represented by two slope elements which can have different gradients depending on the other components of the water balance (discussed below). The lower end of the near-channel slope element is fixed at the channel and the near-channel slope element is assumed to represent 40% of the total slope element width. Any increment of recharge depth is translated into a volume of storage by the area of the slope element and the storativity parameter (S). It is therefore trivial geometry to determine the height of the junction between the two slope elements and the height of the upper end of the far slope element and therefore their gradients at any time step in the model.

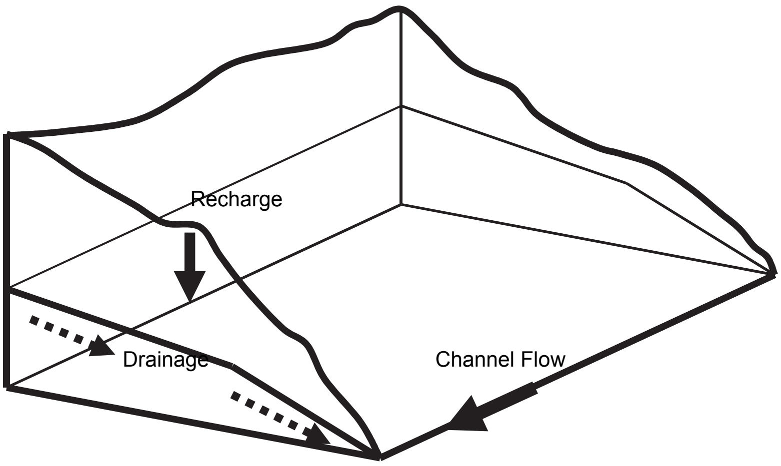

Figure 5.1 Simplified geometric representation of a single slope element as represented in the model.

Figure 5.2 illustrates the possible conditions that can occur with the left hand side diagrams showing the model situation and the right hand side diagrams showing what the model is designed to represent in reality. Figure 5.2A shows a typical situation of a gaining river channel with both slope elements being positive resulting in ***drainage from the groundwater to the channel***. The drainage calculation is based on a transmissivity parameter (T in m^2 d^{-1}) the gradient of the slope element and the total channel length (Figure 5.1). In Figure 5.2A drainage can occur from the far slope element to the near and from the near slope element to the channel. Figure 5.2B illustrates a situation where the groundwater is below the channel, which is represented in the model as a negative near channel slope element. Under these conditions the model can simulate flow from the channel to the groundwater. Figure 5.2C illustrates a situation where the far slope element has been drawn down through abstractions at a point distant from the channel.

The reason why the near channel slope element can become negative is because the model also allows for two other outputs from the groundwater store. The first of these is ***evapotranspiration losses*** from the riparian strip (which have to be satisfied before GW can contribute to stream flow), while the second is ***downstream flow*** to the next sub-basin. The first is determined from the potential evapotranspiration input into the model and a riparian strip factor (RIP, as a % of the total slope element width). The second is based on the T parameter, a downstream GW flow gradient parameter (GWSlope) and the width of the slope element that comes from the basic geometry defined by DDENS and catchment area. Both downstream flow and riparian evapotranspiration losses are reduced as the gradient of the near channel slope element becomes increasingly negative and stops when it reaches the gradient that is equivalent to a parameter defining the rest water level (m).

Determining *a priori* a suitable starting value for the two slope element gradients is not really possible and therefore the model is run twice. In the first run the starting gradients are set to the GWSlope parameter, while the starting values for the second run use the end values of the first simulation. Thus in arid areas the second run of the model is likely to start with a negative near channel gradient.

Channel transmission losses are based on highly uncertain model structural formulations in the absence of sufficient field data to determine more appropriate approaches. Transmission losses are assumed to occur from both incremental flow generated within a single sub-basin, as well as from upstream accumulations of flow in downstream sub-basins. A single parameter is used to define the maximum depth of runoff that can be lost (TLGMax in mm) and the actual losses are based on two power relationships using; 1) the current months discharge relative to TLGMax and 2) the near channel gradient (which must be negative). Further details are available in some of the publications that refer to the model (Hughes, 2004; Hughes et al., 2006).

The model also allows for independent **groundwater abstractions** from both slope elements (two parameters of annual water use plus seasonal distributions).

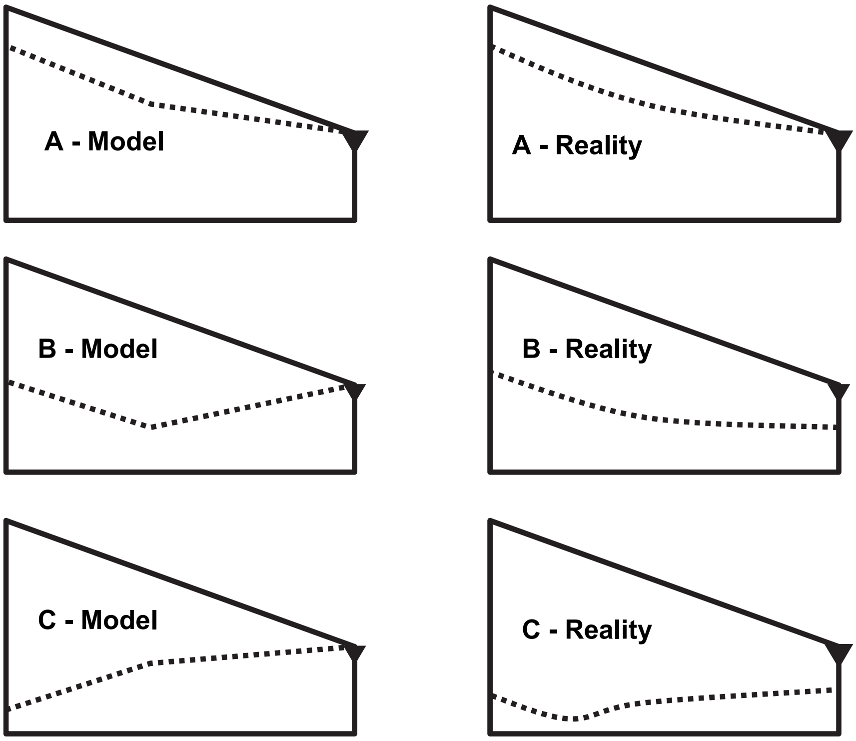

Figure 5.2 Modelled versus real groundwater conditions in a single hillslope element. Dashed lines represent the groundwater levels, the solid upper line represents the surface and the solid triangle represents the river channel. See text for explanation.

5.2 Published model evaluations

The model has already been quite extensively tested in various settings and much of this work has already been subject to peer review (unlike some of the criticisms that have been directed at the model in meeting presentations or internal reports). Some of these are summarised below:

- Hughes (2004): The original presentation of the model.
- Hughes et al. (2006): An extensive application of the model based on calibration in the Okavango River.
- Hughes and Kapangaziwiri (2007), Kapangaziwiri and Hughes (2008): Discussions of the conceptual basis for the parameters of the model and how these could be estimated from physical basin property data.
- Hughes (2009): An application of the model to the Seekoei River for the purposes of environmental flow determinations in ephemeral rivers. Model outputs were confirmed by field observations of groundwater and springs (albeit limited data).

- Hughes et al. (2010a): An assessment of the model in an arid area using the groundwater abstraction parameters. This paper was largely focused on sensitivity tests of some of the parameters in the absence of reliable field data.
- Hughes (2010a): A presentation of evidence for unsaturated fracture flow processes that are a potentially important part of the model conceptualisation.
- Hughes and Mantel (2010a): A discussion of some of the uncertainties in the application of the model to some catchments in South Africa, including the groundwater components.
- Hughes et al. (2010b); Tshimanga et al. (2011): Application of the model to large basins in sub-Saharan Africa (Zambezi and Congo Rivers).
- Kapangaziwiri et al. (2011): A further discussion of uncertainties in the use of the model and the estimation of parameter values, mainly focussed on the low flow estimations and including the groundwater components.
- Hughes et al. (2013): An investigation of spatial scale effects on the determination of the model parameter values with a strong focus on the groundwater parameters.
- Hughes (2013): An overall review of the development and application of the Pitman model.
- Tanner and Hughes (2013): Additional examples of the application of the model including some uncertainty analyses of the groundwater components based on the uncertainties in available data and conceptualisations of surface-groundwater interactions.
- Hughes (2014): A paper that looks at temporal variability in climate inputs and non-stationarity in land use impacts and the effects on calibrated model parameter values (including the groundwater parameters).

It is apparent therefore that the application of the model in a many different climate and hydrogeological settings has been widely published and peer-reviewed. Unfortunately, some of the reviews of the model that have formed parts of other reports (e.g. Allwright et al., 2013) in South Africa have been based solely on the original Hughes (2004) paper and have ignored the later enhancements to and tests of the model. The following sections of this report offer some further site-specific, or component specific, assessments of the model that were designed to answer the questions presented at the start of this section.

5.3 Uncertainty analysis

The companion report to this document (Hughes et al., 2014a) details the developments that have been made with uncertainty approaches applied to the Pitman model. While most of the issues referred to within Hughes et al. (2014a) are associated with the practical implementation of uncertainty analysis, they are also relevant to the use of models for assisting with the understanding of processes through hypothesis testing (Beven, 2012). Essentially, this type of hypothesis testing involves making some assumptions about how the model can represent real processes and testing these using comparisons of the model results with observed information, or known limits to some hydrological processes. This type of approach is facilitated by the use of uncertainty versions of the model where many runs of the model can be made with different parameter sets and the total output examined for parameter combinations that give behavioural results. Behavioural results are those that can be considered to generate outputs that conform to some known response characteristics (e.g. observed stream flow time series or observed groundwater level variations).

The two components of the project have therefore complemented each other and the uncertainty method developments have certainly facilitated the exploration of likely parameter sets that generate behavioural surface water-groundwater interactions. However, as will become evident throughout the following sections of this report, it is not always possible to resolve some of the uncertainties in our understanding or in the input data that are used to force hydrological models. Thus, there are situations where we are not able to obtain 'good' model results, but we are not always able to isolate the reasons for this. The uncertainty versions of the model allow us to identify which parameter combinations will generate the best results, but where these are still poor it is generally not possible (without more information) to attribute this to the contributions of inadequate input data (rainfall, etc.), inadequate or inaccurate observed evaluation data or inadequacies in the model structure. The latter could be further broken down into poor representation of processes and linkages between processes, or inadequate representation of spatial and/or temporal scale variations. These issues are not confined to the Pitman model and apply equally to all environmental models. The interpretation of any models output therefore always contains an element of subjectivity that is very dependent upon the experience of the interpreter and their view of the real world.

