UNDERSTANDING AND MODELLING OF SURFACE AND GROUNDWATER INTERACTIONS.

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ABSTRACT

The connections between surface water and groundwater systems remain poorly understood in many catchments throughout the world and yet they are fundamental to effectively managing water resources. Managing water resources in an integrated manner is not straightforward, particularly if both resources are being utilised, and especially in those regions that suffer problems of data scarcity. This study explores some of the principle issues associated with understanding and practically modelling surface and groundwater interactions. In South Africa, there remains much controversy over the most appropriate type of integrated model to be used and the way forward in terms of the development of the discipline; part of the disagreement stems from the fact that we cannot validate models adequately. This is largely due to traditional forms of model testing having 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. While model structural uncertainties are important to consider, the uncertainty from input data error together with parameter estimation error are often more significant to the overall residual error, and essential to consider if we want to achieve reliable predictions for water resource decisions. While new philosophies and theories on modelling and results validation have been developed (Beven, 2002; Gupta *et al*., 2008), in many cases models are not only still being validated and compared using sparse and uncertain datasets, but also expected to produce reliable predictions based on the flawed data. The approach in this study is focused on fundamental understanding of hydrological systems rather than calibration based modelling and promotes the use of all the available 'hard' and 'soft' data together with thoughtful conceptual examination of the processes occurring in an environment to ensure as far as possible that a model is generating sensible results by simulating the correct processes.

The first part of the thesis focuses on characterising the 'typical' interaction environments found in South Africa. It was found that many traditional perceptual models are not necessarily applicable to South African conditions, largely due to the relative importance of unsaturated zone processes and the complexity of the dominantly fractured rock environments. The interaction environments were categorised into four main 'types' of environment. These include karst, primary, fractured rock (secondary), and alluvial environments. Processes critical to Integrated Water Resource Management (IWRM) were defined within each interaction type as a guideline to setting a model up to realistically represent the dominant processes in the respective settings. The second part of the thesis addressed

the application and evaluation of the modified Pitman model (Hughes, 2004), which allows for surface and groundwater interaction behaviour at the catchment scale to be simulated. The issue is whether, given the different sources of uncertainty in the modelling process, we can differentiate one conceptual flow path from another in trying to refine the understanding and consequently have more faith in model predictions.

Seven example catchments were selected from around South Africa to assess whether reliable integrated assessments can be carried out given the existing data. Specific catchment perceptual models were used to identify the critical processes occurring in each setting and the Pitman model was assessed on whether it could represent them (structural uncertainty). The available knowledge of specific environments or catchments was then examined in an attempt to resolve the parameter uncertainty present within each catchment and ensure the subsequent model setup was correctly representing the process understanding as far as possible. The confidence in the quantitative results inevitably varied with the amount and quality of the data available. While the model was deemed to be robust based on the behavioural results obtained in the majority of the case studies, in many cases a quantitative validation of the outputs was just not possible based on the available data. In these cases, the model was judged on its ability to represent the conceptualisation of the processes occurring in the catchments. 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. Many water resource decisions are still made without adequate account being taken of the uncertainties inherent in assessing the response of hydrological systems. Certainly, with all the possible sources of uncertainty in a data scarce country such as South Africa, pure calibration based modelling is unlikely to produce reliable information for water resource managers as it can produce the right results for the wrong reasons. Thus it becomes essential to incorporate conceptual thinking into the modelling process, so that at the very least we are able to conclude that a model generates estimates that are consistent with, and reflect, our understanding (however limited) of the catchment processes.

It is fairly clear that achieving the optimum model of a hydrological system may be fraught with difficulty, if not impossible. This makes it very difficult from a practitioner’s point of view to decide which model and uncertainty estimation method to use. According to Beven (2009), this may be a transitional problem and in the future it may become clearer as we learn more about how to estimate the uncertainties associated with hydrological systems. Until then, a better understanding of the fundamental and most critical hydrogeological processes should be used to critically test and improve model predictions as far as possible. A major focus of the study was to identify whether the modified Pitman model could provide a practical tool for water resource managers by reliably determining the available water resource. The incorporation of surface and groundwater interaction routines seems to have resulted in a more robust and realistic model of basin hydrology. The overall conclusion is that the model, although simplified, is capable of representing the catchment scale processes that occur under most South African conditions.

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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 and releases it in significant quantities for use.

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 satuaration 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: Water Resources 1990.

WR2005: Water Resources 2005.

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DECLARATION

The following thesis has not been submitted to a university other than Rhodes University, Grahamstown, South Africa. The work presented here is that of the author.

Much of the work carried out during the doctoral research also formed part of a study funded by the Water Research Commission aimed at reducing the uncertainty in water resource estimations. Two reports were submitted as part of the study which directly related to the doctoral research and were aimed at reducing the uncertainty in quantifying surface water and groundwater interactions. All of the case studies examined in this thesis as well as a portion of the literature review were included in the reports. The student was largely responsible for the surface water and groundwater interaction component of the WRC project and carried out all of the modelling for the case studies, except for the Sabie River and Grahamstown study sites. Prior modelling of the Sabie River study site by Prof. Hughes assisted in the setting up of the model for the current study and the current research built upon the earlier modelling foundations. The modelling of the Grahamstown study site was carried out jointly by both Prof. Hughes and the student. Both WRC reports were written by the student and edited by Prof. Hughes.

1 INTRODUCTION

This study explores some of the principle issues 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 impacts of the over-abstraction of groundwater on stream flow in rivers has led to the recognition that both surface water and groundwater are a common resource and has mobilised research toward improving conceptual understanding and toward the development of tools to quantify the interactions between surface and groundwater. These tools, such as integrated models, aim to provide a more realistic analysis of each resource through an improved representation of the real world. This integration is held back 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. Hydrological models have the potential to support integrated water resource management but only if uncertainties in the models are recognised and dealt with appropriately.

Hydrology (both surface and groundwater hydrology) is a difficult science; it aims to represent highly variable and non-stationary processes which occur in catchment systems, many of which are unable to be measured at the scales of interest (Beven, 2012). The conceptual representations of these processes are translated into mathematical form in a model. Different process interpretations together with different mathematical representations results in diverse model structures. These 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 then used to drive the models and generate quantitative information. Approximate implementations (model structures and parameter sets), driven by approximate input data will necessarily produce approximate results (Beven, 2012). Most model developers aim to represent reality as far as possible, and as our understanding of catchment processes has improved, models have tended to become more complex with the incorporation of additional processes. 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. 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), defined in this thesis as the possibility of many different parameter sets within a model structure that might give equally acceptable results when compared with observations. 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.

While structural uncertainties are often difficult to separate from parameter and data input uncertainties, they can be very important if one of the objectives is getting the right answer for the right reason (Kirchner, 2006). This is particularly relevant to simulating surface and groundwater interactions in hydrological models. It is important, however, not to confuse structural uncertainty and structural simplicity (or complexity) as they are not necessarily linked. Arguably, there are two levels of structural uncertainty that should be considered. The first is whether or not certain processes, known to exist in the real world, are represented in a model. The second is whether the algorithms used in the model can adequately represent the non-linearities or thresholds that occur within the relationships between storages and fluxes, or one flux and another. Many of these uncertainties are difficult to resolve given the typically available data that can be used to define and quantify surface and groundwater interactions. Beven (2012) identifies two common ways of assessing the 'correctness' of a model structure and setup. The first is to rely on expert opinion, but bias is inevitably introduced due to the difficulty of finding scientists not committed to one modelling paradigm or another. The second is to test the model against available data for a range of different circumstances. However, fitting a model to historical data is subject to the difficulties of differentiating between the various sources of uncertainty (structure, setup, input data). Beven (2012) suggests a "limit of acceptability" approach to model evaluation as a way of testing models, which would involve thought to define critical experiments that will allow models and their setups to be adequately differentiated.

The approach adopted in this thesis 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 (structural uncertainty). The available knowledge of specific environments or catchments is then examined in an attempt to resolve the equifinality present within each catchment and therefore ensure the subsequent model setup is realistically representing the process understanding as far as possible. While it is difficult to differentiate between input, parameter and structural uncertainty, it should be possible to at least partly identify the 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.

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. A spatially lumped modelling approach in the management of water resources has limitations, especially due to the lack of spatial detail. However, it offers advantages including facilitating a better understanding of large-scale water management issues, assessing the impacts of water allocation and groundwater abstraction on stream flow at the catchment scale and informing water sharing plans. There are, however, still relatively few studies at larger spatial scales which investigate the nature of interacting controls on baseflow generation (Shaman *et al*., 2004). The development of guiding principles for combining data and models at different spatial and temporal scales and extrapolating information between scales remains a challenge.

Any interaction environment becomes more complex as the scale of investigation becomes more detailed. The question is can we reliably quantify the processes occurring at scales appropriate for integrated water resource management? This study proposes the use of a 'compromise' model, consisting of an existing widely used surface water model which has had more explicit groundwater routines incorporated. In the South African context the model (Pitman, 1973) must be acknowledged as one of the most generally accepted hydrological models available due to its long history of use, ease of parameterisation and acknowledged reliability for surface water resource assessments. Groundwater algorithms have recently been developed and incorporated into the model (Hughes, 2004) for the purpose of integrating surface and groundwater and assessing the impacts of groundwater abstraction. The model operates at a catchment scale and the parameters represent spatial averages rather than direct point values from field measurements. While the Pitman model was originally designed to operate at large scales (100's to 1000's km^2), subsequent modifications mean the model can be applied at smaller scales in situations which are particularly heterogeneous or which require more detailed information. There are however, uncertainties associated with the use of the model and its applicability in certain environments. This study therefore attempts to provide a better understanding of the processes involved and to resolve or assess some of the uncertainties associated with the interaction components of the model.

1.1. Background

Historically in South Africa, surface and groundwater were isolated in policy and regulation with groundwater having a private status under the previous South African Water Act (Act 56 of 1956). Groundwater could be used by a borehole owner with no restrictions, and was not included in national water resource estimations. As a result, surface and groundwater practitioners largely worked in isolation and seldom dealt with the interconnectivity between the two disciplines (Hughes, 2004) even though they may have been aware of its importance. The more recent Water Act (Act 36 of 1998) requires that water resources are managed in a holistic and integrated fashion. The new legislation also established the reserve, which includes the basic human needs reserve and the ecological reserve (maintenance of some portion of the natural flow regime in a river to ensure ecological sustainability and preserve biodiversity). All other water uses are subject to a system of allocation licences and general authorisation (Levy and Xu, 2011). The ecological reserve applies to both surface water and groundwater bodies. This has encouraged surface and groundwater practitioners to begin working together and integrating their knowledge of both resources. However, a recognised lack of quantitative information on the groundwater contribution to flow regimes in South Africa still remains (Parsons, 2004). While water resource estimations are still carried out in a non-integrated fashion, the processes associated with the two resources (surface water and groundwater) are inextricably linked and integrated modelling can contribute toward a more robust and complete representation of the catchment processes, thereby resulting in more reliable water resource estimations for both surface water and groundwater assessments. In order to allocate water in accordance with the requirements of the National Water Act, it is imperative that the interdependencies are realised and incorporated into the assessments and allocations of each resource (Kelbe and Germishuyse, 2010).

Nearly fifteen years has passed since the National Water Act (NWA) was promulgated, but there are many factors that contribute to the challenge of successful implementation of the NWA. One of these is the scientific uncertainty that surrounds quantifying surface and groundwater interactions. 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. The setting of a sustainably available quantity of groundwater is a contentious issue, together with the way forward in terms of the integrated management of the resource. Estimating the volumes and fluxes of groundwater in South African aquifers is particularly difficult due to the widespread occurrence of fractured rock aquifers. Currently, groundwater hydrologists tend to focus more on local settings when completing water resource assessments, while surface water hydrologists are more focused on regional investigations that are designed to meet management objectives. Groundwater hydrologists commonly characterise the geological environment and groundwater flow system within an aquifer and surface hydrological processes may be neglected. Surface water hydrologists, on the other hand, have relied on characterising the overall rainfall-runoff response and often neglect groundwater processes. As a result, different methods or models of assessment have been developed for surface and groundwater resource assessments. Surface water models generally operate on relatively large catchment scales and use rainfall, evaporation and catchment conditions like vegetation, soil and topography to produce time series variations of simulated stream flows (Perrin *et al*., 2003; Hughes, 2004; Mouelhi *et al*., 2006). Groundwater models, however, are typically detailed physically based models, which are applied over much smaller areas than surface water models (Sophocleous *et al*., 1995; Prudic *et al*., 2004; Kelbe and Germishuyse, 2010). They mostly simulate steady state conditions and not time series variations. This means that it has been difficult to successfully link surface and groundwater models together largely because of differences in scales and structures of the models, as well as different objectives of the modellers or scientists (Hughes *et al*., 2010).

Due to very few experimental catchments in South Africa in comparison to many developed countries (North America, Europe and Australia), the most common approach to integrated water resource assessments is to estimate average annual fluxes at the surface water catchment scale with baseflow separation techniques and then subtract the groundwater discharge rate from the recharge rate, using a simple water balance. This approach, while useful, ignores spatial and temporal variability and cannot contribute to improved process understanding. This is an important consideration for many regions in South Africa as different processes can produce similar low flow signatures. From a water resources management perspective, it is important to further assess the contribution that other sources of 'baseflow' make to stream flow in different parts of South Africa. Without this information, it is possible that incorrect conclusions will be reached about the impacts of different catchment developments on the flow in rivers. This would include land use developments that affect downward percolation through the unsaturated zone, as well as the abstractions from groundwater sources. This emphasises the care that must be taken in inferring sub-surface processes from an interpretation of stream flow data such as baseflow separation methods, particularly if we are interested in getting the right results for the right reasons.

1.2. Aims and objectives

The ultimate objective of this study is 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. 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.

To achieve this overall the study has the following specific aims:

1. Develop conceptual hypotheses about the surface and groundwater interaction processes in different environments.

2. Assess the conceptual 'correctness' of the Pitman model together with its ability to quantify the processes identified.

3. Investigate uncertainties associated with the structure and application of the groundwater interaction components of the model.

4. Develop recommendations for environments where the model can and cannot be reliably applied.

1.2.1. Develop conceptual hypotheses about the surface and groundwater interaction processes in different environments

It is essential to be able to interpret the results of a modelling exercise in terms of a conceptual understanding of the interaction processes and how the results relate to real hydrological processes. The main processes that are directly or indirectly involved in surface water and groundwater interactions are shown schematically in Figure 1-1. The magnitude and importance of the interactions between groundwater and surface water systems vary considerably in both scale and locality. Deep confined aquifers are unlikely to make a major contribution to groundwater discharge (groundwater baseflow) while small localised groundwater seeps/springs may be the only source of freshwater for human consumption and ecological sustainability in a local setting. Consequently, it is necessary to understand the linkages and interactions in terms of both the spatial and temporal scales when describing the processes (Kelbe and Germishuyse, 2010). In doing this it is also necessary to consider the spatial and temporal scales at which the model is typically applied.

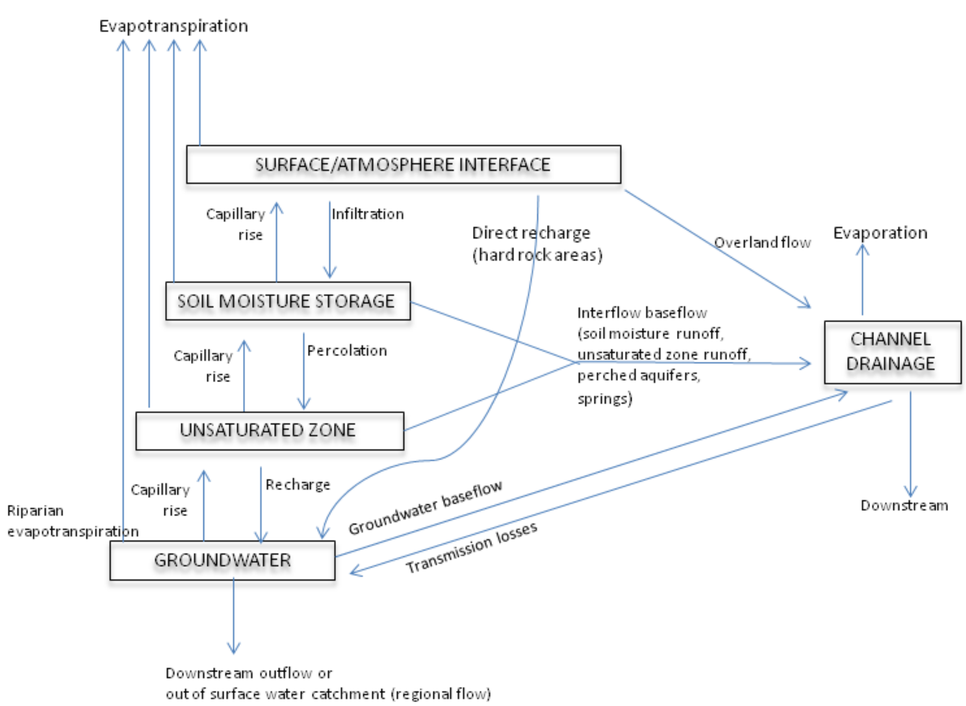


Figure 1-1 **Basic conceptual understanding of surface and groundwater interaction.**

1.2.2. Assess the conceptual 'correctness' of the Pitman model together with its ability to quantify the processes identified.