6. MODEL EVALUATIONS

Some of the model evaluations discussed below are based on comparisons with observed stream flow data. In all cases a group of 5 standard objective functions are used. The first three are the Nash-Sutcliffe coefficient of efficiency using no transformations of the data (CE), natural log transformed data (CE{ln}) and inverse transformed data (CE{Inv}). The other two are the percentage difference in the simulated mean monthly flow (MeanQS) relative to the mean observed flow (MeanQO), using both un-transformed values (Bias) and natural log transformed values (Bias{ln}):

$$\text{Bias} = (\text{MeanQS} - \text{MeanQO}) * 100 / \text{MeanQO} \qquad \text{Equation 6.1}$$

6.1 Assessing the recharge function

As noted in Section 3.3, groundwater recharge can be a complex process that involves different vertical percolation processes in different geological environments that can include macropore or fracture flow and matrix flow. The critical issue is that the model function is based on a non-linear relationship between recharge and unsaturated zone storage, while some authors have suggested that recharge will be more dependent on the rainfall characteristics rather than the storage state of the unsaturated zone. However, it has also been noted that variations in near-surface infiltration can be buffered by the vertical percolations processes in the deeper parts of the unsaturated zone. Arguably, this buffering effect will be related to the storage in the unsaturated zone. In addition, modelling at monthly time-steps means that it is impossible to account for the effects of rainfall intensity.

A series of tests were carried out using simulations of the recharge within quaternary catchment K40A (Southern Cape coastal area). The model was initially set up to simulate the observed flows with recharge estimates that are consistent with values given in GRAII (DWAF, 2005). A series of alternative methods of generating recharge were established using only the net rainfall (after surface runoff simulated by the model) with different Scale, Threshold and Power parameters in equation 6.2. The combinations of the parameters were set to give approximately the same mean monthly recharge as simulated by the original model set up.

$$\text{Recharge} = \text{Scale} * (\text{Net rainfall} - \text{Threshold})^{\text{Power}} \qquad \text{Equation 6.2}$$

The resulting recharge estimates are therefore independent of the storage state of the unsaturated zone. The patterns of variation in monthly recharge were represented as cumulative frequency curves and the parameters (GW, GPOW and SL) of the model formulation (Equation 5.2) were then calibrated to try and produce the same pattern frequency patterns. Figure 6.1 illustrates the results and demonstrates that under many different possible rainfall-only estimates (using Equation 6.1) the Pitman model formulation can generate similar frequency patterns. This does not, of course, mean that individual months will have the same recharge values under the two types of estimate.

It should be stressed that this evaluation does not make any assumptions about which approach to recharge estimation is correct (rainfall-based or moisture storage-based), it merely demonstrates that the method used in the model (moisture storage-based) generates very similar frequency patterns of recharge even if the assumption that recharge is controlled only by rainfall is correct. The main reason is that the patterns of variation of unsaturated zone moisture content will closely follow the patterns of rainfall variation.

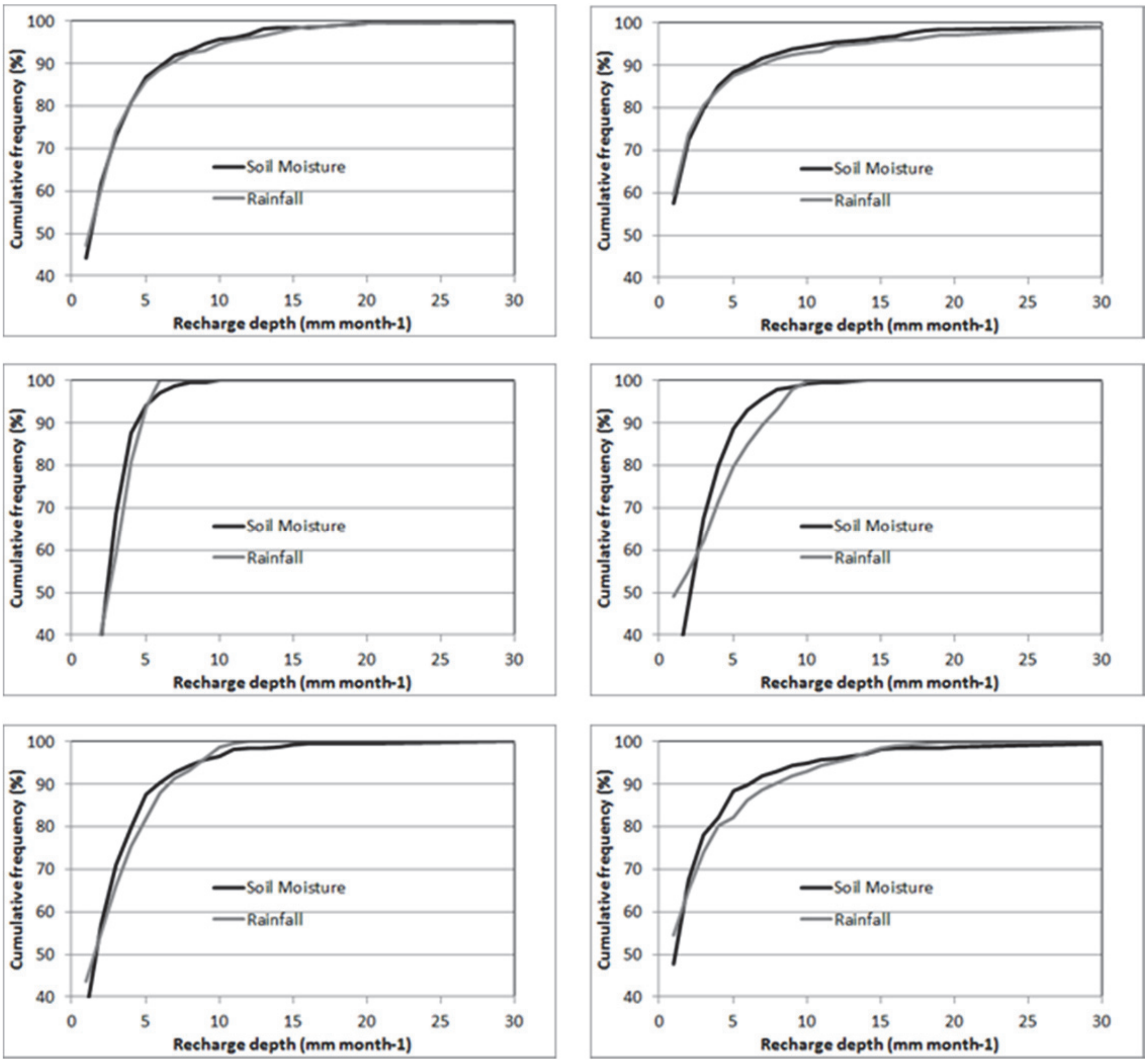

Figure 6.1 Cumulative frequency distributions of recharge simulated using a rainfall-only type model compared with the function in the Pitman model based on the storage state of the unsaturated zone.

6.2 The Ecca River catchment: Unsaturated zone interflow processes

This example was used in Hughes (2010a) and represents an arid area in the Eastern Cape Province with stream flow dominated by very short events following high rainfall. The catchment is characterised by steep slopes and very thin soils except in some alluvial/colluvial areas bordering the downstream parts of the main channel. The geology consists of steeply dipping and highly fractured shales and sandstones. A 9.1 km^2 tributary catchment has been used here to try and identify the source of rare baseflows that occurred during and after two major rainfall events in July and August

1979. Figure 6.2 illustrates that the measured stream flow continued for approximately 110 days after the 2nd rainfall event, despite field observations that little near surface moisture existed some 7 to 10 days after the main rainfall. The prolonged baseflows could not be attributed to groundwater contributions from the regional aquifer as these water levels remained at greater than 15m below the channel bed at all times.

A daily version of the Pitman model (Hughes et al., 2014b) using more or less the same structure as the monthly model has been used to explore possible behavioural parameter values as a guide to likely processes. The previous daily version was slightly modified to include a similar form of the routing equation used in the monthly model (Pitman, 1973) but applied only to the simulated surface runoff. The latter was included to smooth the effect of large daily values of surface runoff and to see if the routing function could account for the prolonged baseflows. The model is run 1 000 times using simple Monte Carlo sampling from uniform parameter distributions. The initial run with FT set to 0, revealed that regardless of the routing parameter value, the baseflows could not be generated. Table 6.1 shows the ranges of the uniform distributions used in the second run, while two of the lines on Figure 6.2 illustrate the full range of the simulated flow depths. Table 6.1 and Figure 6.2 also include a parameter set and one of the best results based on all three objective functions (CE = 0.73, CE (ln) = 0.67, Bias = 4.2%).

The same parameters were applied to a longer period (February 1979 to December 1980) in which there were only two additional runoff events (all < 0.3 mm d^{-1}), while the simulations generated some longer periods of small flows. The implication is that the model structure is too simple to correctly capture the complex surface and sub-surface runoff processes that might include delays in the interflow contributions, re-infiltration of some surface runoff as well as evaporative losses from the channel margins. However, the model has helped to demonstrate the existence of a prolonged interflow component that occurs after heavy rainfall and that cannot be linked to drainage from the regional aquifer nor to drainage from the soil profile. The range of the ST parameter value, as well as the best result are far too high to be attributable to soil moisture storage in a catchment with very thin soils and the conclusion is that these values for storage must represent deeper unsaturated zone storage in the highly fractured shales and sandstones. A consideration of the topography and the orientation of the major bedding planes suggest that approximately 50 to 80 * 10^6 m^3 of unsaturated zone material could contribute to interflow. Based on an ST of 200 mm and with 50 to 70 mm assumed to represent near surface storage (saprolite and soil), an estimated value for the fracture zone storativity is between 0.015 and 0.03. These values do not seem to be unrealistic for highly fractured shales and sandstones, given values reported by a number of different sources.

Table 6.1 Parameters of the daily version of the Pitman model used for the Ecca River example.

Parameter	Uncertainty Range	Best Result
ZMIN (mm d^{-1})	10-30	12.0
ZMAX (mm d^{-1})	150-200	160.0
ST (mm)	150-200	208.0
SL (fraction of ST)	0.01-0.2	0.06
R (-)	0.0 to 0.2	0.1
GW	1.5	1.5
GPOW	5.0	5.0
FT	5-15	9.1
POW	4-10	4.4
K (Routing parameter – h)	5-20	19.0

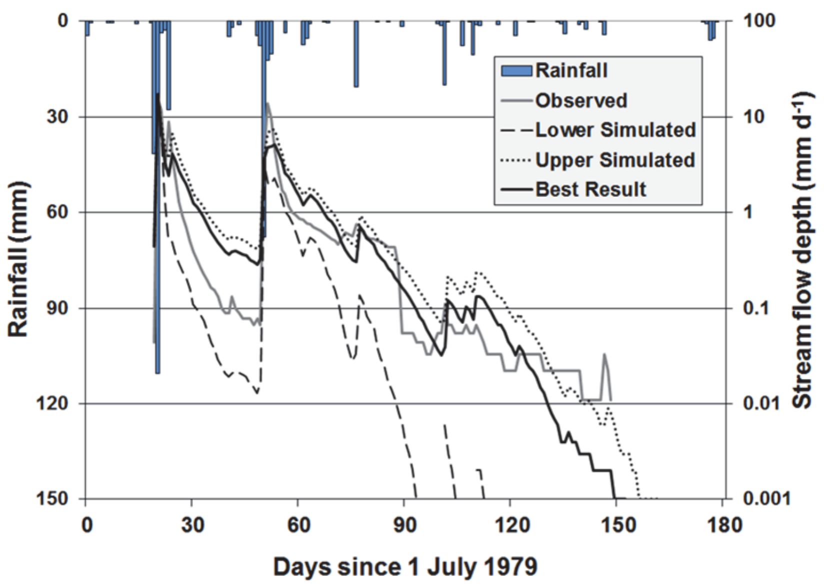

Figure 6.2 Uncertainty analysis results of the application of the daily version of the model on the Ecca catchment.

6.3 The Sabie River catchment: Resolving the source of low flows.

The headwaters of the Sabie River drain the steeply sloping Eastern escarpment of South Africa within Mpumalanga Province. The underlying geology consists of quartzites and shales with a band of dolomite. Figure 6.3 illustrates the daily stream flow response at the gauging station X3H001 (catchment area of 173 km^2) during both wet (1973 to 1978) and dry (1981 to 1986) conditions. The most evident feature of the response is a very high baseflow component during wet years which has higher flow events superimposed upon it. During very dry years (1981 and 1982) this large baseflow is not present and the low flows are relatively constant throughout the year. Hughes (2010a) ascribed this type of response to mixed processes of low flow generation that included a relatively slowly varying groundwater drainage component and a much more seasonal unsaturated zone drainage component that is probably dominated by the dolomitic compartments. The assumption is that the unsaturated zone is highly reactive and that recharge during the wet season of dry years is too low to replenish the storage that would be depleted at the end of the previous dry season.

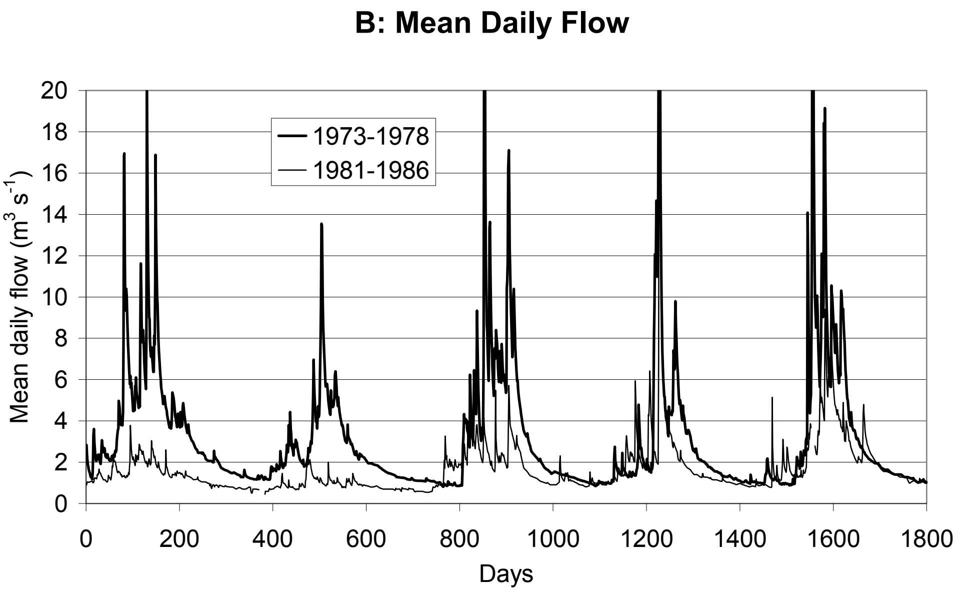

Figure 6.3 Mean daily flow observed time series for the Sabie River at X3H001.

The purpose of the Pitman (monthly) modelling approach was to determine if the model could be used to resolve the source of the low flows; either mostly groundwater drainage or mostly unsaturated zone interflow. The uncertainty version of the model with quite large parameter ranges (generating 10 000 ensembles using independent Monte Carlo sampling of the parameter space) was used together with three different periods of observed data to calculate objective functions and identify the most behavioural parameter sets. Three periods of observed data were used; the total usable record of 1966 to 2005, the dry periods only (1981 to 1984 and 1991 to 1994) and wet periods only (1973 to 1977 and 1996 to 2001). Table 6.2 lists the parameter values and uncertainty ranges that were used in all model runs. Some of the parameter values were kept constant to avoid some of the potential equifinalities and to focus on the dominant sources of low flows. The fixed parameter values were partly based on existing parameter sets, confirmed by some preliminary calibration tests. The catchment was assumed to be covered by 60% plantations and the interception and evapotranspiration parameters were set to reflect these conditions based on previous experience.

Table 6.2 Parameter values and uncertainty ranges used for the Sabie River example

Parameter	Value or range
ZMIN (mm $month^{-1}$)	20
ZMAX (mm $month^{-1}$)	600
ST (mm)	400-800
FT (mm $month^{-1}$)	20-100
POW (-)	2.5
GW (mm $month^{-1}$)	20-100
GPOW (-)	3.5
R (-)	0.5
DDens (-)	0.4
T (m^2 d^{-1})	5-40
S (-)	0.002-0.01
RIP (%)	0.4

The analysis of the results was based on identifying the ensembles that generated Bias and Bias{ln} values of less than ±5% and then ranking these on the basis of the sum of the three CE objective functions (i.e. CE + CE{ln} + CE{Inv}). The most noticeable features of the results was that the sum of the CE values did not change that much within the ensembles that had less than ±5% bias values and that the range of parameter values within these behavioural ensembles was not generally different from the total possible range. This result confirms the well-known existence of equifinality associated with the Pitman model. However, exploring the relationships between parameter combinations suggests some trends amongst the most behavioural ensembles:

- There is a distinct negative relationship between FT and GW and the range of the sum of these parameters in the behavioural ensembles is much lower than within the total 10 000 ensembles.
- The wet period suggests the highest combination of FT and GW, but the resulting parameter sets tend to over-simulate for the total period (Table 6.3).
- T and S are not very identifiable, but there is a generally positive relationship between these two parameters. If higher S values (promoting lower fluctuations in GW storage) are associated with lower T values (slower outflows), these are typically accompanied by higher GW values that will increase the rate of GW storage fluctuation to compensate.

- The highest frequency of behavioural T values are between 10 and 20 $m^2 d^{-1}$, while for S this range lies between 0.006 and 0.009.

The 'best' ensemble results for each set of ensembles, as well as the objective function values measured against the full observed data set (rather than those used to rank the ensembles) are given in Table 6.3 and it is clear that despite the different parameter values, the overall results are similar. The last two columns of Table 6.3 offer two parameter sets that fall within all of the identified behavioural sets, the first of which assumes dominance of interflow, while the second assumes dominance of groundwater recharge. The first is based on the assumption that the dolomitic compartments will be part of the interflow component and that the groundwater will have low storativity and low transmissivity. The second assumes that at least part of the dolomitic storage is part of the groundwater system which consequently has higher recharge, storativity and transmissivity. Figure 6.4 illustrates the model results for these two situations (the first being SC1 and the second being SC2) using seasonal distributions of the mean monthly simulated interflow and groundwater drainage outputs.

Table 6.3 Parameters and objective functions for the 'best' ensembles for each test period compared with the full time series of observed data.

Parameter or Objective function	Period used for ensemble evaluation (calibration)			Alternative parameter sets favouring:	
	Total	Dry	Wet	Interflow	G'Water
ST	669	708	766	750	750
FT	67	68	98	90	50
GW	79	52	97	60	100
FT + GW	146	119	195	150	150
T	13.3	5.4	13.5	10	18
S	0.009	0.007	0.006	0.004	0.008
CE	0.750	0.746	0.738	0.748	0.752
CE{ln}	0.792	0.775	0.770	0.791	0.781
CE{Inv}	0.561	0.561	0.494	0.550	0.543
Sum of CE values	2.103	2.082	2.002	2.089	2.076
Bias (%)	0.69	-3.39	10.93	4.22	-1.77
Bias{ln}	-0.07	-4.16	8.73	2.09	-0.94

It is evident that the overall simulation results are very similar for the two scenarios, but that the source of the total baseflows is very different. Under scenario 1 the mean baseflows are dominated by the interflow component and the seasonal variation in the deeper groundwater response is quite small. Under scenario 2 the groundwater drainage is much greater with a greater amplitude of seasonal variation. This is offset by a smaller interflow response with a lower seasonal variation than under scenario 1. Both scenarios are almost equally able to simulate the low baseflow response in the two extended dry periods of 1981 to 1984 and 1991 to 1994. The overall conclusion is that the model is not able to contribute to distinguishing between unsaturated zone and groundwater contributions to total baseflow in this catchment. Arguably, this could be because the dolomitic compartments that almost certainly represent the majority of the sub-surface storage can be represented in the model as either part of the unsaturated zone or as the real groundwater. Figure 6.4 illustrates that the total baseflow delay is slightly longer for scenario 2 than scenario 1 and this might be reflected in the slightly better low flow results (CE{ln} and CE{Inv}) for scenario 1 than scenario 2, but the differences are small. The dolomitic zones of this catchment have therefore been well represented by both scenarios, but in different ways.

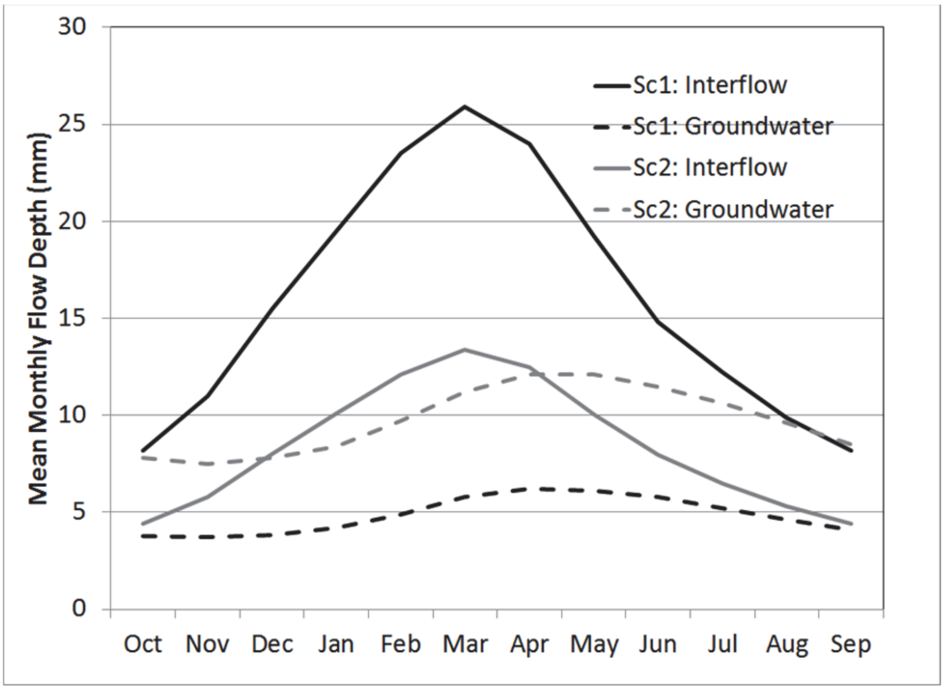

Figure 6.4 Mean monthly simulated components of the total baseflow (interflow and groundwater drainage) for the two example scenarios.