The model is assessed by comparing the simulation outputs with real data (where available), outputs from other models that have already been tested or, at the very least, against a conceptual interpretation of reality. It was recognised at the start of the study that the sources, availability and accuracy of real data would vary considerably within the region and clearly affect the confidence in the comparison. In terms of quantifying the interactions for the purposes of IWRM, real or 'hard' data is preferable as it enables a less subjective assessment of the model outputs. However, the value of qualitative or 'soft' data is often underestimated and can be used to assess a model's outputs based on reasoning and logic. While ideally both types of data would be available, in many situations no 'hard' data exists; in these cases the results of the model are assessed based on an understanding of conditions in those catchments and evaluating whether the outputs from the different model components are realistic.

1.2.3. Investigate uncertainties associated with the structure and application of the groundwater interaction components of the model.

The identification of the sources of uncertainty is essential for any integrated assessment. There will always be higher levels of risk associated with the management of water resources in South Africa due to the high degree of both spatial and temporal variability in available resources, as well as data scarcity. The research focuses specifically on the reliability of the input hydro-meteorological data (typically rainfall and evaporation demand in rainfall-runoff models), the parameter values assigned to the algorithms used to represent the surface and groundwater interaction processes in a basin, model structure which includes spatial and temporal scale issues, adequate process representation, and water use or return flows. While many of these sources of uncertainty are common to any modelling approach, the research aims to isolate the dominant sources of uncertainty relevant to the use of the interaction components of the model.

1.2.4. Develop recommendations for environments where the model can and cannot be reliably applied.

It is accepted that the way in which the groundwater dynamics have been conceptualised within the model will not be appropriate for all situations. While the model structure has been developed to incorporate as many of the main processes found in South Africa as possible, there are processes that are not simulated, which need to be identified to prevent the incorrect application of the model. Additionally there are processes that are not explicitly represented in the model, but can be represented through manipulation of the model setup and parameters. Recommendations for the use of the model in these environments together with information that is pertinent to ensuring an environment is simulated properly must to be identified.

1.3. Research questions and justification

Some of the key research questions being asked in the context of this research are:

- Are there sufficient data to conceptualise the interaction processes for use in 'any' model?
- Can the Pitman model produce realistic results based on the representation of these processes?
- Is it possible to assess if the model is realistically representing these processes?
- Is it possible to assess if the model is producing realistic results?
- Are there specific groundwater processes that the model doesn't represent adequately?
- Can a simple model adequately represent spatially and temporally variable interaction processes?

A major focus of the study is to identify whether the model could provide a practical tool for both surface and groundwater water resource managers by reliably determining the available water resource. The Pitman model is a widely used and trusted model for surface water resource assessments and since the additional groundwater components have been incorporated, has the potential to assist in IWRM. However, the new functions have not been fully tested in the range of environments typically found in South Africa. Due to the frequent lack of actual data with which to assess the model performance, it is often assessed against alternative data and knowledge. Realistically the model can only be evaluated if there is an adequate conceptual understanding of the processes the model is aiming to represent. The study endeavours to create an improved understanding of how the Pitman Model can be used and increase confidence in the use of the model for integrated assessments.

2. LITERATURE REVIEW

2.1. Introduction

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 (Ivkovic, 2009). Comprehensive papers regarding the physical interactions that occur between groundwater and surface water systems have been written internationally (Brunner *et al*., 2009a, Sophocleous, 2002, Winter *et al*., 1998, Winter, 1999 and Woessner, 2000). 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, 2011, Parsons, 2004 and Xu *et al*., 2002). South African research that has focused on the physical processes behind the interactions has tended to be on a smaller scale due to funding and data constraints (Hughes and Sami, 1992; Lorentz *et al*., 2004; Roets *et al*. 2008; Wenningger *et al*., 2008; Kelbe and Germishuyse, 2010).

2.1.1. 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 thesis. 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, 2004; 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; 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 thesis 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 in this thesis will follow the same guidelines.

2.2. Groundwater-surface water interaction: basic principles

2.2.1. Introduction

Contributions, both local and international, have classified different types of interactions that occur between surface and groundwater systems. Some of these papers focus on the processes occurring at detailed scales (Harvey and Bencala, 1993, Winter *et al*., 1998, Sophocleous, 2002, Ivkovic, 2009), while others focus on larger scales for the purpose of water resource estimation (Smakhtin, 2001, Vegter and Pitman, 2003, DWAF, 2008a). 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.

Streams may be classified based on various criteria for different purposes. In this thesis, streams are divided into 3 classes depending on their runoff characteristics namely ephemeral, seasonal and perennial. These characteristics usually give an indication of the degree of connection between the river and the groundwater system. This degree of connection occurs in three basic ways (Winter *et al*., 1998, DWAF, 2004a): (1) Rivers gain water from inflow of groundwater through the streambed (gaining stream); (2) they lose water to groundwater by outflow through the streambed (losing stream); or (3) they do both, gaining in some reaches and losing in other reaches, or gain and lose in the same reach at different times. A perennial stream is defined as having stream flow throughout the year (except perhaps during extreme drought periods). A ephemeral stream can be defined as a stream that flows occasionally and at these times the stream flow consists of mostly event based runoff. A seasonal stream is defined as having intermittent flow which might consist of some baseflow in the wet season but no sustained flows in the dry season. Groundwater also interacts with surface water features such as lakes, wetlands, estuaries and the sea. While these interactions can form an important part of the water cycle in a catchment, the focus of this thesis is on the interactions between groundwater and rivers, as these are the dominant points of interaction in terms of water resource management.

Probably the first conceptual illustration detailing all the possible flow regimes was published by Meinzer (1923). As well as gaining and losing streams, Meinzer (1923) defined perched systems where the stream and water table are separated through an unsaturated zone. Since then, interaction types have been characterised in many ways, with different levels of detail. Nield *et al*. (1994) identified 39 flow regimes of surface and groundwater interaction, with their characteristics controlled by regional water table gradients, recharge to the aquifer, water body length, aquifer anisotropy and the hydraulic resistance of the bottom sediments. Smakhtin (2001) showed how we can envisage a river catchment as a series of interlinked reservoirs. Each of these reservoirs has components of recharge, storage and discharge. The author listed the natural factors which influence discharge of water into a river; these include the distribution and infiltration characteristics of soils, the hydraulic characteristics and extent of the aquifers, the rate, frequency and amount of recharge, the evapotranspiration rates from the basin, distribution of vegetation types, topography and climate. He characterised these factors into those that affect gains and those that affect losses to stream flow. While these processes are included within models in various detail according to the scale and purpose of the exercise, there is some disagreement as to the relative importance of some of the processes. For example a clogging layer within a streambed is often not included in catchment scale investigations due to the high variability over the catchment.

Sophocleous *et al*. (1995) argued that incorporating clogging layers into models (analytical and numerical in this case) has a dramatic effect on the flux volumes in both gaining and losing stream systems.

Sophocleous (2002) classified interactions into three classes; (1) underflow (groundwater flux moves parallel to the river and in the same direction as stream flow), (2) baseflow (groundwater flux moves perpendicular to the river) and (3) mixed (combination of the above). The underflow component is predominant in systems with large channel gradients, small sinuosity's, large width to depth ratios and low river penetrations, in upstream and tributary reaches, and in valley fill depositional environments. Baseflow dominated systems occur under opposite conditions to underflow dominated systems. Mixed flow systems occur where lateral valley slopes are negligible. Braaten and Gates (2003) and Ivkovic (2009) classified rivers according to hydraulic connection and the dominant direction of flux, or water exchange between the surface and groundwater based on the descriptions of Winter *et al*. (1998). Banks *et al*. (2011) demonstrated that the state of connection can change along river reaches, as well as take place concurrently at the same location.

Xu *et al*. (2002) characterised South African rivers by their geomorphic features and then used this classification to derive the aquifer type as well as the type of boundary conditions one should take into account when conceptualising aquifer systems. In upper catchment areas, rivers are characterised by steep profiles, deep incision, inflow from valley sides in humid areas and large bed loads. Middle courses are characterised by bed load deposition and braided channels near mountains. Neotectonic uplift can create an incised convex profile downstream with riffle and pool sequences and in stable areas, meandering rivers. Lower courses are characterised by meanders on wide coastal plains, incision of old meanders where neotectonic uplift has occurred, allogenic rivers in the arid west parts of South Africa with deeper bed deposits and thicker terraces and deep infills in estuaries. This classification was then used to derive aquifer types and boundary conditions that should be taken into account and identified six 'typical' South African interaction scenarios:

- Upper catchment areas

  Type a: Streams without bank storage (e.g. braided rivers). Most likely to occur in mountainous areas. Interflow is likely and often a recharge area for groundwater.
- Middle courses

Type b: Streams controlled by bed morphology (e.g. pool and rifle sequences). This is frequently a gaining system, although transmission losses (losing system) can predominate in some areas.

- Lower courses

  Type c: Streams with bank storage (e.g. meandering rivers). Often in topographically flat areas near a regional base level. Fluvial erosion means these areas may become a bank storage buffer for groundwater.

  Type d: Streams influenced by channel morphology. Interaction with groundwater is at an intermediate scale. Both type c and d are generally gaining systems on a regional scale.

- Special cases

  Type e: Streams dictated by geological structures, especially those caused by neotectonic movements. Occurrence and interaction with groundwater is site specific.

  Type f: Streams with headwaters originating from allogenic source (those that originate and are fed from outside of the area, where precipitation and runoff are sufficient to generate flow), (e.g. Molopo River). This type often occurs in the drier western parts of the country, where streams are ephemeral.

A broad classification of the types of Interactions is detailed below.

2.2.2. Connected systems

Gaining streams

Within gaining stream systems, the piezometric surface must slope laterally towards the stream (effluent stream). Groundwater moves toward and emerges into the stream at all times. The piezometric surface at the stream is permanently above the stream stage and the material between it and the streambed is pervious – porous or fractured. The stream acts as a drain, is effluent and perennial (Vegter and Pitman, 2003). 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. Another important process in determining the volume of aquifer discharge into the stream is the capacity of the riparian strip zone bordering channel margins to evaporate the shallow groundwater. Where the depth to the groundwater is small adjacent to surface water systems, 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).

Losing streams

Within losing stream systems, the piezometric surface is at all times below the streambed level (Influent stream) and slopes downward away from the stream. This classification is characteristic of but, not necessarily limited to, ephemeral streams. The occurrence of transmission losses when the stream is flowing means that the stream recharges the groundwater system. The material between the streambed and piezometric surface is pervious. Transmission losses occur within two main environments over much of South Africa. These include the hard rock environment (river channels often follow lines of structural weakness and surface fracturing, offering an ideal opportunity for infiltration into the channel bed) and alluvial environments, where unconsolidated alluvial material underlies the river channel where losses can be substantial during both low flows and during the early phases of flood events (Smakhtin, 2001). Vegter and Pitman (2003) state that conditions over much of South Africa prevent rivers from being all but minor localised sources of recharge due to the fact that the water table generally follows the topography over the greater part of South Africa which inhibits the lateral expansion of the recharge mound that is being built up below the river by infiltrating water and lastly the fact that the rocky riverbeds and silty channels limit infiltration. Losses can also occur in to relatively dry soils forming the banks of streams which are enhanced by the presence of dense riparian vegetation promoting evapotranspiration. Information on channel losses or transmission losses is lacking although there have been detailed studies undertaken in some areas (Hughes and Sami, 1992; Lange, 2005; Morin *et al*., 2009).

Intermittent streams

The types of interactions observed between surface and groundwater change temporally and spatially in response to natural factors such as climate variability, and in response to anthropogenic factors such as river regulation and surface and groundwater abstraction, which affect the hydraulic gradients between the two systems (Ivkovic, 2009). In intermittent systems, it is assumed that groundwater emerges into streams at intervals such as immediately after recharge episodes. During dry periods groundwater storage is depleted by the effluent seepage or in combination with evapotranspiration from the stream banks and within the catchment. Groundwater may be replenished to a certain extent in the immediate vicinity of the stream by storm runoff. In the absence of rechargeable alluvial deposits and/or porous decomposed rock, replenishment from storm runoff would appear to be of minor importance compared to the volume of water recharged over the catchment area. Recharge from storm runoff can be restricted in its lateral extent as well as volumetrically by low storage capacity of aquifers, short duration of flow events and by low hydraulic heads. Vegter and Pitman (2003) argued that most intermittent streams in South Africa are underlain and bordered by alluvial deposits and /or porous decomposed rock. They assumed that, comparatively speaking, groundwater flow from the hard rock catchment toward the stream is of minor importance and in certain cases of no consequence at all and that the interaction between alluvium and stream is the dominant process. DWAF (2004a) make a distinction between intermittent and mixed systems. In intermittent systems the groundwater level is lower than the bed of the surface water body but depending on the elevation of the water level after recharge events, groundwater may recharge the surface water body. Mixed systems, on the other hand, alternatively lose water to sub-surface material during periods of high river stage and gain water from the same material during low flow periods. This process is commonly termed bank storage. Brunner *et al*. (2009b) discuss the spatial and temporal aspects of the transition from connection to disconnection between surface and groundwater. They based their analysis on numerical simulations and argued that the transition zone can be of significant extent which challenges the commonly made assumption that a system is either connected or disconnected (Wald *et al*., 1986; Covino and McGlynn, 2007). They found that a significant drop of the groundwater table may be required to change the flow regime from connected to disconnected. In addition they showed that in connected and transitional systems, a large transition zone results in an increased spatial variation in infiltration across the surface water body. They identified dimensionality and geometry as critical controls on disconnection with lakes disconnecting more easily than rivers.

2.2.3. Disconnected systems

Vegter and Pitman (2003) identified two scenarios within this classification. A detached stream (the intervening material is more or less impervious, very little or no recharge takes place) and a famished stream (groundwater does not reach the stream because it is permanently being dissipated along its flow path towards the stream by evapotranspiration). Brunner *et al*. (2011) detail disconnected or detached streams and point out that the term disconnected is frequently misunderstood or used in an incorrect way. Although conceptual illustrations of disconnected systems have been published many times, it is only within the last few years that the underlying physics of the disconnection process has been described. Knowing the state of connection is vital for sustainable water management because although lowering the water table beneath a disconnected section of a river will not change the infiltration rate at that point, it can increase the length of stream that is disconnected. According to Brunner *et al*. (2011), a disconnected system not only contributes to groundwater, but the infiltration rates are higher than in a losing, connected system. The authors list the diverse interpretations of a disconnected system (water table below streambed, water table depth greater than twice the stream width etc.) but argue that none of these relationships are correct because they neglect most of the important hydrological variables and do not define where the water table has to be measured. An unsaturated zone can be found in both intermittent and disconnected systems. However, unlike that in intermittent systems, changes in the level of the groundwater table do not affect the rate of infiltration from the stream to the aquifer for disconnected systems. DWAF (2004a) also make a distinction in disconnected systems between detached streams and streams with no connection. Detached streams lose water to the sub-surface material, such as to alluvial aquifers that are not in hydraulic connection with the regional groundwater body, hence do not recharge the regional aquifer. This occurs when the groundwater level is below the surface water level and the two do not influence each other due to impermeable material between the channel and the groundwater.

It is difficult to determine the relevance of the different gain or loss processes to the wide range of climatic, topographic and geological conditions which exist naturally. The literature which deals with experimental studies of low flow generating mechanisms often focuses on areas with shallow water tables which are rarely found in South Africa. Ideally the relative importance of low flow generation mechanisms and factors should be identified before low flow analysis is undertaken. The direction and magnitude of flux between groundwater and river systems will commonly vary along a river reach and depend on the timing and the scale of analysis. It is also important to recognise that the spatial scale used in surface water resource assessment modelling is typically moderate to large scale (100 to 1000's km^2) (Midgley at al., 1994). While it is important to understand that many of the processes that are involved operate at more detailed spatial scales, for the purposes of integrated water resource management, only the dominant processes which generally operate at larger scales need to be quantified. Ultimately the objective is to be able to interpret their impacts at the catchment scale. This is the same problem that faces modellers dealing with only surface water as surface catchment processes can also be highly spatially variable.

2.2.4. Review of anthropogenic factors affecting surface and groundwater interactions

Once a system begins to be exploited, 'natural' conditions are changed. The impacts are likely to include a reduction in overall discharge from the unit, a decrease in overall storage and a modification of overall recharge. Developments such as surface water and groundwater abstractions, the construction of storage dams, an increase in alien vegetation, forestry, mining, agriculture and activities in urban environments all impact the 'natural' system (Parsons, 2004). Storage dams can have a substantial impact on downstream flow volumes and patterns, water use and ecological functioning. The impacts are not only related to the number and size of dams, but also to the extent to which they are used for water supply as well as the nature of the climate and the natural hydrological regimes (Hughes and Mantel, 2010a). In addition, storage dams can induce extensive water logging through artificially recharging under-lying aquifers. Winter (1999) documented the reversal of the direction of groundwater flow resulting from the hydraulic head caused by a reservoir formed by the construction of a dam. Forestry is recognised as a stream flow reduction activity by abstracting water directly from the water table or from the unsaturated zone. This would result in a reduction of baseflow and hence a reduction in stream flow. Alien vegetation has a similar effect on the water balance (Leduc *et al*., 2001; Wittenberg, 2006; Kelbe and Germishuyse, 2010). Long term groundwater level rises in cleared areas can result from increased recharge rates (Leduc *et al*., 2001). This process has resulted in a major rise in salinity in dryland areas of Australia as the increased recharge 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). Groundwater is widely used for irrigation and stock watering purposes. This can result in over-abstraction of groundwater and water logging of the soil through over-irrigation. In addition agricultural use often affects water quality of surface and groundwater systems, through irrigation of saline soils and through the use of fertilizers. In order to carry out mining below the water table, groundwater has to be sealed off or abstracted. However, this impact is offset by the discharge of abstracted groundwater into streams, which in turn reduces the natural variability of stream flow and can significantly modify surface water quality (Parsons, 2004).