6.4 The Breede River catchment: Groundwater uncertainties.

The headwaters of the Breede River (quaternary catchments H10A to C) have been the subject of several hydrological and geohydrological investigations (DWAF, 2007; Hughes and Mantel, 2010a and b; Hughes et al., 2013, Tanner, 2013a). There are a large number of uncertainties associated with modelling this group of catchments, most of these associated with the complex topography, complex geology and extensive water use for the production of deciduous fruit. The topography consists of steep mountain areas surrounding a broad and relatively flat alluvial/colluvial valley. This contributes to many uncertainties in the rainfall inputs, particularly as the mountain areas are ungauged. The geology is complex with steeply folded and faulted sequences of Witteberg sandstones, Bokkeveld mudstones and Table Mountain sandstones (TMS), linked to the alluvial and colluvial valley fill. DWAF (2007a) refers to the possibility of regional transfers of ground water through the TMS in the lower parts of the catchment. Hughes and Mantel (2010a and b) partly concentrated on the uncertainties in the water use from a large number of farm dams that are evident in the catchments, but DWAF (2007a) identified that some of the water use is from groundwater.

The focus of this study was to try and resolve some of the uncertainties in the surface-groundwater interaction water balance at the very course scale of the total catchment area (657 km^2). In attempting to achieve this objective it was necessary to ignore some of the other uncertainties and to make some assumptions about the way in which the surface hydrology can be simulated. Hughes and Mantel (2010b) referred to the large differences between WR90 and WR2005 rainfall estimates and in this study the higher values of the WR2005 database were used. This could inevitably have some impacts on some of the model parameters and therefore it was necessary to re-calibrate the Pitman model to obtain acceptable values for the non-groundwater related parameters. Previous studies (Hughes and Mantel, 2010a and b) identified the possible influence of return flows from the town of Ceres near the outlet of the catchment and the DWA gauge at H1H003. These were included in the present study as a fixed monthly volume of $27 * 10^3$ m^3 based on assumptions about the population size and percentage of water use that is returned through the waste water treatment works.

The uncertainty version of the Pitman model was set up with quite large ranges in all parameters (including water use) and the resulting 10 000 ensembles were evaluated to determine which of the main surface water parameter values could be fixed, given the use of the WR2005 rainfall data. This mainly involved an examination of the middle to upper end of the flow duration curves. The inevitable, and expected, high degree of equifinality illustrated that the low flow regime could be simulated by a large number of different combinations of model parameters. Fixing some of the surface runoff parameters (ZMIN, ZMAX, ST, FT and POW) does not substantially decrease the uncertainty in the groundwater and water use parameters which were the focus of this investigation. Table 6.4 lists the parameters for which uncertainty was retained in the second run of the model.

Table 6.4 Uncertain parameters in the second run of the model for the Breede River catchment.

Parameter	Uncertainty range	Explanation
GW	10 to 60 mm month^{-1}	Accounts for a large uncertainty in recharge identified by Tanner (2013) and reflected in the estimates used in other studies. There will be some equifinality between GW and GPOW, but the assessment will be based on mean monthly recharge.
GPOW	2.5 to 3.5	
Irrig. Area	15 to 50 km^2	This parameter represents the area of irrigation supported by farm dams and Hughes and Mantel (2010a) estimated the median value to be 67 km^2 (equivalent to 39 * 10^6 m^3 y^{-1}, given the demand depths used in the model). The uncertainty range of this parameter and groundwater abstraction reflects a lack of information about the source of irrigation water and the total volume used (DWAF, 2007a).
GW Abstr.	10 to 30 * 10^6 m^3	
Storativity	0.001 to 0.005	Reflects uncertainty in the storativity of the GW body and that part of the storage may be in the hard rock sequences and part may be within the valley fill.
GW Slope	0.008 to 0.05	Reflects the possibility that some of the recharge to the TMS aquifer could be transferred to adjacent catchments (i.e. mismatch between the surface and GW catchments).
Riparian Strip	0.2 to 2.0%	Reflects uncertainties in the evaporative losses from GW in the near channel riparian zones.

The basis of the evaluation was to first select the ensembles that met certain criteria with respect to adequately simulating the observed flows. This evaluation was largely based on the Nash coefficients of efficiency using no transformation (CE) and with natural log transformations (CE{ln}) using the observed flow data since 1970 to avoid too many possibilities of non-stationarity in the water use. The best values for these objective functions were CE = 0.68 and CE{ln} = 0.76. The criteria for behavioural simulations (relative to observed flows) were therefore set using limits of approximately 10% less than the best values (i.e. CE ≥ 0.61 and CE{ln} ≥ 0.68). The % bias values were largely ignored because of the possibility of errors in the observed data, particularly with respect to high flow measurements. However, it was noted that the range of bias values were far less within the ensemble set selected using the CE values. This first selection produced 559 ensembles that can be considered behavioural with respect to the observed flows. The second selection criteria was based on an assumption that the total water use was approximately ±10% of the median value used by

Hughes and Mantel (2010a), i.e. 39 ± 4 * 10^6 m^3. This reduced the set of assumed behavioural ensembles to 175.

Figure 6.5 illustrates the relationship between groundwater recharge and an index of total losses from groundwater using a combination of the GW slope, Riparian strip (RIP) and GW abstraction parameters. Figure 6.5 illustrates that the range of recharge estimates is reduced quite substantially within the behavioural set of ensembles relative to the total ensemble set and that recharge values of less than 3.5 mm $month^{-1}$ are unlikely. The three recharge values given by the GRAII (DWAF, 2005) study are 2.0, 3.0 and 6.3 mm $month^{-1}$, while DWAF (2007a) assumed a value of 6.1 mm $month^{-1}$. Figure 6.5 suggests that both the lower GRAII values are less than likely and that the higher values are possibly the upper limits of the uncertainty in recharge.

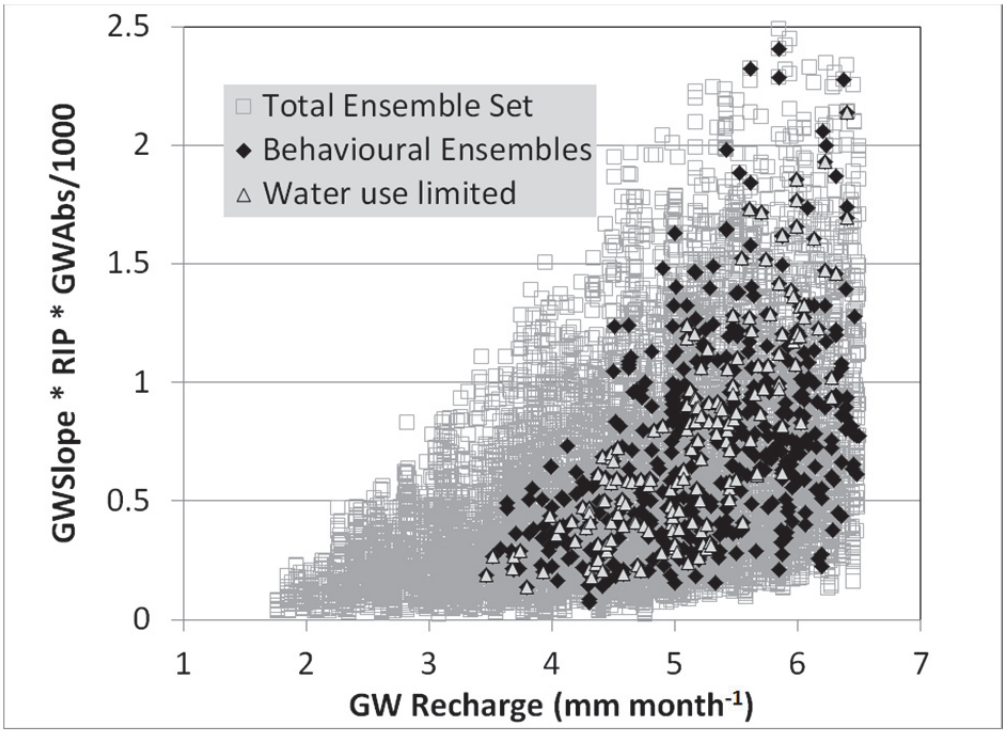

Figure 6.5 Relationship between simulated mean monthly recharge and an index of total losses from groundwater that includes outflows to adjacent catchments, riparian evaporation losses and water abstractions.

Figure 6.6 examines the relationship between the two water uses for the initial behavioural set of ensembles and those considered behavioural with respect to total water use. There is very little difference between these two, but it is perhaps worth noting that there is a greater density of points in the area of relatively high farm dam use and low groundwater abstraction. The DWAF (2007a) estimate of 20.6 * 10^6 m^3 groundwater abstraction is therefore considered to be possible, but less probable than a lower value, on the basis of the results generated by this investigation.

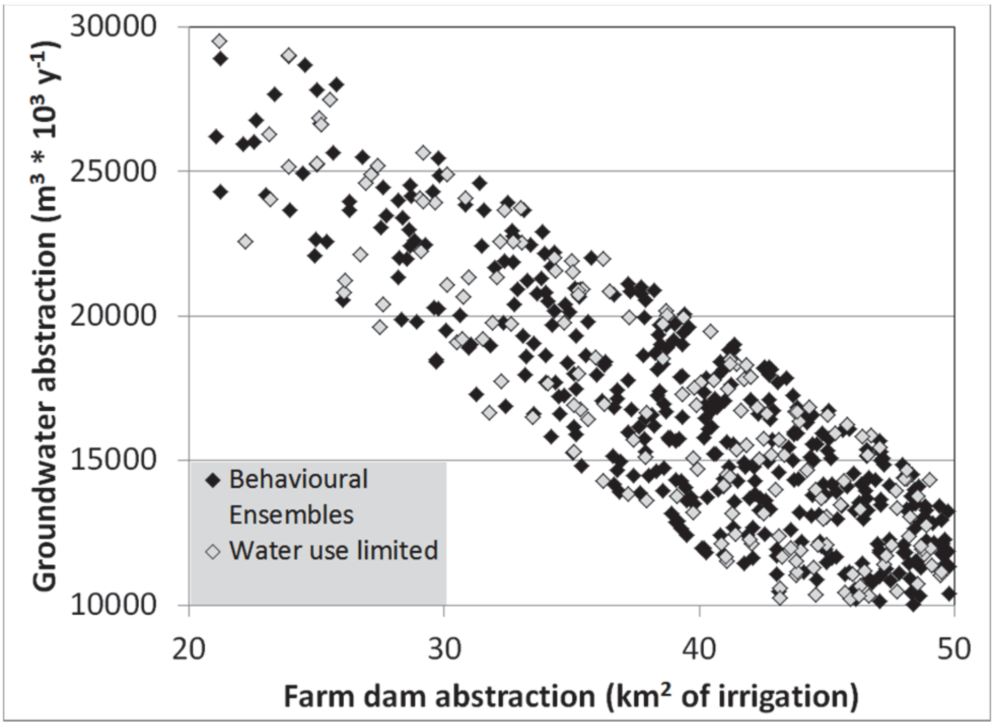

Figure 6.6 Relationship between abstractions from farm dams and from groundwater for the behavioural ensemble set based on comparisons with observed stream flow data (559 ensembles) and the set limited by total water use (175 ensembles).

Figure 6.7 shows the relationship between recharge and groundwater abstraction for the final behavioural ensemble set based on limiting total water use. The left hand side identifies those points that have both low (<0.02) and high (>0.04) values for the GW slope parameter that determines how much groundwater will be transferred to adjacent catchments. Given the other parameters in the model, slopes (or gradient) of 0.02 and 0.04 translate into annual volumes of transfer of 0.33 and 0.66 * 10^6 m^3 y^{-1}, respectively. The DWAF (2007a) estimates ranged from 0.1 to 0.8 * 10^6 m^3 y^{-1}. The results presented here do not really contribute to resolving these uncertainties and the model is largely insensitive to variations in this parameter. This is probably not surprising given the relatively small volumes involved. The right hand side of Figure 6.7 shows the equivalent results based on identifying different values of the riparian strip parameter that determines GW losses to evapotranspiration. The two limiting values of <1.0% and >1.5% used in the diagram represent volumes of approximately 10.9 and 16.4 * 10^6 m^3 y^{-1}, respectively, given no limitation to the supply of water (i.e. no substantial drying out of the groundwater, which is unlikely given the approximately 35 * 10^6 m^3 y^{-1} of recharge). In contrast to the pattern shown for the GW slope parameter, variations in RIP are more clearly identifiable. The higher values of RIP are not considered to be very likely given that this is a highly developed catchment without substantial areas of riparian vegetation. These results point to the likelihood that the recharge range is between 3.5 and 4.5 mm y^{-1} and that the GW abstractions are limited to less than about 18 * 10^6 m^3 y^{-1}, which offers some support for the conclusions that were reached based on the analyses of Figures 6.5 and 6.6.

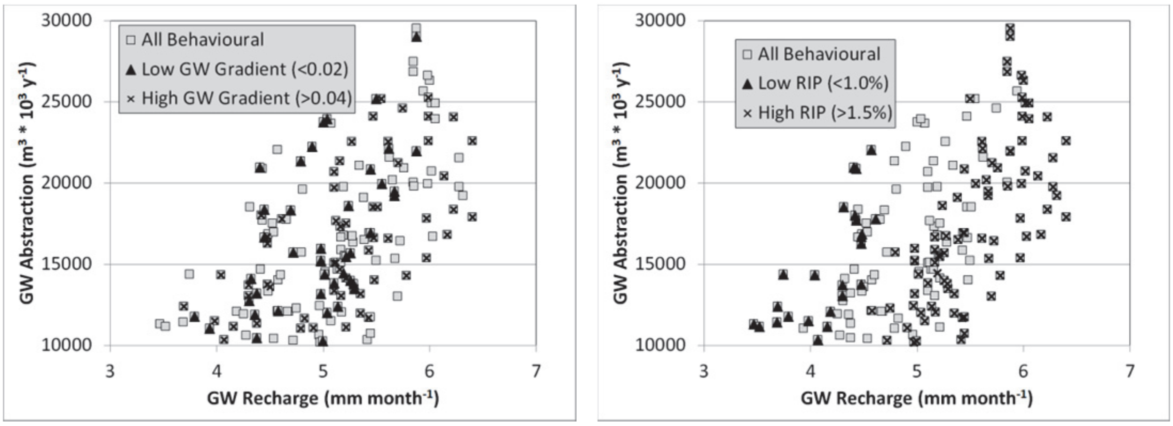

Figure 6.7 Relationships between recharge and abstraction for different extremes of the GW slope (GW Gradient) parameter (left side) and extremes of the riparian strip (RIP) parameter.

In general terms the model is demonstrated to be able to simulate a relatively wide range of different surface groundwater conditions at the scale of the total catchment. This is in stark contrast to the conclusions reached by DWAF (2007b) that suggested that the Pitman model with GW components (the Sami version) was not able to simulate the conditions prevailing in the headwaters of the Breede River catchment. Furthermore, the uncertainty analysis has indicated that some of the uncertainty bounds can be quite substantially reduced, albeit that these conclusions are also dependent upon some additional assumptions (particularly about the most likely range of riparian GW evapotranspiration losses). In summary, the analysis suggests that:

- Mean monthly groundwater recharge should lie between 42 and 54 mm y^{-1}, which represents a volume of 27.6 and 35.5 * 10^6 m^3 y^{-1}.

- Between 10 and 18 * 10^6 m^3 y^{-1} is expected to be abstracted, less than 0.8 * 10^6 m^3 y^{-1} transferred to adjacent catchments, less than about 11 * 10^6 m^3 y^{-1} lost to evapotranspiration and the remainder contributing to stream flows.

While these values are based on using WR2005 rainfall data (mean annual value of 864 mm), using a different value would not make a great deal of difference to the results, but would be balanced by changes to other model parameter values. Further reductions in the uncertainty would have to be based on the collection of additional information. It would clearly be useful to have more accurate information on the amounts of groundwater that are abstracted and obtaining this information is achievable, albeit with a substantial amount of fieldwork and checking. Confirming some of the other water balance components is likely to be much more difficult, if not impossible, at the scale of the whole catchment.

6.5 The Crocodile (West) River catchment: Riparian loss uncertainties.

The approach adopted for the Crocodile River catchment is similar as for the Breede River. The initial uncertainty runs were focussed on identifying appropriate values for the main surface runoff parameters, fixing these and then concentrating on the groundwater uncertainties. The quaternary catchment used in this example is A22C (515 km^2), corresponding to DWA gauge A2H032 a tributary of the Elands River. The whole of the Crocodile River catchment was the subject of a groundwater availability and abstractions study (Vivier et al., 2007), while some of the sub-catchments were included as part of the Tanner (2013) study. One of the major sources of uncertainty identified by Tanner (2013) was the amount of groundwater that is lost to riparian evapotranspiration relative to the recharge and groundwater usage. Mean annual rainfall and potential evaporation are approximately 600 and 1 750 mm y^{-1}, respectively, the topography is relatively flat (except in some areas with incised channels) and the catchment is underlain by fractured quartzites, shales and dolomites. Vivier et al. (2007) referred to relatively high mean annual recharge volumes of 6.7 * 10^6 m^3 y^{-1} (13 mm y^{-1}) and riparian evapotranspiration losses of 11 * 10^6 m^3 y^{-1}, the difference being made up from channel transmission losses. However, this represents a very high value for riparian losses and would require some 1.2% of the total catchment area to be contributing to these losses. An examination of Google Earth images suggests that there could be up to 60 km of channels (effective drainage density of 0.18 km km^{-2}) contributing to groundwater riparian losses and that the riparian strip is less than 50 m. These values are based on the visual evidence of riparian vegetation.

Establishing appropriate parameters for the surface water components of the model proved to be difficult when the model evaluations were based on the objective functions typically used with the model. There is frequently very poor correspondence between simulated and observed flows for individual months, possibly related to poor rainfall data. However, the shape of the observed flow duration curve (FDC) could be well simulated. The evaluation of the simulated groundwater interactions was therefore based on the low flow component of the FDC and specifically the % time of zero flow, which in the observed records was 62%. The interflow function of the model was not used (FT = 0), while the surface runoff parameters were manually calibrated to accurately represent the low frequency of exceedence part of the FDC (20% and lower). Water use data (small farm dams and groundwater) were based on data given in DWAF (2008b). Only GW, GPOW and RIP were then treated as uncertain and Figure 6.8 summarises the results for all ensembles where the simulated 50th FDC percentile is zero. Within these ensembles (5 302) the % time of zero flows varies from 50 to 67% (observed = 62%). The overall conclusion is that the behavioural recharge depth is less than 0.8 mm month^{-1} (4.94 * 10^{6} m^{3} y^{-1} or 1.6% of rainfall) and that the majority is lost to riparian evapotranspiration with some used for irrigation (0.37 * 10^{6} m^{3} y^{-1}, based on DWAF, 2008b).

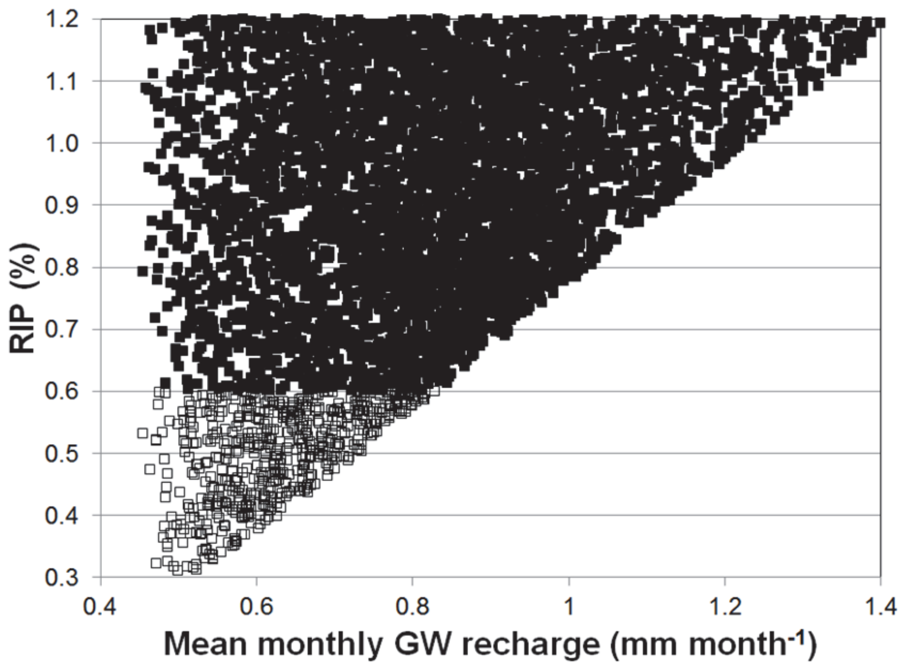

Figure 6.8 Relationship between recharge and riparian strip evapotranspiration parameter (RIP) for the Crocodile River tributary for ensembles with the 50th FDC percentile equal to zero (the open squares are for RIP values of ≤ 0.6%).