In a pre-development water balance, water levels in an aquifer may fluctuate from year to year but remain relatively stable over the long term, i.e. the recharge into a groundwater system equals the discharge out of the system (Sophocleous, 2002; Winter *et al*., 1998). 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 losses occur, there is a time lag between groundwater pumping and the maximisation of stream flow depletion.

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. The setting of a sustainably available quantity of groundwater is a contentious issue. Wright and Xu (2000) suggest that the volume of groundwater licensed to be abstracted could be set at a limit linked to the rate of effective recharge of the area under assessment, the maximum volume abstracted could be set to never surpass the rate of aquifer replenishment. However, they acknowledge that this approach does not explicitly consider the groundwater contribution to baseflow. There are scenarios where an abstraction does not reduce baseflow or intercept regional flow that would have discharged into the river. These include situations where the abstracted water would have discharged to the sea, or where the river is completely disconnected from the aquifer. In some scenarios, abstraction might not impact on stream flow but can result in reduced evapotranspiration due to lowering of the water table, or impact on downstream groundwater users (Rosewarne, 1991; Cleaver *et al*., 2003).

2.3. Geological framework

The classification of major geological environments has been undertaken by a number of authors (Issar and Passachier, 1990; Winter *et al*., 1998; Cook, 2003) Detailed below is a summary of the classification undertaken by Cook (2003) who categorised the major rock types into crystalline rocks, volcanic rocks, carbonate rocks, clastic formations and dykes and sills. Crystalline rocks include intrusive igneous rocks (e.g. granite, diorite, granodiorite, gabbro, dolerite and pegmatite) and metamorphic rocks (e.g. gneiss, quartzite, marble, schist, slate and phyllite). The intrusive igneous rocks tend to form either large intrusive bodies called plutons (e.g. granite, diorite and gabbro) or linear features like dykes and sills (e.g. dolerite, pegmatite). Climate, topography and rock structure are often more important in determining hydrogeological features like storage capacity and transmissivity than rock type. Clauser (1991) compiled data on hydraulic conductivity of crystalline rocks and at the borehole scale the transmissivities generally range between 10^{-2} and 10^{-7} m^2 day^{-1}.

When magma solidifies at or near the ground surface it forms volcanic rock. Lava can be categorised as acidic lava (e.g. rhyolite, dacite and andesite) or basic lava depending on its mineral content. Acidic lava is viscous and therefore forms steep sided domes, while basic lava (basalt) has lower viscosity and can spread extensively (a few tens of metres to hundreds of kilometres) forming thin (less than 1 m to more than 30 m) sheets. Hydraulic conductivity in basalts can form vertically if it cools rapidly to form hexagonal columnar jointing and horizontally from brecciated zones or interbedded pyroclastic deposits. These can form local confined aquifers or perched aquifers underlain by denser basalt units or intruded sills. The storage capacity and transmissivity of volcanic rocks depends on the rate of cooling, the extent of degassing while cooling and on the viscosity of the magma. Total porosity is often around 15% but can range from near to zero to 75% (Wood and Fernandez, 1988).

Carbonate rocks are classified as sedimentary rocks which contain more than 50% carbonate minerals, usually calcite ($CaCO_3$) and dolomite ($CaMg(CO_3)_2$). The term limestone is used for those rocks that contain more than 90% carbonate with calcite as the dominant mineral. If the rock contains between 50 and 90% carbonate it is called arenaceous limestone or argillaceous limestone depending on the relative amounts of quartz and clay minerals. Dolomite clearly has a high amount of dolomite minerals within the rock while chalk is the term used for a fine grained limestone Carbonate rocks are soluble in water rich in carbonic acid which can lead to large conduits for groundwater flow and the development of karst topography. Porosity in carbonate rocks can vary from less than 1% in marbles and some massive limestones to as high as 45% in some chalks and calcareous tuffs. Dolomites are usually more porous than limestones due to reduction in volume as a result of dolomitisation (the replacement of calcite by dolomite). While chalks and some limestones have high porosity, the pores are often small and so primary permeability is low (Greswell *et al.*, 1998). However the rock may acquire high secondary porosity and permeability (Bishop and Lloyd, 1990).

Coarse grained unconsolidated sediments usually have a high hydraulic conductivity and porosity. They can form secondary porosity (fracturing) if the unconsolidated rock has been well cemented. The hydraulic properties of sandstones depend on their textural characteristics which are dependent on the depositional environment together with post depositional changes such as cementation, consolidation and fracturing. Finer grained argillaceous rocks (e.g. shales, siltstone) are formed by compaction and lithification of clay and mud deposits. Shales usually have porosities in the range of 1 to 3%. Intergranular permeabilities in shales and siltstone are usually low (between 10^{-13} and 10^{-9} m s^{-1}), although fracturing can significantly increase hydraulic conductivity. Singhal and Gupta (1999) suggest typical hydraulic conductivities of fractured siltstones and shales of 10^{-7} and 10^{-4} m s^{-1}. In heterogeneous sedimentary environments, thin incompetent beds such as shales will be more intensely fractured compared with thicker units of strong and resistant formations such as mudstones.

Dykes are vertical or steeply inclined intrusive igneous bodies that cut across the bedding planes of pre-existing rocks. They vary in thickness from a few decimetres to hundreds of metres, with widths of 1 – 10 m being most common. They may be from a few metres to several kilometres long and represent feeders for lava flows. Massive and unweathered dykes can form barriers to groundwater movement (Engel *et al*., 1989). Fractured dykes can also form good aquifers. Due to thermal effects, dykes can also cause fracturing of adjacent rock (Sami, 1996). Sills are nearly horizontal tabular bodies that commonly follow the bedding of enclosing sedimentary rocks or lava flows. Some sills can be extremely thick and extend over large areas. Due to their low permeability without fracturing, sills may cause the formation of perched water tables.

The geology of South Africa is shown in Figure 2-1. The effective description of groundwater occurrence in a large geologically complex area such as South Africa requires sub-areas with similar characteristics to be delineated. Authors that have attempted to classify South Africa’s aquifers include DWAF (2003a), MacDonald and Davies (2000) and Vegter (2000). Detailed below is a summary of the classification undertaken by Le Maitre and Colvin (2008). The primary lithologies were grouped into six broad principal aquifer types based on the type of permeability and groundwater flow and discharge regimes found in the country. These principle aquifer types were based on the hydrogeological provinces described by Issar and Passachier (1990). This original classification was modified by Le Maitre and Colvin (2008) and is detailed in Table 2-1.

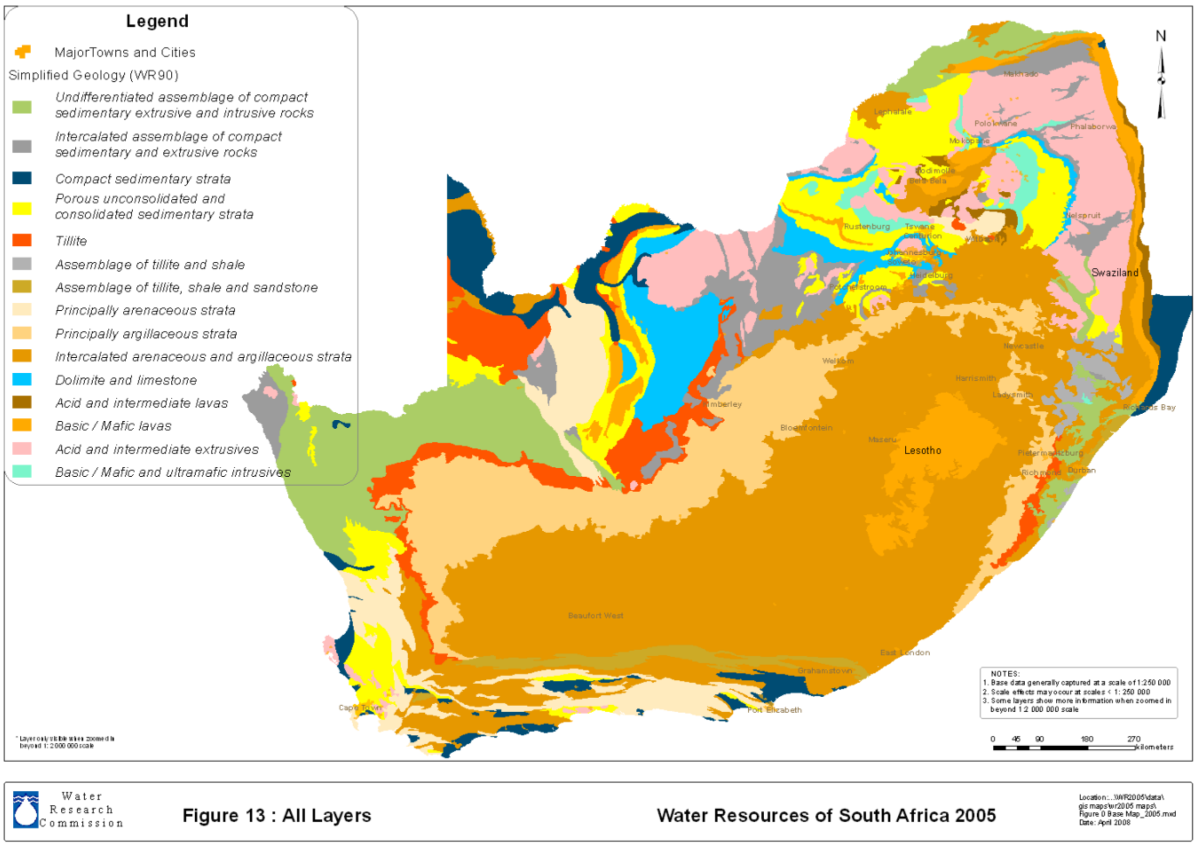


Figure 2-1 The simplified geology of South Africa (Bailey, 2007).

The largest of the six principle aquifer types is associated with fractured sedimentary rock, covering 54.6% of the land area in South Africa, Swaziland and Lesotho and dominated by the shales and sandstones of the Karoo Supergroup. The second largest aquifer type consists of unconsolidated deposits covering 17.5% of the land area. Unconsolidated aquifers are found in the interior where alluvial sediments have accumulated due to the low relief, especially in the Kalahari basin as well as marine and aeolian deposits along the southwest and northeast coasts. The basement complex and younger granites (11.9%) contain secondary aquifers in the weathered zone and permeable structures. The extrusive igneous rocks (8.7%) occur patchily across the country and only cover a large area in the Maloti-Drakensburg Mountains and along the north eastern border. Karoo dykes and sills (4.6%) and carbonates (2.7%) are the least extensive although the carbonate systems are the most productive of all the aquifer types.

Table 2-1 Summary of the properties of the 6 principle aquifer types based on hydrogeological terrains and their influence on the nature of groundwater discharge to rivers (Le Maitre and Colvin, 2008).

Principal aquifer type	Rock types	Aquifer characteristics	Nature of discharge to rivers
Basement complex and younger granites	Granites, gneisses, green-stones and similar rocks	Secondary aquifer; limited water storage mainly in the regolith (weathered material), or in faults and fractures	Limited volumes of groundwater and discrete discharge to rivers but can be sustained through springs or localised discharges where there are significant fractures or (major) faults
Carbonates	Dolomites and limestones	Secondary aquifer; water storage is in the solution cavities created by dissolving the carbonates (high permeability and storativity); often compartmentalised by impermeable dykes	Large volumes of water and sustained flows in the dolomites and in some situations in the limestones; typically discharged through eyes, springs and wetlands
Extrusives	Basalts and similar rocks from outflows of lava	Secondary aquifer; limited water storage mainly in the regolith zones or in faults and fractures; some permeability in vesicles	Very limited volumes of groundwater and dis-charge to rivers but can be sustained through springs or localised discharges where there are significant fractures or (major) faults
Fractured meta-sedimentary	Sedimentary rocks which have been fractured, intruded and metamorphosed to varying degrees; e.g. shales and sandstones of the Karoo group (covering much of the interior) and the	Secondary aquifer, the water is found in fractures, including bedding planes, joints and faults at multiple scales; sometimes limited to weathered material	Discrete and linear discharge from the fractures and faults and from contact zones between the sandstones and the interbedded shale layers or the underlying basement rocks; evident as springs, seeps and wetlands but often hidden as rivers are frequently sited on major structures. Wide range of permeability and storativity depending on lithology and structural history.

	sandstones of the Table Mountain Group		
Karoo dykes and sills	Intrusive lavas forming dykes (linear, roughly vertical) and sills (near horizontal)	Secondary aquifer; limited water storage mainly in the chill zones and fractures; and limited subsequent weathering.	Variable volumes of water which usually emerge as springs, seeps or wetlands, generally associated with Karoo rocks with limited groundwater storativity
Unconsolidated deposits	Formed from aeolian, alluvial or colluvial material; sometimes from marine deposits; often extensive (e.g. Kalahari, Zululand); along lower parts of many river systems.	Primary aquifer with inter-granular permeability and some secondary permeability associated with crete lenses and bioturbations; variable permeability from coarse sands and gravels to finer material and clays.	Diffuse discharge from primary aquifers with moderate to high storativity. In areas with low gradients the groundwater is recharged during high river flows or floods and then released to sustain flows once river water levels drop below the water table; water can be stored in floodplain aquifers where it is accessible to floodplain plant communities such as gallery forests; groundwater often present beneath the river bed when there is no surface flow

2.4. 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. Any interaction environment becomes more complex as the scale of investigation becomes more detailed. The question is, can we reliably quantify the processes occurring, at a scale large enough to contribute to integrated water resource management.

2.4.1. 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 (Bear *et al*., 1993; 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, and the field can be considered in many respects to still be underdeveloped. Part of the problem lies in defining the type of information required, on the basis of realistic expectations. The inherent structural and hydraulic complexity of fractured rock severely limits the type and quality of data that can be obtained from field measurements and ultimately, the level of uncertainty in model prediction. An important consideration when beginning an investigation is that the problem of interest, its measurements and interpretation, are a function of scale. Therefore before the stages of 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).

Local and regional geological activity such as folding and faulting lead to an irregular distribution of fractures and zones of high and low fracture density usually develop. The orientation, connectivity characteristics and variations in the density of the fracture zones are expected to have strong influences on the movement (direction and rate) of groundwater in both the saturated zone (the aquifer) and the generally unsaturated zone through which recharging water has to pass. The importance of preferential flow in controlling recharge is shown by many studies in Botswana and South Africa (both areas dominated by fractured rocks). Large scale recharge investigations have been conducted in central and eastern Botswana (Selaolo *et al*., 1996; de Vries *et al*., 2000).

The properties of fractured rock aquifers can vary substantially between regions due to differences in rock types and varying tectonic stresses. For example, boreholes drilled in the Table Mountain Group (TMG) Sandstone in South Africa (which has undergone intense folding) are high yielding (Roets *et al*., 2008; Parsons, 2009), while wells in the Karoo Formation (relatively flat lying meta-sediments), also in South Africa, have extremely low yields (Sami, 1996). Fractured zones around intrusive features such as dykes and sills often form good aquifers. Due to thermal effects, dykes can also cause fracturing of adjacent rock. Sami (1996) found the yield of boreholes adjacent to dolerite dykes intruding the fractured sandstone/mudstone Karoo Aquifer, to be significantly higher than elsewhere in the basin. Pumping tests in Botswana indicate that dykes which are thicker than 10 m serve as groundwater barriers, but those of smaller width are permeable as they develop cooling joints and fractures (Buckley and Zeil, 1984).

Aerial photography, remote sensing data, published and field geological maps, borehole and tunnel fracture studies, data from tracer, cross-hole and geophysical tests as well as other standard hydrogeological analyses can provide inferences on fracture set properties. Field evidence of fracture zones in bedrock exposures and the occurrence of frequent intrusions can be used to infer degrees of fracturing, as can information on folding and faulting. However, there is typically not enough information available to characterise fractured rock aquifers from a hydrogeological perspective, despite the fact that existing geological and hydrogeological maps can be used to identify their existence. Groundwater discharge from fractured rock aquifers can be estimated by measuring the discharge of streams that drain fractured rock catchments (Misstear and Fitzsimmons, 2007), or from measuring concentrations of solutes within streams and applying solute mass balance methods (Abiye *et al*., 2011). However, the spatial variability of groundwater inflows may be increased by water inputs from irregularly spaced fractures (Cook, 2003). In environments where the stream flows directly on the fractured bedrock, groundwater discharge to streams may occur only in discrete locations associated with individual fractures (Oxtobee and Novakowski, 2002). In such settings, stable isotopes and vegetation can be used to help locate discharge points (Levy and Xu, 2012). Oxtobee and Novakowski (2002) quantified the interactions between surface and groundwater in a fractured rock environment using air-photo interpretation, detailed stream surveys, electrical conductivity and temperature surveys, isotopic analysis, mixing calculations and point measurements of hydraulic head and discharge obtained using mini-piezometers, seepage meters and weirs. They identified that the groundwater contribution to baseflow occurred through discrete point sources associated with the exposure of fractures. Similarly Sami and Hughes (1996) compared groundwater recharge simulated by a conceptual model (VTI model) with mean annual recharge rates derived from a chloride mass balance for fractured sedimentary rocks in the Karoo Basin, South Africa. Although the model performed well, the results suggested that recharge is highly variable in space and that at local sub-area scales the comparison was less reliable. Banks *et al*. (2011) undertook a detailed regional investigation using a combination of hydraulic, hydrochemical and tracer based techniques and produced a detailed perceptual model of the surface and groundwater interactions over an entire river system (headwaters to the sea). The study demonstrated the dominance of hydrogeological and hydroclimatic controls on the state of connection in a catchment and also highlighted the complexities inherent in fractured rock environments. Although the spatial variability of groundwater discharge into streams is a process that must be examined in any investigation, at the larger scale at which water is managed, it may not be necessary to specifically identify individual inflow (recharge) and outflow (groundwater baseflow) points. The smaller the scale of the investigation the more important it becomes to characterise the fractures and faults. While any number of field techniques can be used to characterise hard rock aquifers, pumping tests are often the preferred technique for gathering hydrogeological data in these environments (Krásný and Sharp, Jr, 2007).