6.6 The Gamagara River: Channel transmission losses and aquifer de-watering

Simulations of catchment scale channel transmission losses are extremely difficult to validate due to the lack of real data (Shannafield and Cook, 2014). The Gamagara River (27.84^{o}S 22.98^{o}E) in the Northern Cape Province of South Africa represents a situation where it is possible to use some observed groundwater level data to assess the ability of the model to simulate the transfer of water from the channel during rare flow events as well as the effects of these transfers on groundwater dynamics. This is a very arid area with mean annual rainfall of 400 to 600 mm y^{-1} and high potential evaporation rates (> 2 000 mm y^{-1}). The Sishen open-cast iron ore mine is located within the catchment and aquifer de-watering began in 1967 with annual abstractions stabilising at between 10 and 15 * 10^6 m^3 y^{-1} after about 1981 (Bredenkamp and De Jager, 2011). The assumption is that the de-watering resulted in additional transmission losses from the rare flow events, an assumption supported by anecdotal evidence about reductions in the frequency of flow events from local residents (Bredenkamp and De Jager, 2011). A series of observation boreholes indicate that prior to de-watering groundwater levels near the river fluctuated between 10 m to less than 1 m below the surface with a low amplitude of variation. After de-watering levels dropped to as much as 60m below the surface, but also showed greater degrees of fluctuation that were associated with heavy rainfall events in the catchment. Fluctuations in water level for boreholes located away from the de-watered zone and distant from the channel are approximately between 5 and 30 m below the surface. The model was set up with four sub-catchments, two representing tributary inflows (A=356 and B=1 451 km^2), one representing the de-watered zone (C=316 km^2) and one below the de-watered zone (D=1 522 km^2). There are no observed stream flow data and therefore the patterns of simulated stream flows (before losses due to de-watering) were aligned to existing model results (Midgley *et al*., 1994) and the focus of the study was on the simulated impacts of channel losses on borehole levels. As the model does not simulate groundwater levels directly (Figure 6.9), it was necessary to use the simulated GW slope element gradients (α in Figure 6.9a) to estimate GW depths.

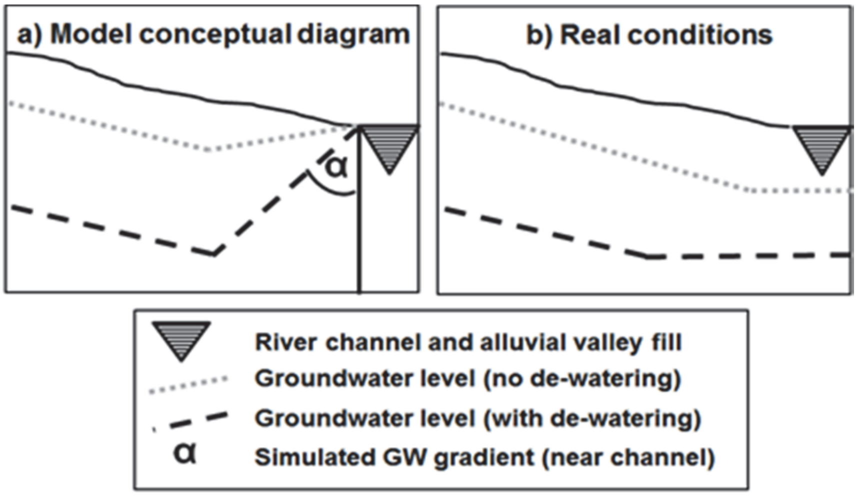

Figure 6.9 Groundwater conditions for the Gamagara River example: a) model concepts, b) real conditions.

In the absence of de-watering the inclusion of the channel transmission loss function in the model has very small effects on the simulated flows largely because of the relatively low volume of groundwater storage available. The simulations of groundwater levels in the sub-catchments are a reasonable reflection of the observed data (Figure 6.10a and b). Borehole BH-162 is close to the outlet of the tributary A, while boreholes BH-193 and BH-396 are just upstream and downstream of the boundary between sub-catchments C and D. Figure 6.10c illustrates the observed and simulated groundwater levels for sub-catchment C after de-watering based on an annual abstraction rate of 13 * 10^6 m^3 y^{-1}. Figure 6.10d shows the results for sub-catchment D (with an additional borehole (BH-214) that is close to BH-396) for simulations with and without de-watering and spanning the period when the de-watering reached a peak between 1977 and 1981. The model was established with groundwater parameters taken from previous reports (DWAF, 2005; Bredenkamp and De Jager, 2011) but one of the observations was that the storativity value for sub-catchment C had to be reduced (from 0.015 to 0.008) for the simulation based on dewatering. This was considered justified on the assumption that before de-watering the majority of the groundwater fluctuations were within the alluvial material or near surface fractured rock, while afterwards they are within more massive rock layers.

Figures 6.10a and 6.10b suggest that the simulations unaffected by de-watering generate results that approximately represent the observed borehole level variations, despite the highly conceptualise nature of the model algorithm (Figure 6.9a). Figure 6.10c suggests that the variations in a borehole close to the river channel (BH-195) within the main de-watering zone are also simulated quite well after de-watering began, but that the rest water level parameter value of 55 m (below which there are no storage changes) could possibly be improved. Borehole BH-195 is some 2.5 km distant from the channel and the delayed effects of transmission loss increments to groundwater are evident, but these cannot be simulated by the model. Figure 6.10d shows the results for the downstream sub-catchment, spans periods than exclude and include de-watering and includes simulations with and without de-watering effects. The de-watering effects here are related to reduced transmission losses and lower sub-surface movements from sub-catchment C after de-watering. The observed borehole levels are closer to the no de-watering simulations before about 1976, but closely follow the de-watered simulations afterwards. The simulations suggest that the reduction in total mean monthly stream flow from de-watering is approximately 20%. Some of the uncertainties inherent in these results include a lack of detailed information about how much groundwater was abstracted between 1967 (de-watering started) and when it stabilised after about 1981. Additional uncertainties exist in the simulations of the flow event volumes and therefore the volumes available for transmission losses, as well as the very simplified loss function used in the model. Despite these, and the limitations of connecting simulated groundwater gradients with actual groundwater levels, the model has adequately reproduced the real world situation.

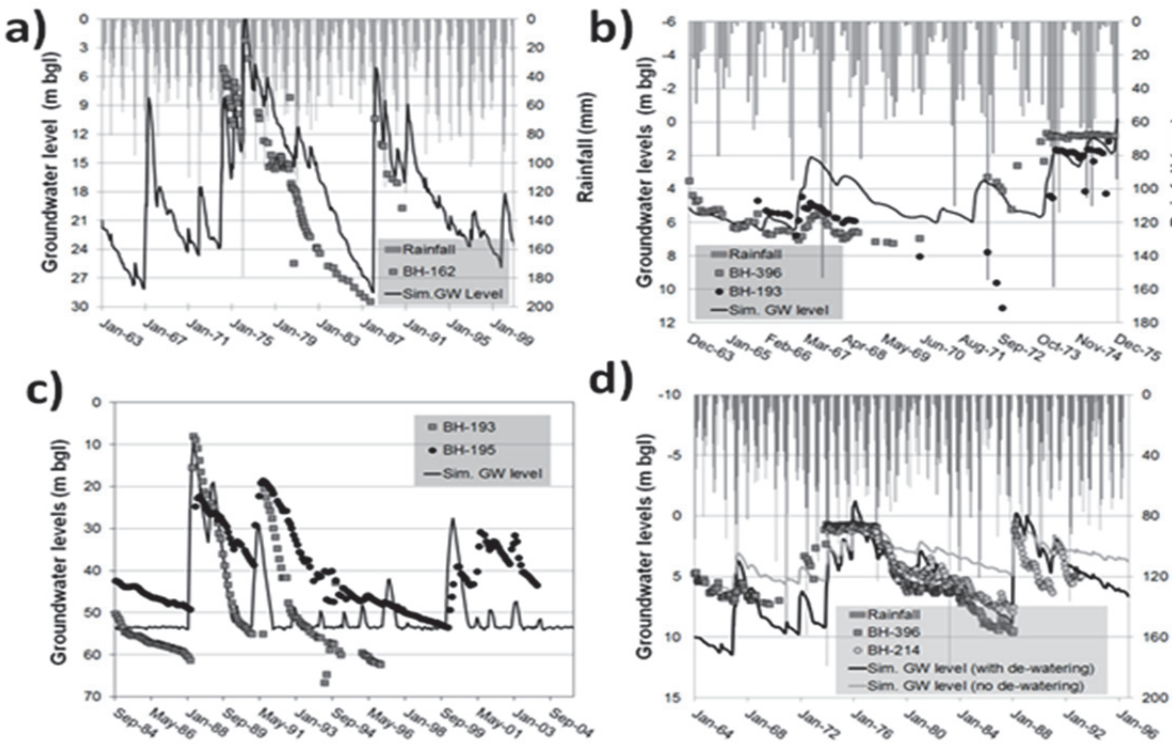

Figure 6.10 Gamagara River results (simulated and observed groundwater levels: a) Sub-catchment A; b) Sub-catchment C without de-watering; c) Sub-catchment C with de-watering; d) Sub-catchment D with and without de-watering (m bgl = metres below ground level).

6.7 The Molopo Dolomitic Eye: Recharge and karst aquifer uncertainties

Accurate characterizations of surface and groundwater interactions in South African karst environments are highly uncertain, and models have had great difficulty simulating the processes in these settings. While there are observed flow data in this sub-catchment (flow gauge D4H014), wide ranging recharge estimates together with uncertain catchment divide boundaries render it difficult to characterize this environment satisfactorily. The dolomitic eye, located in catchment D41A, forms the outlet of a well-developed karst aquifer. According to DWA (Midgley *et al*., 1994) the upstream catchment area of the gauging station is 22 km^2. Preliminary calculations of a simple water balance in the catchment (upstream of the dolomitic eye) indicated that either the surface water catchment is much larger than that reported by DWA (Midgley *et al.*, 1994) or that the groundwater catchment boundary is larger than that of the surface water. In situations where the groundwater catchment size is unknown, normalised baseflow from springs have been used to provide reasonably accurate estimates of the catchment area (Bailly-Comte *et al*., 2009). The three recharge values given by GRAII (DWAF, 2005) for the catchment (D41A) are 7.43, 26.02 and 42.33 mm y^{-1}. These values equate to between 1.5 and 8.3% of mean annual precipitation. With no surface runoff in the sub-catchment, the spring outflow should consist predominantly of groundwater. The outflow from the spring and the recharge estimates for the region have been used to estimate an approximate groundwater catchment area. A catchment area of 22 km^2 (Midgley *et al*., 1994) suggests a volume of between 0.16 and 0.92 × 10^6 m^3 y^{-1}(based on the GRA II recharge estimates), which is far less than the mean annual flow of D4H014 (10.60 × 10^6 m^3 y^{-1}) based on the stream flow gauging records. An examination of a topographic map of the area indicates that the upstream surface catchment area is approximately 120 km^2, however, this is still too small to generate the volume of observed flow, given the recharge estimates reported by GRA II (Table 4-8). It is, therefore, probable that the groundwater catchment boundary is larger than the surface water catchment boundary. The estimates for the (rather large) range of possible recharge values are given in Table 6.5.