Almost 98% of aquifers in South Africa are classified as secondary aquifers (Parsons, 2004). The most widespread hard rock aquifer type in South Africa is the fractured metasedimentary type which incorporates 55% of South Africa's area (Le Maitre and Colvin, 2008). This aquifer type is dominated by the shales and sandstones of the Karoo Supergroup and the quartzites of the Table Mountain Group (generally believed to hold greater volumes of groundwater than the Karoo Supergroup). As such, studies have been carried out on both aquifer types (van Tonder and Kirchner, 1990; Sami and Hughes, 1996; van Tonder *et al*, 2001; DWAF, 2008a; Parsons, 2009; Roets *et al*., 2008; Botha and Cloot, 2004). South African studies are generally carried out with far less data than the previously mentioned international investigations. Conrad and Adams (2007) used GIS to assess recharge in fractured rocks over South Africa using limited data sets. Issar and Kotze (2007) used environmental isotopes to establish a hydrogeological perceptual model for flow in the fractured quartzite rocks of the Table Mountain Group. Botha and van Rooy (2001) used geological mapping, geophysical methods and pump tests to characterise the Nebo Granite aquifer in the Northern Province of South Africa.

Particular challenges faced during the characterisation of these aquifers include locating suitable structures, estimating possible recharge and analysing the aquifer response to pumping. Voss (2003), for this reason, suggested that *a priori* characterisation of fracture systems might be impossible. It has been suggested (Clauser, 1991; 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 background that corresponds with a representative storage, the smallest scale above which practically no change in mean values occurs (although high permeability zones such as dolorite dyke intrusions and regional fault zone structures should be borne in mind (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.

2.4.2. Alluvial aquifers

The interaction processes within alluvial aquifers can be complex and can take the form of both groundwater contributions to stream flow and surface water contributions to groundwater (common in arid and semi-arid regions). The alluvium can be recharged by rain, surface water runoff, spring flow, flood recharge from rivers or by groundwater from the surrounding geology. 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 (Dixon and Chiswell, 1992; Hughes and Sami, 1992; Subyani, 2004) while others have investigated scenarios where the alluvial aquifers recharge the stream (Negral *et al*., 2003; DWAF, 2008b). Studies documenting complex interaction scenarios within alluvium (Sophocleous, 1992; Morrice *et al*., 1998; Osman and Bruen, 2002; Tooth *et al*., 2002) have highlighted the importance of a sound perceptual model as well as the danger of oversimplifying the interaction processes.

If the alluvial aquifer is dry (in ephemeral rivers), attenuation of channel flow can occur, due to infiltration through the river bed and banks (transmission losses). 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 the flow would occur through recharge of the aquifer, 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.

De Vries and Simmers (2002) carried out an overview of types, processes and estimation of recharge in dry climates and concluded that quantitative estimates from all principal recharge mechanisms are very uncertain. However, it is generally accepted that indirect recharge by transmission losses increase with aridity (Lloyd, 1986; De Vries and Simmers, 2002). Owen (1994) found that for lateral plain aquifers, recharge depends on the permeability of the aquifer, the distance from the channel and the duration of river flow. The alluvial aquifers in the Mzingwane catchment in Zimbabwe are recharged principally from stream flow and Moyce *et al*. (2006) found that no stream flow occurs until the channel aquifer is saturated (normally early in the rainy season). Walters (1990) determined that the proportion of flow lost from the upstream hydrograph varies significantly over different areas. His study areas varied from negligible over a 32 km reach (Texas) to more than 90% over a 12 km reach (Arizona). Small scale studies (using point source measurements) have been carried out in alluvial aquifers examining the effects of the geological composition of alluvium (Morrice *et al*., 1997) and clogging layer effects using MODFLOW (Osman and Bruen, 2002). Although detailed process studies are often unable to contribute significantly to improving large scale process understanding for water resources research, they do highlight the potential complexity of alluvial aquifers which needs to be taken into account when constructing accurate perceptual models. Tooth *et al*. (2002) characterised the geomorphology of the Nyl River and floodplain in the semi-arid Northern Province of South Africa. The summer flooding occurs primarily as sheetflow which deposits clay layers which seal the floodplain surface, prolonging inundation and limiting groundwater recharge to the floodplain margins.

Channel losses into the alluvium can be estimated by measuring discharge at two points in the channel system. If hydrometric data are available upstream and downstream of a channel reach, inflow and outflow volumes may be compared and transmission losses quantified. Lange (2005) suggested that volumes of transmission loss may then be related to flow and channel characteristics by means of regression analysis, although this approach can be complicated by unknown lateral inflows. Lange (2005) assessed the temporal dynamics of transmission losses in the Kuiseb River in Namibia, not by using a traditional water balance study but by applying a mathematical flow routing scheme with parameters not calibrated with runoff information. Although the losses were not quantified, 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 based on kinematic flow with components accounting for channel bed infiltration. Most of the model parameters were obtained from field surveys (not detailed in the paper) and GIS analyses. 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). Alluvium can cause interpretive difficulties in connection with hydrograph separation (Chen *et al.*, 2006) procedures. There have been authors, however, that have derived techniques for estimating the effects on the downstream hydrograph (Hisdal and Tallaksen., 2003; Kurt, 2007; Aksoy *et al*., 2008). Neglecting transmission losses in any modelling system designed to simulate runoff characteristics from semi-arid or arid catchments could result in serious errors, such as an over estimation of the surface runoff from catchments with similar characteristics. There is a need for further documentation and understanding of the nature of transmission losses in different environments, as well as the incorporation of this understanding in water resource estimation techniques (Hughes and Sami, 1992).

2.4.3. Karst aquifers

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 (Atkinson, 1977; Larocque *et al*., 1998). Karst systems are often characterised by substantial surface and groundwater interactions via processes such as aquifer recharge by losing streams (Larocque *et al*., 1998), fracture and conduit connections between surface and groundwater and spring flow contributions to surface water (Katz *et al*., 1997; Swart *et al*., 2003). Although much research has been carried out on karst aquifers (White, 1993, 1998, 2002), a thorough understanding of surface and groundwater interactions in karst environments is lacking (Musgrove *et al*., 2010). Karst aquifer systems often respond rapidly to changes in environmental and climatic conditions (Malard and Chapuis, 1995; Mahler and Massei, 2007). 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 preferable when managing the system in an integrated fashion.

Investigations into the surface and groundwater interaction processes of dolomite aquifers have utilised different methods including conceptual modelling (Le Moine at al., 2008), numerical modelling (Swart *et al*., 2003; Hughes *et al*., 2008), exploration drilling (Kafri and Foster, 1989), pump tests (Winde and Erasmus, 2011) and upstream and downstream flow gauge measurements and temperature time series (Bailly-Comte *et al*., 2009). Variations in spring discharge and spring water compositions have been used to characterise karst environments and investigate the processes that control groundwater quality (Wong *et al*., 2012). Le Moine *et al*. (2008) successfully simulated the La Rochefoucauld-Touvre karstic system in France using a lumped rainfall-runoff model (GR4J) and argue that a simple conceptual model has advantages over more detailed numerical models in karst systems. The GR4J model incorporates intercatchment (regional) groundwater flow (common in many dolomite aquifers) and a lag and route function, both of which aided in improving the simulation of the catchment.

A main feature of the South African dolomites is that they are subdivided into smaller compartments by thin vertical dyke intrusions, which act as barriers to the lateral drainage of groundwater. This gives rise to the occurrence of springs ('eyes') at the topographical lowest point where these dykes force the groundwater to overflow as springs. Most of the dolomitic eyes in South Africa are dammed springs that are formed when dykes dam the groundwater in the aquifer resulting in rising groundwater levels (Kafri and Foster, 1989 and 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). Most of the literature on dolomitic aquifers in South Africa is focused on the geological origins of the dolomite or the engineering challenges faced during construction on the formations (van Rooy and Witthuser, 2008; Hobbs, 2008). The major outflow points ('Eyes') from dolomitic aquifers in South Africa are also fairly well documented as they have represented reliable sources of water for the local communities for many years (Buttrick *et al*., 1993; Witthuser and Holland, 2008). Although, the extensive water use due to the dewatering of the compartments by mining operations and subsequent release of large quantities of waste water into downstream dolomitic aquifers, has resulted in difficulty in formulating definite conclusions from the data available. Comprehensive geological and hydrogeological maps exist for South Africa which indicate the locations of dolomite. It is difficult to derive the depth of an aquifer and it is usually obtained by examining drilling logs of boreholes and the deepest water strikes.

Kafri and Foster (1989) carried out a field investigation examining the hydrogeology of the Malmani dolomite in the Klip River and Natalspruit basins of South Africa. They examined existing borehole data, and carried out geological mapping and exploration drilling. They found complex aquifers extensively compartmentalised by dolerite dykes with numerous perched aquifers lying above dolerite sills. Although groundwater levels in the different compartments were variable, all appeared to be in hydrological connection with the Klip River and regionally the groundwater flow converged on the main stream and southwards toward the Vaal River (Stream and groundwater levels were similar throughout). There is very little surface runoff in the catchment (the authors estimated 5% of MAP), while recharge of the aquifers was highly uncertain and estimates ranged between 2.5 – 28% of MAP). Swart *et al*. (2003) encountered very similar characteristics in the Chuniespoort Group dolomite on the Far West Rand, South Africa. They used data from the mines (responsible for extensive dewatering and perforation of the dolerite dykes in order to hydraulically connect the aquifer compartments) and a geohydrological model to assess the future of the numerous dolomitic springs once the dewatering ceased. They concluded that the dry springs will resume flow once the water levels recover even with the now hydraulically connected aquifer compartments.

2.4.4. Primary aquifers

Primary aquifers can include a wide variety of aquifers and contributions to the field are numerous. Primary aquifer types include sand aquifers (coastal, dune, lowland floodplains etc.) (Delin and Landon, 2002; Trojan *et al*., 2003; Kelbe and Germishuyse, 2010), chalk aquifers (Darling and Bath, 1988) and glacial deposits (Waddington *et al*., 1993; Anderson, 2004; 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 (Sudicky, 1986; Dagon, 1989; Killey *et al*., 1991; Le Blanc *et al*., 1991), and the potential heterogeneity of many of these aquifers (Dagon, 1990; Ptak and Teutsch, 1994; Schad and Teutsch, 1994). 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. There still remains a high degree of uncertainty when simulating catchments dominated by primary aquifers. 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 extend over several orders of magnitude (Rehfeldt *et al*., 1989; Schad and Teutsch, 1991). Lateral flow such as interflow as well as perched aquifers can be present due to clay lenses and macropores present within the aquifer.

It is more straightforward to characterise recharge rates into primary aquifers than fractured rock environments, although the presence of macropores can cause a degree of preferential flow. Recharge estimates in northern Senegal based on chloride data indicated that recharge rates are highest where Quaternary sands are thickest and decrease to lower values where finer textured soils occur (Gaye and Edmunds, 1996). In Australia, Allison *et al*. (1990) estimated rates of recharge into primary aquifers rich in clay at 1 to 9 mm year^{-1} and in sandy aquifers recharge rates were from 1 to 50 mm year^{-1}. Pumping tests are traditionally used for the determination of average or effective values of hydraulic parameters, such as transmissivity and storativity. However, in heterogeneous porous aquifers, the interpretation of the values inferred from pumping test data may vary with time and space.

In South Africa, there are three significant coastal aquifers; these are the Cape Flats, Langebaan and the Zululand/Mozambique aquifer (Steyl and Dennis, 2010). Generally the aquifers on the west coast (Cape Flats, Langebaan and numerous smaller primary aquifers) consist of course grained, clean sand while the sub-tropical east coast aquifers (Zululand/Mozambique aquifer) are composed of sand, silt and clay layers. The Cape Flats aquifer extends for 630 km^2 and is an important source of water for Cape Town, although it is heavily polluted in some areas. Recharge estimates have varied between 40% (Gerber, 1980) and 15 to 35% (Vandoolaeghe, 1989), both using the water balance method. Conrad *et al*. (2004) carried out a review of the recharge estimates for all the primary aquifers of the west coast and concluded that a vertical recharge figure of 8% is conservative. According to Conrad *et al*. (2004), the majority of the groundwater recharge is from underlying and adjacent fractured rock aquifers. Artificial recharge is carried out in some of the aquifers to promote improved sustainable development (Wright and du Toit, 1996). The authors describe the ‘typical’ west coast primary aquifers as being high yielding, with variable aquifer thickness. The aquifers are characterised by shallow groundwater tables which mimic the topography (although the surface and groundwater catchments do not always coincide). There are springs throughout the area and although the groundwater levels are high, they drop below the river level in the summer months and interaction ceases (seasonal river) (Conrad *et al*., 2004 and DWAF, 2008a).

The Zululand aquifer extends for 7000 km^2 along 1250 km of coastline. Although much of the aquifer falls within the protected St Lucia wetland park, forestry activities within the region has caused a reduction in water levels towards the north of the aquifer (Steyl and Dennis, 2010). Recharge to the east coast aquifers is predominantly from rainfall (Rawlins and Kelbe, 1991). Methods utilised to investigate surface and groundwater interactions were detailed in a series of case studies carried out by the University of Zululand. These methods included the monitoring of groundwater levels, numerical modelling (MODFLOW) and hydrograph analysis techniques (Kelbe and Germishuyse, 2010).

2.4.5. Unsaturated zone processes

Unsaturated zone water is defined in this thesis 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. This water can also be termed interflow which is often associated with emerging water from the unsaturated zone such as soil water which contributes to baseflow (seeps), springs formed by the emergence of water before it reaches the regional groundwater (from perched water tables, the lateral orientation of fracture zones or from the lateral movement of percolating water which encounters a less permeable layer) (Figure 2-2). Contributions from the unsaturated zone can make up substantial portions of baseflow in many catchments. Groundwater abstraction may not impact at all on unsaturated zone water from springs, seeps, perched water tables and interflow (DWAF, 2004a) unless induced recharge through the lowering of the groundwater table represents a major factor (Bromley *et al*., 2001; Lia and Xia, 2004).

It is not straightforward to investigate processes in the unsaturated zone and there is therefore, a lack of information available on the processes which take place (Scanlon *et al*., 2006). This is particularly true in arid and semi-arid environments where the distance between surface water and groundwater levels can be high and the interactions difficult to observe. Many traditional perceptual models envisage the unsaturated zone as a small area between the surface and shallow groundwater table (Landon *et al*., 1999; Seibert *et al*., 2003) and while this is correct in many parts of the world, in semi-arid or arid regions the unsaturated zone can be hundreds of metres thick (Izbicki *et al*., 2000; de Vries *et al*., 2000). Consequently, many assumptions based on traditional perceptual models of interaction regarding the processes which occur in the unsaturated zone are incorrect in these environments.

Interflow can occur in different ways which include:

- the drainage of near channel storages such as channel bank soils, alluvium and wetland areas (Smakhtin, 2001).
- Perched water table occurring above a zone of reduced percolation (Figure 2-2A).
- Favourable geometric configuration of fractured networks may lead to formulation of interflow (Figure 2-2B).
- Perched water table occurring within a zone of localised increased lateral conductivity.

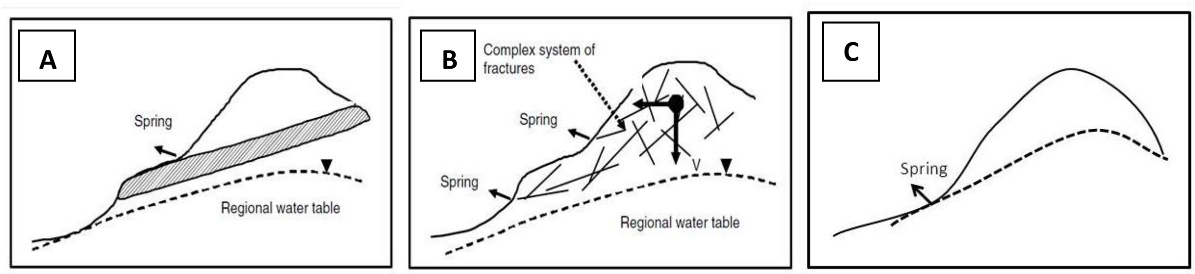


Figure 2-2 A shows a spring resulting from a confining layer; B illustrates the scenario where springs result from a lateral component to otherwise percolating recharge water; C illustrates the scenario where springs result from the intersection of the regional groundwater table with the surface (adapted from Hughes, 2010a).

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 but was removed by vapour transport while infiltrating water from intermittent stream flow (in the floodplain area) took 180 to 260 years to reach the regional water table (clay layers impeded the downward movement of water). According to Scanlon *et al*. (2006) unsaturated zone flow in arid and semi-arid environments is frequently highly spatially variable and water can take from less than a year to thousands of years to reach the water table. As a result the authors reinforced the need to use careful consideration of spatial and temporal scales in selecting approaches used to characterise unsaturated zone processes.

Unsaturated storage zones often recharge during and soon after precipitation events. Springs most often occur in steeply sloping terrain and depending on the type of spring, can account for prolonged baseflow following rainfall events. 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 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. 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).

Vegetation cover is also important to be aware of as it largely determines the proportion of rainfall that reaches the soil and may also influence infiltration, percolation and deep drainage, and the available storage capacity of soil profiles (Le Maitre *et al*., 1999). It is essential that the processes occurring in the unsaturated zone are included in a perceptual model formulated for any environment, as these processes can form a major component of the water balance. Seibert *et al*. (2003) argue that many models do not fully address interactions between unsaturated zone storage and saturated zone storage which can lead to unrealistic simulations.

2.4.6. 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, 2005a) 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 (Toth, 1963; 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 2-3). 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.

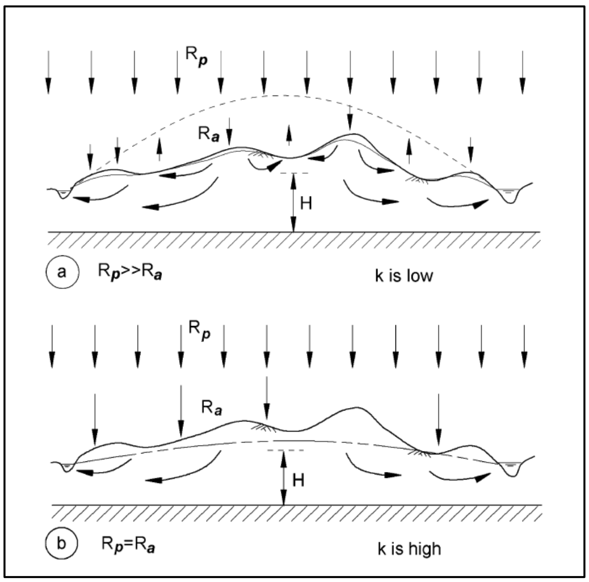


Figure 2-3 (a) Low permeable aquifer with topography controlled water table; (b) highly permeable aquifer with a recharge controlled water table (Haitjema and Mitchell-Bruker, 2005).

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 and required by the NWA. 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 authors also stated that in their experience, the majority of groundwater assessments show an acceptable correlation between groundwater head elevation and topography. They stated that 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.

2.4.7. 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 increasing anthropogenic impacts downstream masking 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 it 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 large, 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 includes how one can upscale local information on soils, vegetation, groundwater and surface and groundwater interactions to the scale of a surface catchment, and whether catchment scale modelling of surface and groundwater interactions can yield reliable results.

2.5. Observations and analysis

2.5.1. Introduction

There are numerous direct and indirect methods (hydrological, hydrogeological, tracer, GIS-based and geophysics based) available to investigate, characterise and quantify the interactions between surface and groundwater. Each method will have its strengths and limitations. Kalbus *et al.* (2006), provide an overview of the field methods typically used in surface and groundwater interaction investigations. Considerations for choosing appropriate methods are given including spatial and temporal scales, uncertainties and limitations in application. The methods differ in resolution, sampled volume, and time scales (Table 2-2). The authors concluded that a multi-scale approach combining multiple techniques can considerably reduce uncertainties and constrain estimates of fluxes between surface and groundwater. The authors state that inaccuracies are inherent in all methods to determine interactions between surface and groundwater, so that an analysis of uncertainties along with any measurement is indispensible. Levy and Xu (2012) provide South African examples of investigations utilising the various methods. Due to very few experimental catchments in South Africa in comparison to many developed countries (North America, Europe and Australia), the most common approach is to estimate average annual fluxes at the surface water catchment scale with baseflow separation techniques and then subtract the groundwater discharge rate from the recharge rate. This approach, while useful, ignores spatial and temporal variability and cannot contribute to improved process understanding.

While experimental catchments are very valuable in improving process understanding and the modelling thereof, they are very expensive to establish and maintain. In addition, most process based research in hydrology has focused on short time scales in small catchments (Sidle, 2006) reflecting the logistical issues associated with setting up large scale experimental catchments. Issues limiting this type of research include the lack of tools that allow processes to be extrapolated from point scales to larger catchment scales and the fact that larger catchments have increasing anthropogenic impacts downstream, thereby masking natural variability (Tetzlaff and Soulsby, 2008). Consequently, there are few examples of these catchments that have been designed for the purpose of model development or testing in developing countries such as South Africa. In some situations, quantifying appropriate parameter values for the models is relatively straightforward while for others field based data is required to differentiate between the processes that dominate the patterns of low flow in rivers. It is not straightforward to design an experimental catchment that produces data useful for assessing interactions at the larger scales appropriate for water resource management (Sidle, 2006).

The type of data obtained from experimental catchments can vary considerably (Levy and Xu, 2012) and it is difficult to upscale to the scale at which water is allocated and managed. Experimental data alone is rarely useful for regional scale assessments and often have to be combined with other methods or models. Hence, the validation of models using experimental data is extremely uncertain. Catchment and regional scale assessments (Tetzlaff and Soulsby, 2008) are used to support management of surface and groundwater resources, while more detailed field investigations (Seyler *et al*., 2009) are required for site specific details to support, for example, the siting of boreholes. It is difficult to generalise about interaction between surface and groundwater at a river or catchment scale (Parsons, 2004). However, methods need to be developed that can be used for integrated water resource development that could then be supported, where necessary, with more detailed modelling and/or data collection.

Table 2-2 **Spatial scales of the different methods available to measure interactions between surface and groundwater. The spatial scale is given as a radius or distance of influence in metres. Pm represents point measurements. Adapted from Kalbus *et al*. (2006) and Levy and Xu (2012).**

	parameter	method	Appropriate for estimates in discrete locations and/or fractured bedrock settings	Allows quantification of exchange	Spatial measuring scale				
					pm	1	10	10^2	10^3
	seepage flux	seepage meters	Yes	Yes	■				
	temperature gradient	temperature profiles	Yes	Yes	■				
		vegetation mapping	Yes	No		■	■		
Methods based on Darcy's Law	hydraulic head	piezometers	No	Yes	■				
	hydraulic conductivity	grain size analyses	No	Yes	■				
		permeameter tests	No	Yes	■				
		slug tests	No	Yes	■	■			
		pumping tests	No	Yes	■	■	■	■	■
	porosity	sediment sample analyses	No	Yes	■				
	groundwater velocity	tracer tests	No	Yes	■	■	■		
Mass balance approaches	groundwater component	incremental stream flow	No	Yes			■	■	■
		Upstream/downstream flow measurements	No	*Yes*			■	■	■
		hydrograph separation	No	Yes				■	■
		Geochemistry and temperature	Yes	Yes			■	■	■
		Stable isotopes ^2H and ^{18}O	Yes	Yes			■	■	■
		heat tracer	Yes	Yes		■	■	■	
Modelling		Conceptual modelling	Possibly	Yes				■	■
		Numerical modelling	Yes	Yes		■	■	■	■

Regional scale groundwater mapping, while not simple, is essential if integrated water resource management on national scales is to be achieved. Countries such as Korea (Lee *et al*., 2007), Taiwan (Shiang-Kueen, 1998), Mexico (Carrera-Hernandez and Gaskin, 2007) and the United States

(Sophocleous, 1992; Dumouchelle and Schiefer, 2002; Szilagyi *et al*., 2005; Sophocleous, 2010) have undertaken this type of mapping successfully. South Africa has produced a national groundwater database (DWAF, 2005a) which provides estimates of storativity, transmissivity, maximum depth to groundwater and annual recharge at a spatial scale (approximately 100 to 1000 km^2) appropriate for typical hydrological model applications. This database has contributed to beginning the integration process between surface and groundwater (which were previously managed separately); however, there is still a large amount of uncertainty associated with the values in the database largely due to the data scarcity in the country.

Many countries, including South Africa, are committing resources for improved understanding and quantification of the interactions between surface and groundwater (Winter, 1998; DWAF, 2005a). In many of these countries, the two resources were managed separately, until the interconnectedness between them was realised. Historically in South Africa, water resource management focused on surface water research with only a limited inclusion of groundwater investigations. The more recently produced databases however (Table 2-3), have incorporated both surface water and groundwater reflecting the drive toward more integrated management.

Table 2-3 Water resource databases available in South Africa.

Year	Prepared by:	Focus	Database Name	Access
1986		SW	Management of the Water Resources of the Republic of South Africa	DWA (Department of Water Affairs)
1994	Midgley, D.C., Pitman, W.V., and Middleton, B.J.	SW	Surface Water Resources of South Africa 1990, Volumes I to VI (WR90)	WRC Report No. 298/1.1/94 to 298/6.1/94
1995	Vegter, J.R.	GW	A set of National Groundwater maps with explanation booklets.	WRC Report No. TT 74/95 / DWA
1996	Vegter, J.R.	GW	Groundwater harvest potential map of South Africa and explanation report.	DWA
1997	Schulze, R.E., Maharaj, M., Lynch, S.D., Howe, B.J. and Melvil-Thompson, B.	SW	South African atlas of agrohydrology and climatology.	WRC report TT82/96, ACRU report 46
1998	Baron, Seward and Seymour	GW	The Groundwater Harvest Potential Map of the Republic of South Africa.	
2000	Vegter, J.R.	GW	Groundwater development in South Africa and an introduction to the hydrogeology of groundwater regions.	WRC Report No. TT 134/00.
1995-2003		GW	Groundwater Resource Assessment Phase I	DWA
2001	Haupt, C.	SW	Water Resources Situation Assessment.	Groundwater Resources of South Africa
2003-2005		GW/SW	Groundwater Resource Assessment Phase II	DWA
2005	Conrad, J	GW/SW	Preparation and production of a series of GIS-based maps to identify areas where groundwater contributes to baseflow.	WRC report no. K5/1498. GEOSS Report No. G2005/02-1.
2007	Bailey, A.	GW/SW	The Water resources 2005 project (WR2005)	DWA
2012	DWA	GW	National Groundwater Archive	DWA

2.5.2. Observations and analysis based on surface water data

Low flow measures and indices from stream flow data

Hydrological methods have been developed for analysing the low flow regimes of rivers using stream flow time series. Smakhtin (2001) detailed the wide range of analysis methods available including flow duration curves and low flow frequency analysis. Indices or signatures of the hydrological response characteristics of a catchment are often used when data are scarce. Indices have been developed to represent the main drivers of effects in a catchment with a minimum of information required. Indices can be derived from output or input-output time series measured within a catchment, such as precipitation, evapotranspiration (or temperature) and stream flow (or other response variables) time series. Such response characteristics are often specific to a particular catchment and differences in response between catchments can be identified and hopefully explained. Examples include common descriptors of hydrograph shape such as runoff ratio and time to peak flow (Yadav *et al*., 2007). Indices can be based on similarity measures across a landscape, which compares parts of the landscape with other similar parts in terms of hydrological response (Blöschl *et al*., 2007). Indices have also been developed to describe overall variability in regional hydrological regimes, and to quantify flow characteristics that are believed to be sensitive to human impacts. Olden and Poff (2003) reviewed 171 currently available hydrological indices in order to aid hydrologists in the selection of appropriate ones.

Le Maitre and Colvin (2008), used river flow statistics to estimate the contribution of groundwater to river flow regimes and found that non-climatic factors are also important determinants of flow regimes. They found strong correlations with rainfall for the Baseflow Index (BFI) and the Coefficient of Variation Index (CVI) or Hydrological Index (HI) showing that the rainfall is positively related to the proportion of total flow that is baseflow and that higher rainfall results in lower flow variability. They also examined the relationships between the dominant aquifer type and selected flow statistics (Hydrological index, baseflow index, coefficient of variation index, percentage time of zero flows and flow concentration statistics) within each quaternary catchment. They found the relationships to be complex and variable and that broad scale geological and geomorphological distribution patterns play an important role in catchment water storage and its effects on the river flow regimes and the relative importance of groundwater. Although they found strong links between the flow regimes and the surrounding geology, these were often masked by large scale patterns in climate, and heterogeneous geology and geomorphology within a catchment. While they did not find a single statistic which can provide an obvious index of the groundwater contribution to a river system, they do recommend the baseflow concentration (expressed as the vector sum of the monthly baseflow, which essentially measures how evenly baseflow is distributed over 1 year) and the percentage zero flows for further testing.

In South Africa, the national water resources data set provides observed stream flow data for all gauged catchments together with naturalised flow data for every quaternary (sub-catchment) catchment in the country. The naturalised data represents the removal of all unnatural influences on the observed stream flow, and subsequent extrapolation to the remaining ungauged catchments. This 'natural' stream flow data for the entire country is useful although the process of naturalisation is highly uncertain. Low flow methods applied to the observed stream flow data must contend with often uncertain flow data, while low flow methods applied to the naturalised flow data are equally uncertain in terms of the naturalisation process.

Hydrograph recession analysis and baseflow separation

The characteristics of flow in streams during the dry season have long been recognised as different from those experienced during and following storm rainfall. Therefore a total flow hydrograph is often considered to consist of two main components; baseflow and direct runoff. Many traditional hydrograph separation techniques have focused on trying to distinguish between rapidly occurring surface runoff, slower moving interflow and even slower discharge from groundwater (Freeze and Cherry, 1979). Hughes *et al*. (2003) argues that these assumptions are only applicable in small catchments as the processes occurring in larger catchments are more complex and hydrograph shapes can be affected by a multitude of processes, some dominated by topography, others by sub-surface (soils and geology) characteristics and others by spatial variations in rainfall inputs. However, recession analysis and baseflow separation techniques have been used successfully to separate a hydrograph into its two principal components (quick and high flow from slow and low flow) in many investigations (Nathan and McMahon, 1990; Wittenberg and Sivapalan, 1999; Aksoy *et al*., 2008). These methods are, however, not capable of identifying the source of the low flows (baseflow). Groundwater hydrologists have historically defined baseflow as the groundwater portion of low flow in streams (Hall, 1968; DWAF, 2004a). Therefore, the estimates of baseflow derived from baseflow separation methods are often larger than estimates derived from observations of recharge and water table behaviour (Hughes, 2004). If reliable water resource assessments are to be undertaken, the relative contribution to baseflow from groundwater and interflow needs to be established. This is an important consideration when making decisions about licensing groundwater abstractions in the context of the effect of the abstractions on surface water. It also further emphasises the care that must be taken in inferring sub-surface processes from an interpretation of stream flow data (Hughes, 2010a).

Hydrograph recession analysis is essentially modelling baseflow recession to determine the surface runoff for single event flows (Hammond and Han, 2006). A popular method used in the literature is to consider the recession from a linear store, producing a simple exponential recession curve. However, the use of a simple linear store has been criticised and Wittenberg (1999, 2003) has suggested that recession curves are modelled better by nonlinear recession curves. However, this work has been limited to baseflow recessions where there is little or no influence from recent storm runoff (Hammond and Han, 2006). The basis for the two classes of equations can be justified physically (Wittenberg, 1999). The method which utilises a linear store was first noted in the literature by Boussinesq in 1877 and was further developed by Maillet (1905). The exponential function implies that the groundwater aquifer behaves like a single linear reservoir with storage linearly proportional to outflow. Wittenburg (1999) however, argues that in reality storage and retention effects are nonlinear which must also be assumed for the groundwater and its discharge.

While hydrograph recession analysis is applied to single event flows, baseflow separation techniques have been developed for use on continuous time series data. Techniques for separating baseflow from direct runoff can be found in the literature. Perhaps the most well-known baseflow separation techniques include that developed by Herold (1980), the UK smoothed minima method (Gustard *et al*., 1989; Wahl and Wahl, 1995; Mazvimavi *et al*., 2004) and the recursive digital filter (Nathan and McMahon, 1990; Hughes *et al*., 2003; Szilagyi, 2004). The hydrograph separation technique developed by Herold (1980) was used in the Water Resources 1990 project (Midgley *et al*., 1994) to estimate the baseflow for about 2000 quaternary catchments in South Africa, Lesotho and Swaziland. Herold (1980) suggested that the current groundwater component results from the combined effect of decay of previous groundwater discharge and previous stream flow increase. The growth and decay factors are estimated through visual calibration so that a realistic separation is achieved (Levy and Xu, 2011).

Xu *et al*. (2002) examined the results of hydrograph separation in the light of the hydrogeomorphologic setting of the reach of river being considered. They aimed to make hydrograph separation techniques more meaningful by providing a conceptual framework where time series of groundwater discharge to streams can be estimated with reasonable accuracy. They suggested different ranges of values for the growth and decay factors based on the river segment location (upper, middle or lower part of the catchment), the likely importance of bank storage, whether the river is braided or meandering and whether the water table is above or below the river stage. Although still subjective in nature, the hydrogeomorphological approach aims to be able to give a range of values of what the groundwater discharge is likely to be. Although, the author's offer criteria for identifying when interflow might be an important component of the hydrograph, this is subjective and subject to large uncertainty. Currently no baseflow separation technique is able to reliably differentiate between groundwater discharge and interflow.