Table 6.5 Range of possible groundwater catchment sizes based on spring outflow and recharge estimates (7.43 mm, 26.02 mm and 42.33 mm y^{-1}; DWAF, 2005).

Catchment area (km^2)	Recharge (mm y^{-1})	Spring outflow (10^6 m^3 y^{-1})
120	42.33	5.04
260	42.33	10.92
330	33.00	10.92
420	26.02	10.92
1475	7.43	10.92

The final catchment area in Table 6.5 is the area required to generate the observed spring outflow volume with the lowest GRA II recharge estimate. Considering the advanced karstic development in the aquifer and the extremely large catchment size required to generate the spring outflow using the lower recharge estimate, it seems likely that the mid to upper recharge estimates are more realistic. Employing a recharge depth of 33 mm y^{-1}, a catchment size of approximately 330 km^2 would be required to generate the measured outflow volume. Unfortunately the large range of recharge values prevents a more accurate identification of the groundwater catchment size. The uncertain version of the model was run for four different catchment sizes (Table 6.6) using the same uncertain parameter set. The recharge parameters (GW and GPOW) were set to represent the full range of possible recharge estimated by GRAII (DWAF, 2005). Additional uncertainty was introduced via the parameters representing aquifer storativity (S), transmissivity (T) and the riparian strip factor (RIP).

Table 6.6. Objective functions for the 'best' ensembles from each of the four different catchment sizes.

Catchment area	120 km^2	330 km^2	420 km^2	1475 km^2
Nash Coefficient (untransformed)	-2.05	0.33	0.33	0.30
Nash coefficient (ln transformed)	-3.12	0.38	0.39	0.29
% Bias in simulated monthly flows	-50.17	-4.71	-5.92	0.44
% Bias in simulated monthly ln(flows)	353.51	1.19	10.27	-13.60
Recharge (mm y^{-1})	48.53	37.10	28.63	10.56
Number of behavioural ensembles*	0	209	304	4

*Based on CE>0.2; % Diff 20%

As anticipated, the model run using a catchment area of 120 km^2 did not yield any behavioural ensembles, and the 'best' ensemble (Table 6.6) was achieved with a recharge value higher than the GRAII estimates. Similarly, incorporating a catchment area of 1475 km^2 did not yield many behavioural ensembles and indicates that a recharge value within the mid to upper range (26.02 mm to 42.33 mm y^{-1}; DWAF, 2005) of GRAII estimates is more likely. A groundwater catchment size of approximately 300 to 400 km^2 seems most probable considering the GRAII recharge estimates, the observed outflow and the model simulation results.

Figure 6.11 shows the comparison between the simulated outflows and the observed flow data, and Table 6.7 details the corresponding parameter sets. The dotted line represents a catchment area of 330 km^2 and is the 'best' ensemble from the uncertain model run. The grey line which represents a catchment area of 440 km^2, however, has been calibrated using the single run version of the model to better achieve the peaks of the hydrograph. This was necessary to try and account for the period from 1988 to 1993 where the simulated flow volumes are significantly higher than the observed flow data. No amount of calibration was able to improve the simulation for this period without adversely affecting the remaining well-simulated periods. This period results in the uncertain version of the model 'smoothing' the simulated flow to achieve better statistics when compared to the observed flow. It is not straightforward to identify the reason for this volume disparity, especially as the later years are reasonably well simulated. While there were some rainfall discrepancies between the WR90, WR2005 and a local dataset, they could not account for the flow differences. A period of large abstraction during this time could account for the lower observed volumes, although such a period of increased abstraction does not seem to be likely. While there are clearly housing developments surrounding the pool of the dolomitic eye upstream of the gauging station, no data were available to quantify the patterns of abstraction over the gauged record period. Flow gauging errors could have resulted in a period of incorrectly measured flow, but no additional information is available to confirm or deny such a possibility. A further possibility that has been noted in other Karst spring situations is the existence of lag effects related to storage thresholds or other complex effects within the Karst formations. However, these lag effects are not evident at the Molopo site. The general trend of increased outflow related to increased rainfall (and therefore recharge) is evident through most of the time series except the 1988 to 1993 period.

This setting represents a fairly typical example of karst interaction environments in South Africa in terms of both physical environment and data uncertainty. While parameter values such as transmissivity and storativity will always be uncertain and difficult to characterise, more accurate recharge estimates could significantly reduce the uncertainty associated with both low flow analysis and groundwater catchment size. Unknown variables such as abstractions and evapotranspiration from alluvium filled depressions could also have a significant impact on the results of any model simulation. Considering the lack of data available, the Pitman model was able to capture the general pattern of flow and give a relatively clear indication as to the likely range of recharge and groundwater catchment boundary.

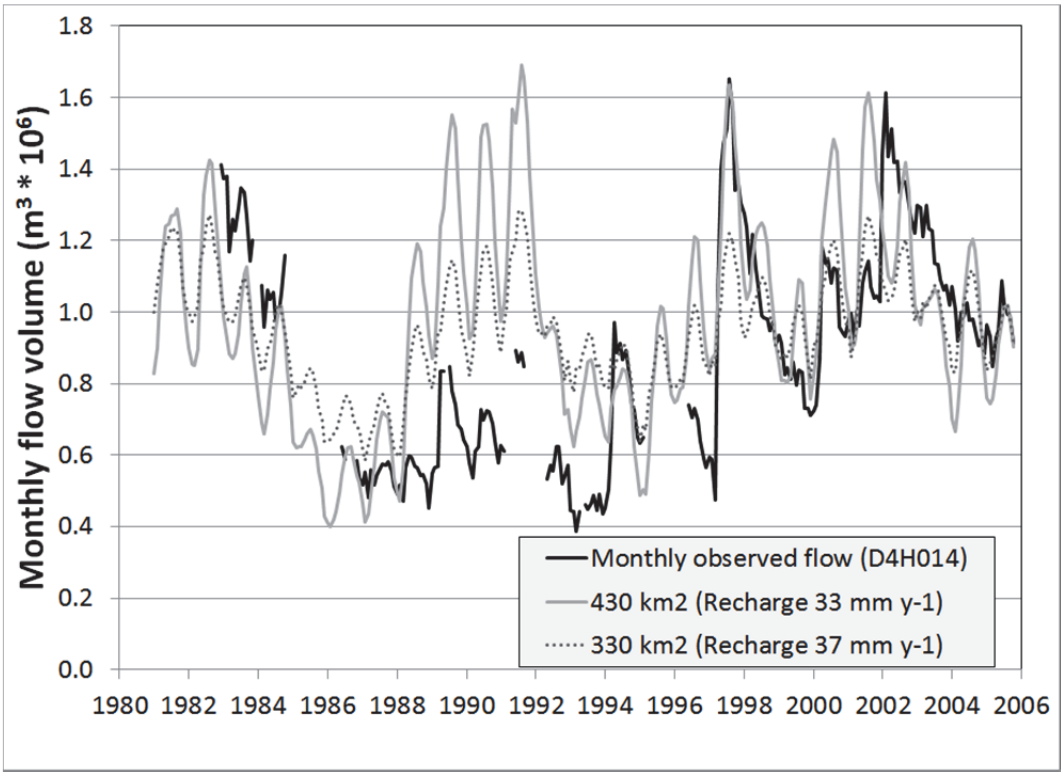

Figure 6.11 Comparison of observed spring flow from the Molopo Dolomitic eye and simulated outflow from the spring.

Table 6.7 Model parameters for the single run versions of the model shown in Figure 6.11.

Catchment size	330km^2	430km^2
ST	280	250
FT	0	0
POW	0	0
GW	20	20
GPOW	2.5	3.0
Drainage density	0.25	0.30
Transmissivity (m^2 d^{-1})	23	25
Storativity	0.011	0.008
Riparian Strip Factor (%)	0.3	0.2

7. CONCLUSIONS

From a modelling perspective, the overall conclusion of the study is that the Pitman model with the revised groundwater routines is a useful tool for exploring different hypotheses (Beven, 2012) about some of the processes involved in SW/GW interactions at the catchment scale. However, in the absence of critical field information, the model is not always able to help resolve some of the uncertainties that exist. Unfortunately, such information is frequently not available and therefore these uncertainties remain regardless of the type of model that is used. Some of the uncertainties can be partially resolved using conceptual 'common-sense' plus limited field information to support such tentative conclusions. However, there is a need to clearly identify those uncertainties that remain unresolved and determine which ones are critical from a water management perspective and design appropriate data collection and research programmes. One of the advantages of an integrated model, such as the Pitman model, is that it is able to assist with the identification of these uncertainties.

With reference to the questions raised in the introduction (Figure 1), Note A that refers to the nature of the recharge process is considered here to be largely irrelevant. The Ecca River example clearly indicates that deep unsaturated zone lateral drainage has to be considered as a potentially important process. Hughes (2009) indicated that it will not always be possible to determine whether such processes are associated with fracture alignment or the existence of perched aquifers, while in some catchments (Sabie River), distinguishing between the low flow contributions to stream flow from lateral unsaturated zone drainage and drainage from the underlying aquifer may not be possible without further field information. Note D refers to the interactions between aquifers and the channel network, and specifically what area of the catchment is likely to be affected by riparian zone evapotranspiration from groundwater. Both the Breede and Crocodile River examples addressed this issue, one in a relatively wet and mountainous area, the other in a more semi-arid situation with ephemeral flows. In both cases some of the uncertainties in the groundwater volumetric balance were reduced with the help of the model together with some assumptions based on field information about the likely area of riparian evapotranspiration losses. In both catchments the total model equifinality was reduced by first of all establishing (and fixing) the parameters of the model that focus on the generation of surface water balance components.

The question raised in Note E will always be difficult to answer, but the Gamagara River example provides some further insights into the process, albeit with a high degree of uncertainty. In this example, relatively low transmission losses were simulated under natural conditions when groundwater levels were always quite close to the surface. However, the model suggests that increasing the available storage through aquifer de-watering can substantially impact on the losses from stream flow. The stream flow simulations could not be validated as there are no observed data, but the model was able to simulate observed borehole level variations quite successfully. Note F was marginally addressed in the Breede River example, but transfer volumes referred to by previous studies (DWAF, 2007a) are not a major component of the total water balance. The Molopo Dolomite eye example also addressed Note F, although in a very karst specific environment. Indications are that surface water and groundwater catchment divides in karst environments can be very different and knowing this difference is essential to correctly representing the processes in a model. The Pitman model was able to successfully constrain both the recharge and groundwater catchment size. Finally, Note G was not really addressed as most of the abstractions in both the Breede and Crocodile River catchments are close to the river channels and the open-cast mine in the Gamagara River example is also relatively close to the channel.