The understanding gained from low flow methods applied to stream flow data can be extrapolated to ungauged basins using regional regression, regional prediction curves, regional mapping and other methods of spatial interpolation of low flow characteristics, as well as low flow estimation from synthetic stream flow time series (synthesised stream flow data based on interpolations and extrapolation of the data between gauged and ungauged catchments). Contributions of this type are numerous and varied (Nathan and McMahon, 1992; McIntyre *et al*., 2005; Wagener and Wheater, 2006; Yadav *et al*., 2007), however, due to the poor network of gauged stations and the spatial and temporal variations in low flow cycles introduced by variations in land use, these methods are highly uncertain.

2.5.3. Hydrochemistry and isotope studies

The use of hydrochemistry and isotope studies can be compared to hydrograph separation in that they can be used to separate stream flow into the different contributing water sources. However, these tracers have the added ability to identify and trace the processes occurring in a catchment in a far more specific way then hydrograph separation studies. The use of multiple geochemical tracers to characterise surface and groundwater interactions have been described by Swarenski *et al*. (2001) and Sholkovitz *et al*. (2003) and the examples of these types of investigations are numerous (Soulsby *et al*., 1999; Rodgers *et al*., 2004; Carey and Quinton, 2004). Tracer studies can be utilised in a various ways, from small (Wenningger *et al*., 2008) to large scale studies (Tetzlaff and Soulsby, 2008), and in many different hydrological environments. Oxtobee and Novakowski (2002) successfully used stream and aquifer temperatures, electrical conductivity, stable isotope analysis as well as hydrometric measurements in a catchment in Canada to identify and quantify groundwater discharge in a fractured bedrock environment. There are numerous types of geochemical and isotopic tracers including chloride (de Vries *et al*., 2002), stable isotopes (Turner and Townley, 2006), major chemical parameters such as sodium, nitrate, silica and conductivity (Oxtobee and Novakowski, 2002), tritium (Selaolo, 1998), alkalinity (Rodgers *et al*., 2004), electrical conductivity (Harvey *et al*., 1996), isotopes of radon (Cook *et al*., 2003), chlorofluorocarbons (Cook *et al*., 2003), strontium (Negral *et al*., 2003) and radium (Kraemer, 2005).

Kalbus *et al* (2006) detail the numerous types of investigations that can be carried out using isotope tracers (although investigations frequently combine hydrochemistry and isotope analysis). Stable isotopic tracers like stable oxygen and hydrogen isotopes can be used to distinguish rainfall event flow from pre-event flow (rainfall will have a different isotopic composition than water already in the catchment). Deuterium and ^{18}O have been used to quantify groundwater discharge to surface water (Space *et al*., 1991), trace transmission losses from surface water systems (Lawson *et al*., 2008) and distinguish the sources of groundwater recharge (Blasch and Bryson, 2007). Turner and Townley (2006) used both chloride concentrations and stable isotopes to identify the area of an aquifer that was being recharged by lake transmission losses in Western Australia based on the fact that isotopic enrichment was occurring in the lakes due to evaporation. The methods are best applied together as individually each method has inherent uncertainties. Isotopic enrichment by evaporation can occur through exchange processes across the air-water interface and the water can reach an upper limit of enrichment, while chloride can be enriched in the sub-surface due to transpiration from shallow water tables.

Tracer based hydrograph separation data yield groundwater discharge rates from reach to catchment scale and have even been used to monitor the pathways of inter-basin groundwater transfer in Costa Rica (Genereux *et al*., 2002). Tetzlaff and Soulsby (2008) detail the advantages of tracer investigations when investigating large scale catchments. The authors state that geochemical and isotopic tracers can reflect the integration of smaller scale hydrological processes that underline emergent properties of catchment response at larger scales. They combined hydrometric data with insights from geochemical and isotopic tracers to examine the spatial and temporal variation in baseflow characteristics in a large river system (1849 km^2) in Scotland. By monitoring similarities and differences in tracer behaviour, the authors conceptualised and estimated quantitatively the relative importance of different catchment water sources in generating river baseflow. On a smaller scale, the differences in concentrations of environmental tracers between surface and groundwater can be used to identify and delineate zones of groundwater discharge or recharge, provided that the differences are sufficiently large. Hydro-chemical tracers are often used to determine the fractions of water flowing along different sub-surface pathways (Cook and Herczeg, 2000). Historical tracers, such as bomb-pulse tritium and chlorine-36, have proved useful in delineating preferential flow in many regions (Nativ *et al*., 1995; de Vries *et al*., 2000).

2.5.4. Observations and analysis based on sub-surface data

Kelbe and Germishuyse (2010) outline three steps considered integral to a hydrogeological evaluation of the interaction between surface and groundwater. These include defining the aquifer (hydrostratigraphic units), preparing a water budget and defining the flow system components. The basic data requirements for developing the perceptual model include those used to define the physical framework:

- Geological maps and cross sections showing the areal and vertical extent and boundaries of the system.
- Topographical map showing surface water bodies and divides.
- Contour maps showing the elevation of the base of the aquifer and confining beds.
- Isopach maps showing the thickness of the aquifer and confining beds.
- Maps showing the extent and thickness of streams and lake sediments.

and those used to define the hydrogeological framework:

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

Hydrogeologists then typically use this information to determine the spatial extent of the aquifer, incorporating all the inflow and outflow zones. In surface and groundwater interaction investigations, the delineation of the intervening surface boundary is required. Similarly, the transport of water within the aquifer to the interactive surface boundary requires the delineation of the aquifer unit boundaries (using mapping techniques). The resulting information is then commonly used to parameterize groundwater models. This methodology highlights the technical approach adopted by groundwater hydrologists and the data intensity of the methodology.

The list of data requirements outlined by Kelbe and Germishuyse (2010) represents the ideal set of data necessary to formulate a perceptual model and quantify the processes occurring within that model. In most investigations, however, the available data is far less than ideal and a perceptual model must still be constructed with the data on hand. Once the perceptual model has been constructed using the available data, the main features of the aquifer are used to construct a mass balance. For the aquifer, the recharge, discharge and storage are determined. This section attempts to identify which components of this “wish list” are available and can be practically utilised to quantify the interactions between surface and groundwater, as well as identify those components which are impractical or not widely available in a data scarce country such as South Africa.

Defining the physical framework

A wealth of information can be gained from topographical and geological maps, which are available for most countries. These can be used to identify the regional geology in relation to the topography and surface water divides, and cross sections can identify aquifer layers and provide information about aquifer thicknesses and extent. This initial investigation can give an idea as to the type of interaction environments to be anticipated, and an early perceptual model can be formulated. Data such as contour maps, isopach maps and information on the extent and thickness of streams and lake sediments is not widely available. There are, however, tools and methods available for the specification of the upper and lower boundaries of an aquifer. GIS can assemble the primary data (aerial photographs, satellite imagery, etc.) usually based on elevation contours. Any data that is available (such as borehole logs, bathymetric surveys, river cross sections, any geophysical data etc.) is often point source or limited spatial data and can be extrapolated using interpolation and extrapolation techniques (Paillet, 1993;

Kelbe and Germishuyse, 2010; Munz *et al.*, 2011). In hydrogeological studies this process aims to generate information such as upper and lower aquifer boundaries and boundary position of vertical flow processes such as recharge, for use in groundwater models.

Utilising all available data sources is valuable as overlaps and links in the different types of data can thoroughly strengthen the perceptual model. While this is an important process to follow in a data scarce region, it is time consuming and not practical to carry out at larger scales on a national basis. There are not many regions in a data poor country such as South Africa that have sufficient data to produce robust information in the form of detailed maps on a meaningful scale (for water resources purposes). It is important to acknowledge the uncertainties inherent in these maps when utilising the resulting information.

Defining the hydrogeological framework

Some of the most valuable information in surface and groundwater interaction studies is piezometric surface or water table elevation data. Often this is a sparse network of information so needs to go through similar interpolation procedures detailed to produce a water level that can be incorporated into the relevant model. The most reliable water level information is obtained from monitoring boreholes (boreholes not affected by abstraction), otherwise water level measurements can be obtained from abstraction boreholes that have ceased pumping for a sufficient amount of time for the water level to recover. The DWA database in South Africa has a substantial volume of borehole data that is freely available (DWAF, 2012a). However, like many other regions, there is a huge amount of data that has not been included in the database and in many areas the coverage is extremely poor or unreliable.

Surface water hydrographs are available in many countries and while the South African network is fairly sparse, there is still a lot of accessible data. The biggest problem with surface water hydrographs is the unaccounted for and poorly documented upstream human interferences which mask the natural variability in the stream flows. Even where stream flow gauges exist, the natural flow regime is therefore effectively ungauged (McMillan *et al.*, 2010). Long term records of groundwater hydrographs are even rarer in many countries and the data available are also impacted by unknown volumes of groundwater abstractions. Where they are available, borehole hydrographs can be very useful for measuring the recharge and discharge of an aquifer. It is important that the monitoring borehole is well placed and is not near a groundwater extraction point, if the data obtained hopes to be meaningful. According to Kelbe and Germishuyse (2010), borehole hydrographs exhibit similar characteristics to a stream hydrograph with three characteristic recession rates (quick, intermediate and base recessions) although they found that the groundwater recession rates were more linear than the exponential rates observed in most of the stream hydrographs.

The hydraulic conductivity or permeability of an aquifer is the main hydraulic factor that determines the flow within an aquifer according to Darcy's Law (Freeze and Cherry, 1979). Although point measurements using permeameters or grain size analysis can be carried out to determine permeability, these are unreliable due to the heterogeneous nature of most aquifers (even primary aquifers). Probably the most reliable method of measuring the hydraulic conductivity of an aquifer is from pump tests or slug and bail tests in boreholes (although this method can be unreliable in a secondary aquifer setting where the fractures are not well connected). Details of these analyses and tests can be found in Kalbus *et al*. (2006). Certainly maps and cross sections of these aquifer characteristics are almost impossible to reliably define on a large scale in a country such as South Africa where the majority of aquifers consist of fractured rock. However, regional mapping at less detailed levels which incorporates relative degrees of fracturing and identifies areas of higher hydraulic conductivity such as dykes, sills and major fault zones would be very useful for integrated water resource management. Similarly characterising the storage properties of fractured rock aquifers is very difficult. Other types of relatively homogeneous aquifers such as alluvial or primary aquifers can be defined more accurately using pump tests and slug and bail tests, however issues such as heterogeneity and scale are still relevant (Schad and Teutsch, 1994).

Defining the hydraulic conductivity of the sediments lining streams and lakes would require a fairly detailed field investigation using techniques such as grain size analysis, in situ permeameter tests, constant head permeameter tests or a constant head injection of water through a screened piezometer (Kalbus *et al*., 2006). This information would only be realistically obtained in small scale investigations. However, there are authors who argue that information on clogging layers is essential in surface and groundwater interaction studies (DWAF, 2006a; Osman and Bruen, 2002), and sediment properties and thickness could probably be inferred (with much uncertainty) using the perceptual model.

Spatial and temporal distribution of rates of the major water balance components like evapotranspiration, recharge and groundwater baseflow need a model to characterise. The level of detail necessary would be determined by the purpose of the investigation which would dictate the type of model and therefore data required. Without a detailed field investigation, parameter values for these components are highly uncertain in heterogeneous environments such as South Africa (Ricciardi, 2009; Refsgaard *et al*., 2012).

2.5.5. Constructing a water balance

Unsaturated zone processes

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). Previous studies have found substantial volume differences in water samples taken in the unsaturated zone using suction and gravity samplers (Brandi-Dohrn *et al*., 1996) and cores (Fleming and Butters, 1995). Landon *et al*. (1999) attributed this to water moving at different rates and pathways. Stable isotopes of oxygen ($\delta^{18}O$) and hydrogen (δD) are conservative tracers of water and have been used to identify the sources and pathways of water in the unsaturated zone (Payne, 1988; Izbicki, 2000). Darling and Bath (1988) determined that matrix water extracted from cores was isotopically different from gravity drainage water in unsaturated chalk. In most of these 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 focussed 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). In sandy soils, preferential flow occurs as funnelling in layered, sandy soils (Kung, 1990) and unstable fingering flow in homogeneous sandy soils (Selker *et al*., 1992). 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). In complex fractured rock environments water movement may be dominated by fracture zones which are almost impossible to characterise (Voss, 2003; Hughes, 2010a) although the connectivity of fractures is a critical feature controlling water movement in the unsaturated zone (Berkowitz, 2002). Should interflow form a substantial portion of the baseflow in a catchment, one of the implications would be that volumes of recharge estimated using the near surface water balance will not be the same as the volume of recharge that reaches the regional groundwater table. Estimates based on hydrochemistry or isotopic signals would be more reliable (de Vries *et al*., 2000; Butler and Verhagen, 2001). In the absence of detailed field data when characterising an interaction environment, data such as groundwater levels (thickness of unsaturated zone), vegetation density and type (evapotranspiration volumes), geology, topography, soil characteristics (which can give some idea of infiltration and field capacities) and climate information can lead to a much clearer understanding of the expected dominant processes in the unsaturated zone (Sousa *et al*., 2012).

Recharge

Groundwater recharge varies significantly over a catchment both spatially and temporally, especially in arid and semi-arid zones and it is not straightforward to trace its path on a scale that might be useful for water resource management. Due to the difficulty of measuring recharge in a heterogeneous environment, and in upscaling any data that is collected, many recharge estimates vary considerably for the same region (de Vries *et al*., 2000). Nevertheless, the reliable quantification of recharge on a large scale is required for integrated water resource management (IWRM) because recharge is often the variable to which the simulated results are most sensitive (Delin *et al*., 2006). The processes behind the recharge of groundwater can be complex and DWAF (2004b) gives a detailed overview of the important recharge mechanisms and identified processes. These include infiltration capacities and field capacities of soils which define the thresholds controlling the movement of water through the soil, as well as the permeability and storage capacity of an aquifer which defines the volume of water it is able to accept. Other relevant processes to consider include rainfall (amount, type, intensity and duration), evaporation, surface slope, vegetation type, storm runoff, interception, transpiration, the presence of macropores and fracturing, and moisture retention capacity of the unsaturated zone (DWAF, 2004b). 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 (1997) also distinguishes:

- 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 there are many techniques that can be used to estimate recharge rates, all have limitations and no technique has yet been identified which does not include substantial uncertainty (Simmer, 1987; Bredenkamp *et al*., 1995). Therefore application of multiple methods is recommended in estimating recharge because of the limitations inherent in each method (Delin *et al*., 2006). DWAF (2004b) categorises methods to determine recharge rates as being either physical or chemical. Physical methods attempt to estimate recharge from water balances calculated either from hydrometeorlogic measurements, direct estimates of soil water fluxes based on soil physics or changes in the aquifer's saturated volume based on water table fluctuations. Chemical methods are based on the distribution of a tracer (commonly ^{2}H, ^{3}H, ^{14}C, ^{18}O and Cl) in the saturated or unsaturated zone (Table 2-4). 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 (Gieske, 1995; Bredenkamp *et al*., 1995; Meyer and Godfrey, 1995), isotope applications (Verhagen, 1979), water balance method (Fleisher, 1981; Kirchner *et al*., 1991) and conceptual modelling and rainfall analysis (Rawlins and Kelbe, 1998). Delin *et al*. (2006) compared results of a regional approach to multiple local scale values to determine the applicability of regional scale estimates at the local scale. They found that the regional estimates (based on RORA, a basin scale analysis of stream flow records using a recession curve displacement technique) compared favourably to most of the local scale estimates (based on water table fluctuations and age dating of groundwater), while results from an unsaturated zone water balance were inconsistent with the results from the other methods (both regional and local scale).

Table 2-4 Overview of common recharge estimation methods (adapted from Xu and Beekman (eds.) (2003).

Method	**Principle**	**References**
Surface water zone		
Physical methods		
Hydrograph separation	Stream hydrograph separation: outflow, evapotranspiration and abstraction balances recharge	Burns (2002)
Water budget (catchment or channel scale)	Recharge derived from difference in flow upstream and downstream accounting for evapotranspiration, in and outflow and channel storage change	Lerner *et al*. (1990);
Watershed modelling	Numerical rainfall-runoff modelling, recharge estimated as a residual term	Sami and Hughes (1996)
Rainfall-recharge relationship		Vivier (2009)
Unsaturated zone		
Physical methods		
Lysimeter	Drainage proportional to moisture flux/recharge	Bredenkamp *et al*. (1995)
Seepage meter	Point measurements of the velocity and direction of flow	Rosenberry and Morin, 2004; Swarzeski *et al*., 2005
Unsaturated flow modelling	Unsaturated flow simulations e.g. by using numerical solutions to Richards equation	Sami and Hughes (1996)
Zero flux plane	Soil moisture storage changes below zero flux plane (zero vertical hydraulic gradient) proportional to moisture flux/recharge	Selaolo (1998)
Soil moisture balance	Unsaturated flow balance, recharge estimated as a residual term	
Tracer methods		

Chloride mass balance	Profiling: drainage inversely proportional to Cl in pore water	Beekman *et al*. (1997)
Historical	Vertical distribution of tracer as a result of activities in the past (^{3}H)	Gieske (1995)
Natural tracers	Water balance based on isotopes ^{2}H, ^{18}O and ^{4}He	
Saturated-unsaturated zone		
Physical methods		
Cumulative rainfall departure	Water level response from recharge proportional to cumulative rainfall departure	Xu and van Tonder (2002)
EARTH	Lumped distributed model simulating water level fluctuations by coupling climatic, soil moisture and groundwater level data	Van der Lee and Gehrels (1997)
Water table fluctuation	Water level response proportional to recharge/discharge	Bredenkamp *et al*. (1995)
Tracer methods		
Chloride mass balance	Amount of Cl into the system balanced by amount of Cl out of the system for negligible surface runoff/run on	Selaolo (1998)
Saturated zone		
Physical methods		
Groundwater modelling	Recharge inversely derived from numerical modelling groundwater flow and calibrating on hydraulic heads/groundwater ages	Gieske (1995)
Saturated volume fluctuation	Water balance over time based on averaged groundwater levels from monitoring boreholes	Bredenkamp *et al*. (1995)
Equal volume – spring flow	Water balance at catchment scales	Bredenkamp *et al*. (1995)
Tracer methods		
Groundwater dating	Age gradient derived from tracers, inversely proportional to recharge. Recharge unconfined aquifer based on vertical age gradient (^{3}H, CFCs, $^{3}H/^{3}He$); Recharge confined aquifer based on horizontal age gradient (^{14}C)	Weaver and Talma (1999)

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 (e.g. Lloyd, 1986; Sami, 1992). In these areas extreme local variability in recharge results from focused recharge beneath ephemeral streams and preferential flow mostly in fractured systems (Sharma and Hughes, 1985; Kirchner *et al*., 1991; 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 (Lloyd, 1986). 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. DWAF (2004b) concluded that the chloride mass balance, cumulative rainfall departure, the EARTH model, groundwater modelling, saturated volume fluctuation and the water table fluctuation methods to be most applicable in arid and semi-arid regions. Of these, the chloride mass balance is the easiest, least expensive and most widely applied (Scanlon *et al*., 2006) while groundwater modelling is the most difficult and expensive.

The most common methods used to determine recharge in South Africa involve mass balance and/or the use of numerical models. Within South Africa's predominantly fractured rock aquifers, preferential recharge is a significant process. Due to the difficulty of measuring recharge in such a heterogeneous environment, most recharge figures are estimated by using a model to link recharge to rainfall (Kelbe and Germishuyse, 2010). While authors (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. Despite many contributions to recharge processes and their quantification, there is still not enough information available about the processes and their temporal variability at the catchment scale (especially in arid and semi-arid regions with large unsaturated zones). Without reliable estimates of recharge patterns, the other groundwater balance components of any model will remain highly uncertain.

Evapotranspiration

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. Reliable measurements of hydrological variables, including evapotranspiration, are therefore useful for validating model processes, and quantifying individual components of the water balance.

Vegetation affects aquifers by directly extracting groundwater from saturated strata and reducing the proportion of rainfall that is eventually recharged by interfering with the passage of precipitation from the atmosphere to the water table in recharge areas. The size of the total vapour losses associated with vegetation cover is dominated, in many countries including South Africa by the transpiration component and to a lesser degree by interception (Scott and Lesch, 1997; Le Maitre *et al*., 1999). 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% (Larcher, 1983; Le Maitre *et al*., 1999). Evapotranspiration utilises water from all available water sources depending on factors such as rooting depth and density, the availability of water and the physiology of the plants. The extraction of groundwater most often occurs either through deep routed vegetation or through vegetation situated in riparian areas (along channel margins). Vegetation of different types can transpire water from the soil profile, regolith, saprolite and rock fractures (Stone and Comerford, 1994). Numerous studies (Issar *et al*., 1984; Lima *et al*., 1990; Gee *et al*., 1994) 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) (Lima *et al*., 1990; Stone and Kalisz, 1991). In arid and semi-arid environments, many plants have shallow, spreading root systems. While investigations into the vegetation-groundwater interactions often focus on alluvial aquifers (Stanford and Ward, 1993) or primary aquifers (Rawlins and Kelbe, 1991, Scott *et al*., 1993), research into the relationships in fractured rock environments is lacking (Scott and Lesch, 1997). Many investigations only examine processes in the shallow soil zone (Dodd *et al*., 1984) but the roots of many vegetation types extend much deeper (Thorburn and Walker, 1994). Many reviews show that root depth is generally only limited by water tables or by the soil or regolith characteristics that prevent rooting (Canadell *et al*., 1996; Jackson *et al*., 1996). 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).

Potential evaporation data (available nationally for most countries) together with some knowledge of vegetation types and riparian strip widths (Bond *et al*., 2002) can be used to estimate evapotranspiration rates. There are numerous methods for estimating evapotranspiration (Serrat-Capdevila *et al*., 2011), with the most common including the Penman and Penman-Monteith models (which use values for atmospheric radiation, temperature, humidity and wind speed). Instrumentation for the estimation of evaporation includes the eddy correlation or Bowen ratio (Kelbe and Germishuyse, 2010).

Emergence

Recharging 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:

$$\boldsymbol{R = ET - BF_{GW} - DS_{GW} - ABS + CL + US_{GW} + \Delta S} \qquad \textbf{Equation 1}$$

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

Attempting to quantify the remaining groundwater outflow processes is essentially working out what happens to the recharge that has moved into a system. Sub-surface processes such as downstream and regional groundwater flows are not possible to directly measure. Similarly groundwater baseflow cannot be directly measured only surmised using indirect methods (likewise with channel transmission losses). Some idea of the relative importance of many of these outflow components can be gained by examining information on the topography, regional groundwater gradient and hydraulic conductivity of the environment. Lastly, the remaining significant output process includes groundwater use and depending on the quality of records in a region can be poorly or well defined. 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 etc. 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 would be able to approximately estimate storage in these heterogeneous environments.

2.6. Modelling

2.6.1. Introduction

Models generally aim to represent the complex and spatially distributed interaction of water and energy by means of mathematical equations. Computer based models attempt to represent the conceptual understanding of these interactions 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

The models described below are categorised into simple, complex and compromise models, however, it is not always straightforward to group certain types of models together as there are often overlaps within different model structures. The following definitions provide a simple reference for the sections below:

SIMPLE MODELS: are water balance models. While they do sometimes retain some physical representation, most do not include a distributed representation (i.e. gridded spatial representation) of a catchment. Their complexity varies between models consisting of a single uncomplicated equation to fairly elaborate models which consist of a series of intricate equations. Also termed NON-PHYSICALLY BASED MODELS, CONCEPTUAL MODELS, NON-DISTRIBUTED MODELS, and LUMPED MODELS.

COMPLEX MODELS: are usually numerical groundwater models. These are based upon mathematical equations that describe fundamental physical processes. They are mostly distributed models which operate over a large number of elements or grid squares. Also termed PHYSICALLY BASED MODELS, NUMERICAL MODELS, and DISTRIBUTED MODELS.

INTEGRATED MODELS: are models which incorporate elements of both surface water and groundwater systems and are often models which are specifically designed for either surface water or groundwater modelling and have incorporated additional components in order to become more integrated. These models can be complex (physically based) or simple (non-physically based).

COMPROMISE MODELS: are positioned between simple and complex models. Compromise models are often advantageous in that they are neither too simple (cannot simulate the processes in sufficient detail) nor to complex (unrealistic data requirements and a high level of irresolvable equifinality).

GROUNDWATER MODELS: are focused primarily on the representation of sub-surface processes. They are usually distributed numerical models although there are examples of simple, lumped groundwater models.

SURFACE WATER MODELS: are focused primarily on the representation of surface water systems. They are usually lumped, conceptual type models although there are examples of complex, distributed surface water models.

SEMI-DISTRIBUTED MODELS: are often lumped models which have been developed to be applied at smaller spatial scales thereby offering a more detailed representation of a catchment without applying a fully gridded spatial representation.

2.6.2. Simple modelling approaches

Simple or non-physically based modelling approaches are based upon the principle of the conservation of mass and endeavour to ensure that chances of incorrect accounting and subsequent incorrect distribution of the water resource are minimised (Wright and Xu, 2000). Van Tonder and Kirchner (1990) concluded that the groundwater balance method is the only method which yielded reliable estimates of groundwater recharge. The complexity of non-physically based modelling approaches varies between models consisting of a single uncomplicated equation to fairly elaborate models which consist of a series of intricate equations. An example of a simple model is that formulated by Wright and Xu (2000). Their approach follows a simple equation, i.e. the recharge equals the discharge at a steady state in an unexploited hydrological unit. If the unit is exploited, the water balance is as follows:

Adjusted recharge – reduced discharge – pumping + storage = 0   Equation 2

More complicated non-physically based or water balance models include conceptual rainfall-runoff models which are frequently used in hydrological modelling. The model equations are based upon a simple interpretation of the physical processes acting upon the inputs and outputs of a system. Often the catchment is conceptualised as an assemblage of interconnected storages, with individual storages representing key aspects of, or processes within, the system. The controlling equations satisfy the water balance (Xu and Singh, 1998).

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*., 2012). 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; Jakeman and Hornberger, 1993; Young *et al*., 1996; Le Moine *et al*., 2008) 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 uncertainty in 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". 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, 2004a). Further criticisms were outlined by DWAF (2006a) 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?

2.6.3. Complex modelling approaches

The most complex or physically based models are usually numerical groundwater models. These are based upon mathematical equations that describe fundamental physical processes. They are mostly 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, 2011). Numerical models of surface-groundwater interaction combine numerical solutions of equations for surface water routing and groundwater flow using mostly 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, 2006a).

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. The choice between using a FD or FE solution is largely personal and proponents of a particular method can easily support the strengths of their preferred solution strategy. According to Simpson and Clement (2003), the main difference between them is in how the numerical schemes spatially average the variation of material properties. In their comparison, the algorithms performed equivalently for a one dimensional problem but for a two dimensional problem the FD solution was plagued by numerical errors. They argued that the FE approach avoided many of the same problems due to the intrinsic averaging of material properties and improved representation of specified flux boundaries.

Complex physically based 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 (to the limit possible 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 the ability to represent this variation in the dominant hydrological processes. However, if we don’t have the catchment knowledge, which is frequently the case in data scarce countries, 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).

2.6.4. Compromise models

Compromise models are positioned between simple (non-physically based) and complex (physically based) models and in the context of surface and groundwater interaction modelling are often models which are specifically designed for either surface water or groundwater modelling and have incorporated additional components in order to become more integrated. There have been approaches by both surface and groundwater hydrologists to support integration and many of the models detailed below are examples of newly integrated models. Compromise models are often advantageous in that they are neither too simple (cannot simulate the processes in sufficient detail) nor to complex (unrealistic data requirements and a high level of irresolvable equifinality), although integrated models can vary in complexity between highly complex (physically based) or less complex (non-physically based). There are many types of conceptual and numerical models and the examples selected for review are a selection of some of the commonly used kinds. The reviews that follow are organised according to Figure 2-4 and Table 2-5, and start with the simplest models working up to the most complex. Many of these models are reviewed in more detail in DWAF (2004a) and DWAF (2006a).

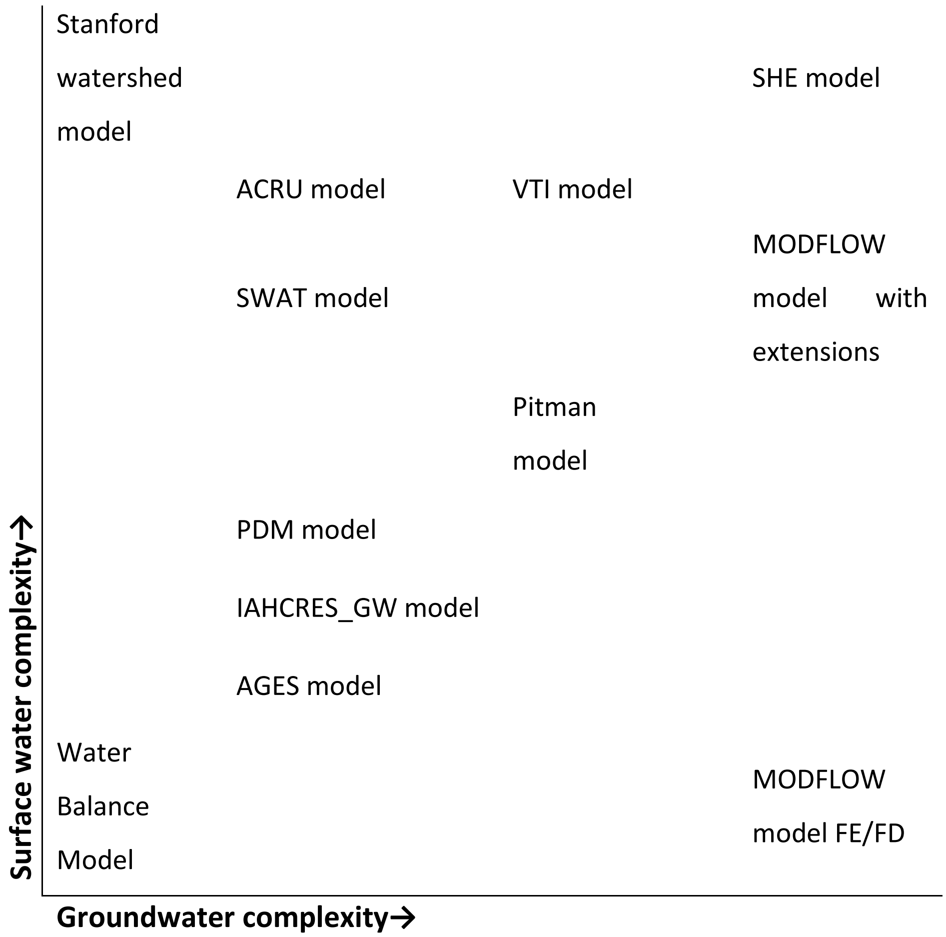


Figure 2-4 Examples of surface and groundwater models arranged according to complexity in their surface and groundwater components.

The **IHACRES_GW model** (Identification of unit Hydrographs and Component flows from Rainfall, Evaporation and Stream flow data) (Jakeman and Hornberger, 1993) is a robust simple model used worldwide (Littlewood, 2002; Dye and Croke, 2003) which has had a groundwater model component module incorporated (Ivkovic *et al*., 2005). The IHACRES_GW model takes effective rainfall as input, and stream flow is generated as model output. It is a parsimonious model with just four parameters calibrated through visual inspection of stream flow data and flow duration curves to achieve the best model fit. The groundwater storage is conceptualised as a single reservoir with outflows including groundwater baseflow, and groundwater extractions and other losses. The rainfall is portioned into surface runoff and groundwater baseflow. The model assumes that the low flow signal represents groundwater baseflow and ignores other potential sources of baseflow. The structure of the model and the mass balance equation used for groundwater storage is given in Figure 2-5.

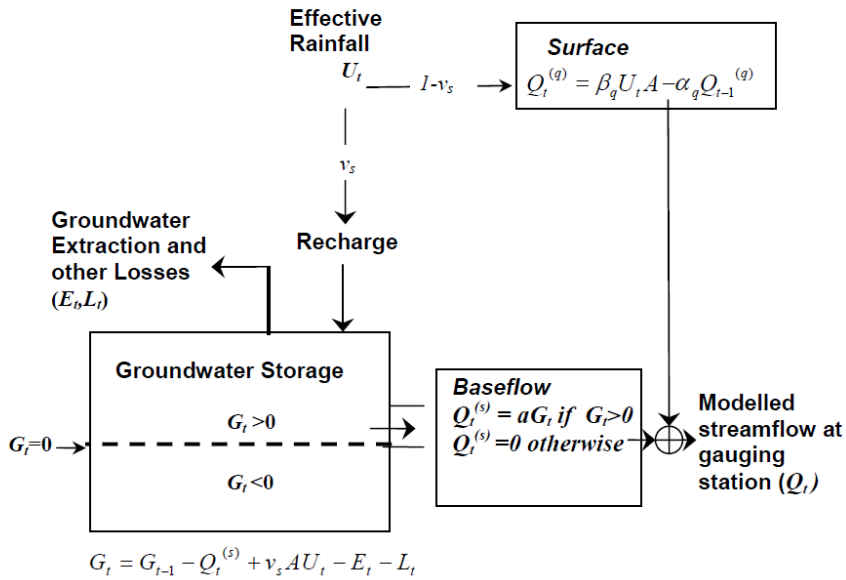


Figure 2-5 IAHCRES_GW model structure (Ivkovic, 2006).

where E_t is the groundwater extraction at time step t, which is obtained from available data. L_t represents any losses from groundwater storage at time step t, including sub-surface outflow below the level of the stream gauging station, evapotranspiration and other losses (or gains if the loss term is negative such as would be the case with irrigation returns and river infiltration resulting in groundwater inflow). L_t is calibrated using stream flow data through fitting the parameter to the decay observed in the stream hydrograph (Ivkovic, 2006). The advantage of the IAHCRES_GW model is its simplicity and few parameters, however, the disadvantages of the model include it not being applicable in ungauged basins and the inability to simulate unsaturated zone processes.

A regional groundwater flow balance model developed in South Africa is the **AGES (Vivier) model**. The model was developed in 2005 for the purposes of groundwater management and subsequently has undergone updates (Vivier *et al.*, 2007; Vivier, 2009). The conceptual model is based on that derived by Wright and Xu (2000) and determines the groundwater flow volumes per quaternary catchment at a given management constraint. It simulates variable recharge from statistical rainfall distributions. Baseflow calculations are based on Darcy's Law with a gradient determined by topography. The model accounts for recharge from dam seepage, groundwater supported wetlands, abstraction, mine dewatering, irrigation with return flows, riparian vegetation, alien vegetation, the groundwater component of baseflow and the environmental water requirement (minimum volume of stream flow required to environmental requirements). The model can also be constrained using baseflow volumes from stream flow data, which should be more and not less than the groundwater component of baseflow. The aim of the methodology is not to determine the actual groundwater balance as the frequent data scarcity means it is often an impossible task, but rather aims to determine groundwater flow balances that can be allocated for groundwater management purposes.