Hydrological models will never be adequate substitutes for real data, but there are many situations where real data are not, and never will be, available. This is especially true with respect to quantifying the water balance components of SW/GW interactions at the catchment scale. This study has demonstrated that a relatively simple hydrological model, operating at relatively coarse spatial and temporal scales, can be useful for quantifying these components, albeit with a relatively large amount of uncertainty remaining. The model is also able to point to areas of uncertainty that could be the focus of future field investigations that might include environmental tracer studies (Tetzlaff and Soulsby, 2008) to resolve some of the model unknowns. Such studies are relatively rare within South Africa but arguably will offer one of the best opportunities for further reduction in the uncertainty in the way in which models represent real world processes and therefore more reliable simulations.

It is hoped that this report will go some way towards counteracting the very negative attitude towards the revised Pitman model (either the Rhodes University version used here or the Sami version) that has prevailed within some hydrogeological communities in South Africa. Given that the whole motivation for developing the model was to improve the way in which surface water and groundwater simulations can be integrated, it is unfortunate that it seemed only to increase the divisions between surface and groundwater hydrologists. As mentioned many times in the past, there was never any intention to challenge the need for existing types of groundwater models. The purpose of these models is frequently different to that of the Pitman model and previous experiences suggested that existing groundwater models could not fulfil some of the needs that the modifications for the Pitman were made for. It is also interesting to note that, subsequent to the development of the modifications, other research projects that have been designed to investigate surface water-groundwater interactions have been initiated by some of the models fiercest critics. At least some of these have developed even simpler water balance routines that ignore some of the processes that have been included in the Pitman model. It is not possible to reference these studies as they have yet to be published. The authors of this report would contend that the processes that have been included in the Pitman model are the minimum that are necessary to deal with the variety of situations found in South Africa (and other parts of the sub-continent). This contention has been partly supported by the example studies presented in this report and within the PhD thesis of Dr Jane Tanner (2013). Further simplification is therefore not considered as a viable option.

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APPENDIX A: GLOSSARY OF TERMS

ALLUVIUM: a general term for all detrital deposits resulting from the operations of modern rivers, including the sediments laid down in river beds, flood plains, lakes, fans at the foot of mountain slopes and estuaries.

AQUIFER: a geological formation which contains water.

AQUITARD: a confining bed that retards but does not completely stop the flow of water to or from an adjacent aquifer; while it may not readily yield water to boreholes and springs, it may act as a storage unit.

ARTESIAN: artesian conditions obtain when the hydrostatic pressure exerted on an aquifer is great enough to cause the water to rise above the water table.

BANK STORAGE: water that percolates laterally from a river in flood into the adjacent geological material, some of which may flow back into the river during low-flow conditions.

BASEFLOW: sustained low flow in a river during dry or fair weather conditions, but not necessarily all contributed by groundwater; includes contributions from both interflow and groundwater discharge (Parsons, 2004).

BERG WAAS: Berg Water Availability Assessment Study.

CONCEPTUAL MODEL: a computer model with equations based upon a simple interpretation of the physical processes acting upon the inputs and outputs of a system. A PERCEPTUAL MODEL can also be termed a conceptual model.

CONFINED AQUIFER: an aquifer bounded above and below by confining beds in which groundwater is under greater pressure than that of the aquifer.

DIFFUSE RECHARGE: is spatially distributed and results from widespread percolation through the whole vadose zone. It is defined as water added to the groundwater store in excess of soil-moisture deficits and evapotranspiration, by direct vertical percolation of precipitation through the unsaturated zone. Also termed direct recharge.

DOLOMITE: term applied to a carbonate rock composed predominantly of dolomite.

DOLERITE: a medium grained intrusive igneous rock of basaltic composition.

DWA: Department of Water Affairs.

DWAF: Department of Water Affairs and Forestry.

DYKE: a tabular body of intrusive igneous rock that cuts across the layering or structural fabric of the host rock.

EFFLUENT STREAM: a stream with a piezometric surface lower than the groundwater surface. The stream is fed directly by groundwater. Also termed gaining stream.

EPHEMERAL RIVERS: a stream that flows occasionally and at these times the stream flow consists of mostly event based runoff. These rivers have a limited (if any) baseflow component with no groundwater discharge.

EQUIFINALITY: the possibility of many different parameter sets within a model structure that might give equally acceptable results when compared with observations.

FAULT: a fracture in earth materials, along which the opposite sides have been relatively, displaced parallel to the plane of movement.

FLUX: rate of groundwater flow per unit width of aquifer.

FRACTURE: cracks, joints, faults or other breaks in the rock that can enhance water movement.

FRACTURED ROCK AQUIFER: an aquifer where water resides in fractures which are expected to have strong influences on the movement (direction and rate) of groundwater. Also termed HARD ROCK AQUIFER AND SECONDARY AQUIFER.

FRACTURE FLOW: water movement in either the unsaturated or saturated zone that occurs in fractures and fissures.

GRA II: Groundwater Resource Assessment, Phase II.

GRDM: Groundwater Resource Directed Measures.

GRIP: Groundwater Resource Information Project.

GROUNDWATER: that part of the subsurface water which is in the zone of saturation below the regional groundwater level. In this thesis, water in a perched aquifer is not considered groundwater but rather unsaturated zone water.

GRU: Groundwater Management Unit.

HARD ROCK AQUIFER: see FRACTURED ROCK AQUIFER.

HYDRAULIC HEAD: the height of the exposed surface of a body of water above a specified sub-surface point.

HYDRAULIC CONDUCTIVITY: measure of the ease with which water will pass through earth material; defined as the rate of flow through a cross-section of one square metre under a unit hydraulic gradient at right angles to the direction of flow (in m/d).

HYDRAULIC GRADIENT: the slope of the water table or piezometric surface is a ratio between the difference of elevation (hydraulic head) and the distances between the two points of measurement. The rate and direction of water movement in an aquifer are determined by the permeability and the hydraulic gradient.

INDIRECT RECHARGE: results from percolation to the water table following runoff and localisation in fractures, ponding in low lying areas or lakes and through the beds of surface water bodies.

INFLUENT STREAM: a stream with a piezometric surface higher than the groundwater surface and which discharges into the underlying groundwater system through transmission losses. Also termed losing stream.

INTERFLOW: the flow of water along unsaturated flow paths (above the regional groundwater table) that can move both vertically and laterally before discharging into other water bodies. Interflow can include soil moisture flow, unsaturated fracture flow or water from perched aquifers. See also UNSATURATED ZONE FLOW.

IWRM: Integrated Water Resource Management.

KARST AQUIFER: a body of soluble rock that conducts water principally via enhanced (conduit or tertiary) porosity formed by the dissolution of the rock. Karst aquifers include a wide variety of more or less karstified limestone from the less developed or diffuse flow aquifers to the highly localised or conduit flow aquifers.

NGA: National Groundwater Archive.

NGWD: National Groundwater Database.

NWA: National Water Act.

PERCEPTUAL MODEL: the qualitative understanding of the processes occurring in an environment or catchment. Can also be termed conceptual model or conceptual understanding.

PERCHED AQUIFER: a local body of water above an impermeable layer of very limited extent such as a lens of clay within a sandstone bed.

PERENNIAL STREAM: a stream that flows throughout the year (except perhaps during extreme drought periods).

PERMEABILITY: the measure of the ability of earth materials to transmit a fluid. It depends largely on the size of pore spaces and their connectedness. Defined as the volume of fluid discharged from a unit area of an aquifer under unit hydraulic gradient in unit time (expressed as $m^3/m^2/d$ or m/d); not to be confused with *hydraulic conductivity* which relates specifically to the movement of water.

PHYSICALLY BASED MODELS: models with parameter values which have a physical interpretation and which represent spatial variability in the parameter values.

PIEZOMETRIC SURFACE: an imaginary surface representing the piezometric pressure or hydraulic head throughout all or part of a confined or semi-confined aquifer; analogous to the water table of an unconfined aquifer.

POROSITY: ratio of the volume of interstices in a soil or rock to the total volume, usually stated as a percentage.

PRIMARY AQUIFER: an aquifer in which groundwater moves through the original interstices of the geological formation, i.e. sand grains. Primary aquifers can include a wide variety of aquifers which include sand aquifers (coastal, dune, lowland floodplains etc.), chalk aquifers and glacial deposits. Also included in this category are regolith or saprolite aquifers.

QUATERNARY CATCHMENT: a fourth order catchment (basic hydrological unit) in a hierarchal classification system in which a primary catchment is the major unit. The quaternary catchment is the basic unit for water resources management in South Africa.

RECHARGE: the addition of water to the groundwater table (not including perched aquifers or any other form of interflow).

REGOLITH: fragmented or unconsolidated rock material of residual or transported origin, comprising rock debris, alluvium, aeolian deposits, till, loess and *in situ* weathered and decomposed rock and typically overlies bedrock; it includes soil.

RIPARIAN: area of land directly adjacent to a stream or river, influenced by stream-induced or related processes.

SAPROLITE: a soft, earthy, clay-rich and totally decomposed rock, formed in place by weathering of rocks. Structures that were in the unweathered rock are preserved in saprolite.

SEASONAL RIVER: a stream with intermittent flow which might consist of some baseflow in the wet season but no sustained flow in the dry season. Interaction with groundwater depends on the fluctuating position of the water table, ranging from effluent streams in the wet season to influent streams in the dry season.

SECONDARY AQUIFER: an aquifer in which groundwater moves through secondary openings and interstices, which developed after the rocks were formed.

SEMI-CONFINED AQUIFER: an aquifer that is partly confined by layers of lower permeability material through which recharge and discharge may occur, also referred to as a *leaky aquifer*.

STORATIVITY: the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head. Also termed storage coefficient.

TDS: Total Dissolved Solids.

TMG: Table Mountain Group.

TRANSMISSIVITY: the rate at which a volume of water is transmitted through a unit width of aquifer under a unit hydraulic head (m^2/d); product of the thickness and average hydraulic conductivity of an aquifer.

UNCONFINED AQUIFER: an aquifer with no confining layer between the water table and the ground surface where the water table is free to fluctuate.

UNSATURATED ZONE: or vadose zone is that part of the geological stratum above the groundwater table where interstices and voids contain a combination of air and water.

UNSATURATED ZONE FLOW: the flow of water along unsaturated flow paths (above the regional groundwater table) that can move both vertically and laterally before discharging into other water bodies. Unsaturated zone flow can include soil moisture flow, unsaturated fracture flow or water from perched aquifers. See also INTERFLOW.

WARMS: Water Authorisation Registration and Management System.

WR90, WR2005: Water Resources 1990 and 2005.