The **Probability Distributed Model (PDM)** is a fairly general conceptual rainfall-runoff model developed in the UK for flow forecasting (although it has been used in the arid environment of Australia). The PDM has been designed more as a toolkit of model components than as a fixed model construct which means extending a component to include additional processes is relatively straightforward. Several options are available in the overall model formulation which means a broad range of hydrological behaviours can be represented. Moore and Bell (2002) adapted solutions to the Horton-Izzard equation resulting from the conceptual model of groundwater storage and developed a generic model component for representing groundwater storage under the influence of pumped abstractions, spring flows and underflows (downstream groundwater flows). The aquifer storage is represented in a nonlinear fashion and this can be extended to represent ephemeral flows (which occur when recharge does not offset groundwater losses). Runoff is generated from saturated probability distributed stores (representing the fast pathways to the catchment outlet). The groundwater storage, representing routing of water to the catchment outlet via slow pathways, is usually taken to be of cubic form, with outflow proportional to the cube of the amount of water in storage. This storage can be depleted by pumped abstractions from groundwater, downstream groundwater losses and groundwater fed springs. The outflow from both surface and groundwater storages together with any fixed flow from other sources (compensation releases from reservoirs or constant abstractions) forms the model output. While a soil moisture store is included in the model, the additional components have been developed for a groundwater dominated system so interflow and perched aquifers are not explicitly simulated.

The **Pitman Model** is a conceptual rainfall-runoff monthly time scale model. Originally the model consisted of eleven catchment parameters to simulate runoff (Pitman, 1973), but has since been modified in two separate ways (Hughes, 2004; Sami, 2006) to incorporate more explicit groundwater routines. Both new versions are based upon similar concepts and simulate all the major groundwater processes (Figure 2-6). The main differences between the two versions of the model are that the Sami version simulates actual groundwater levels while the Hughes version simulates a regional groundwater gradient (a groundwater level on such a large scale is deemed to variable and uncertain). Both versions calculate baseflow and transmission losses with a non-linear model as a function of the gradients between the surface and groundwater levels (Sami, 2006) or slope (Hughes, 2004) in the catchment. In addition the Sami (2006) version can allegedly simulate the spatial effects of a single borehole extraction point while the Hughes (2004) version maintains that the simplicity and scale of the model cannot represent individual borehole effects (groundwater abstraction has to be large scale to have any effect on the groundwater store of an entire catchment). Both versions are semi-distributed with groundwater components based on relatively simple geometry. In the Hughes (2004) version, recharge calculations are based on a maximum monthly recharge parameter, the status of the unsaturated zone storage and a non-linear power function. The recharge estimate is added to the groundwater store while losses from the groundwater store include riparian evapotranspiration at the channel margin, groundwater discharge to the channel and groundwater flow to downstream catchments. Groundwater abstractions are also allowed for in a simplified way. Water balance calculations are then used within each time interval of the model to update the groundwater slope. Both models simulate processes such as transmission losses and interflow through the unsaturated zone. The models have been criticised by groundwater hydrologists for being too simple for South Africa's heterogeneous conditions.

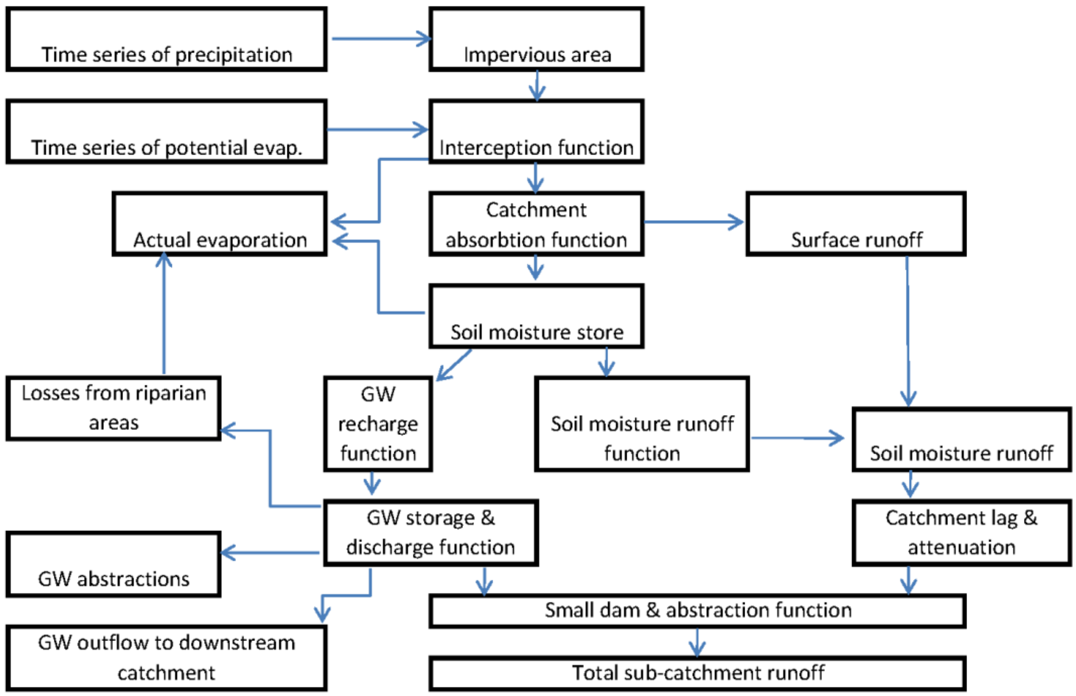


Figure 2-6 **Conceptual structure of the Hughes (2004) version of the Pitman Model, although the structure could also represent the Sami (2006) version (adapted from Hughes, 2004).**

The **SWAT (Soil and Water Assessment Tool)** (Arnold *et al.*, 1998) model is a quasi-distributed physical model with sub-basins modelled as a lumped unit. It is a continuous time scale model using a daily time step. It was developed to predict the impact of land management practices on water, sediment and agricultural chemical yields and therefore simulates numerous processes such as canopy storage, infiltration, redistribution, evapotranspiration, lateral sub-surface runoff, pond and tributary channel functions. SWAT includes two soil layers in which water percolates downward as the upper layer reaches field capacity. Interflow is simulated as lateral sub-surface flow within the soil profile and is calculated at the same time as soil moisture redistribution is calculated (a kinematic storage model predicts lateral flow in both soil layers). Hydraulic conductivity, slope and soil water content variations are all accounted for, with upward flow also simulated.

River channels are categorised into the main channel and tributary channels (which do not receive groundwater baseflow). Water which percolates through the soil zones is partitioned into two recharge parts, the first portion recharges a shallow unconfined aquifer and the second, a deep confined aquifer (for the purpose of simulating regional flows). Outputs from the shallow aquifer include groundwater baseflow, evapotranspiration from tree roots, percolation into the deep regional aquifer and groundwater abstractions. A recession constant derived from daily stream flow records is used to lag groundwater flow from the aquifer to the stream. The groundwater component is modelled as a lumped system, therefore cannot handle variable pumping. Sophocleous *et al*. (1999) and Sophocleous and Perkins (2000) replaced the groundwater component of SWAT with MODFLOW and constructed a comprehensive basin model called SWATMOD which uses linking utilities to transfer data between the models and can be run from either model.

The model comprehensively covers the hydrological processes of a system and is one of the few models that includes regional groundwater flows. However, the model does not directly simulate surface and groundwater interactions (uses a recession constant to simulate baseflow) and does not simulate transmission losses. Since it is a lumped model it suffers from many of the same criticisms that face conceptual models (cannot position boreholes or simulate groundwater levels).

The **ACRU model** (Agricultural Catchment Research Unit) (Smithers and Schulze, 2004) is a lumped agro-hydrological model which has been tested under South African conditions. It is a multi-layer soil water budget model designed for small sub-catchments (>30km^2). The equations and objective functions used in ACRU have been explained and discussed in detail in Schulze (1995) and Smithers and Schulze (2004). The developers refer to the ACRU model as a multi-purpose, multi-level, integrated physical-conceptual model that can simulate total evaporation, soil water and reservoir storages, land cover and abstraction impacts on water resources and stream flow at a daily time step. ACRU simulates groundwater flow based on the assumption of a homogeneous and unconfined aquifer and requires data such as depth of water table, capillary fringe height and saturated hydraulic conductivity. The response coefficients applied to simulate interflow, groundwater recharge and groundwater baseflow require calibration as they do not have a physical meaning. The model simulates the soil component of the water cycle in detail and requires soil parameters such as soil porosity, field capacity, permanent wilting point and thicknesses of both the top soil and sub-soil horizons. An overview of the concepts behind the model are given in Figure 2-7.

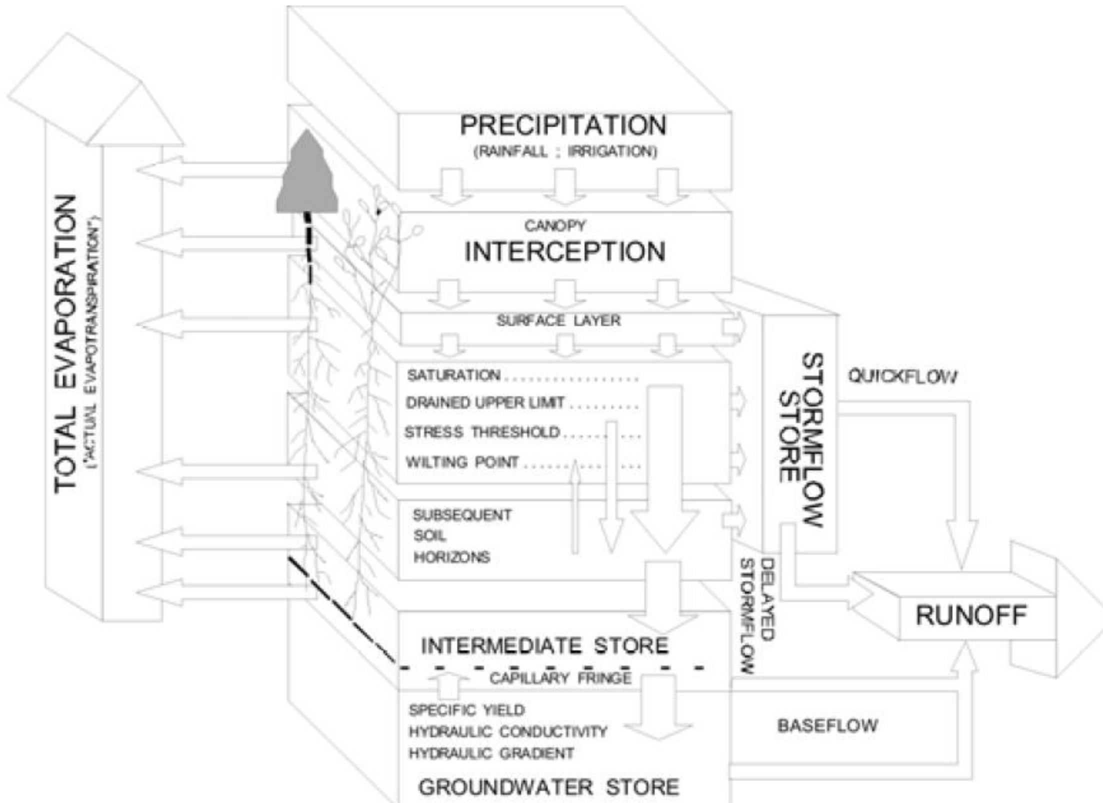


Figure 2-7 Representation of the water budget in the ACRU model (Smithers and Schulze, 2004).

For the purposes of surface and groundwater interactions, the ACRU model does not simulate groundwater processes explicitly enough and does not simulate interflow in the unsaturated zone. DWAF (2004a) criticised the ACRU model for being unable to simulate surface and groundwater interactions as baseflow is calculated by recession constants and cannot simulate transmission losses. The model is relatively data intensive,

The **VTI model** (Hughes and Sami, 1994) is a semi-distributed rainfall-runoff model, which simulates most of the interaction processes which occur in South Africa (interflow, baseflow from perched aquifers or unsaturated zone springs, groundwater baseflow, transmission losses and regional groundwater flow). The soil profile is divided into two layers with both layers contributing to interflow at saturation (based on a normal soil moisture distribution). Recharge water moves from the lower soil layer into the groundwater store and is corrected by lateral and vertical distribution factors to account for the spatial variation in recharge zones. Percolating recharge water is assumed to be uniformly distributed and is represented in the model as a storage zone termed PSTORE. Drainage from PSTORE can be vertical (to the regional groundwater) or lateral (spring flow into the stream channel). Percolation from PSTORE to the regional groundwater increments the water table which is represented by a vector with a slope equal to the hydraulic gradient. If this vector intercepts the ground surface a saturated seepage face is created longitudinally along the channel. Groundwater baseflow is calculated as the product of this seepage face and the hydraulic gradient. Groundwater transfers to other sub-catchments are calculated as the product of the adjusted transmissivity (incremented to account for increasing flow rates as the groundwater depth rises above the rest water level) and the hydraulic gradient (also incremented as the water table rises).

The algorithms are physically based and although it provides a comprehensive water balance, is relatively data and skills intensive. It does not model groundwater levels directly and cannot account for single borehole impacts on a catchment. DWAF (2004a) concluded that in South Africa, only the VTI model conceptualised most of the known surface-sub-surface interactions although its skills requirements and parameterisation issues limit its use.

MODFLOW (McDonald and Harbaugh, 1988, Harbaugh, 2005) is a finite difference numerical model that simulates groundwater flow in confined and unconfined aquifers in three dimensions. As a modular model, additional packages can be easily incorporated to enhance its capability. As a standalone model it does not simulate unsaturated zone flow or transmission losses but additional packages have been developed to incorporate these processes. Groundwater baseflow and transmission losses can be simulated using the RIVER (Harbaugh and McDonald, 1996) package. This package contains routines that calculate flow between rivers and underlying aquifers using Darcy's law based on the head difference and sediment conductance of the river bed. Water in the channel is not directly simulated and the river is treated as an infinite source of water of constant head, which means the model can generate more transmission losses than there is water in the channel. The model can therefore not simulate variable transmission losses and groundwater baseflows depending on the volume of flow in the river. For this purpose the STREAM-ROUTING (SFR1 and SFR2) (Niswonger and Prudic, 2006) package can be used instead, which is designed to account for the amount of flow in streams and then simulate the interactions. Streams are divided into reaches (individual cells in the finite difference grid) and segments (group of reaches connected in a downstream order). The package allows for inputs and outputs from runoff, precipitation and evapotranspiration, and can simulate transmission losses although the estimation of the parameters for describing relative permeability is challenging. Springs are simulated using the DRAIN or SEEPAGE (Batelaan and De Smedt, 2005) package, which can simulate both gravity and artesian springs. However, this package cannot simulate springs which are not connected to the regional groundwater body. Other relevant packages include the WETLAND (Restrepo *et al*., 1998) package, RESERVOIR (Fenske *et al*., 1996) package (for lakes and reservoirs), DAFLOW (Jobson and Harbaugh, 1999) package (for flow routing in upland gaining streams) and MODBRANCH (Swain and Wexler, 1996) package (for unsteady flow in a network of single open channel reaches).

While MODFLOW is widely used and simulates realistic heterogeneous conditions, it is data intensive and cannot provide a water balance for the entire water cycle (surface runoff is not simulated). In addition the model cannot generate a time series of groundwater baseflow unless a time series of recharge is available. The model has also been criticised for incorrect assumptions regarding some of the process characterisations (Brunner *et al*., 2010).

The **Stanford Watershed IV Model** (Crawford and Linsley, 1966) was one of the first operational, lumped or 'conceptual' models to be developed. The model is a complex general catchment model designed to represent the whole hydrological cycle in an integrated way. It was developed as a physically based model which simulates a wide variety of processes such as overland flow, infiltration, soil moisture, evaporation, transpiration, interflow and groundwater baseflow on a short (hourly) time step. The soil moisture storage is divided into two zones, the upper zone is relatively shallow and water can be removed by gravity drainage or evaporation, water is removed from the lower zone by gravity drainage and transpiration. The model was one of the first hydrological models to use "nominal" soil moisture storages and the continuous variability of assignments of water to moisture storages was unique. The use of cumulative frequency distributions for infiltration rates at a point in time to model areal infiltration and evaporation were also unique at the time of development. The successor to the Stanford Model is the HSPF (Hydrologic Simulation Program Fortran) (Wittemore and Beebe, 2000) model, and is essentially the PC based adaption of the Stanford Watershed Model.

HSPF is a lumped parameter, continuous simulation non-point source model that simulates groundwater storage as a simple reservoir. The model is a complex model and while it provides powerful simulation potential when used by knowledgeable personal, Wittemore and Beebe (2000) claim that achieving this state of expertise is a formidable task. HSPF has undergone a series of updates and code changes since its adaptation from the early Stanford Watershed Model and predecessor (EPA, 1996). The HSPF is not a distributed or physically based model and therefore unable to handle land use management effects at the field scale. It does not explicitly model groundwater levels and is unable to simulate transmission losses.

The **SHE and MIKE-SHE models.** SHE (Abbot *et al.*, 1986) is a physically based distributed model developed for water resource management and development (Figure 2-8). The SHE model simulates a wide variety of processes although only at small scales. The hydrological processes are modelled by finite difference solutions of the equations of mass, energy and momentum conservation or by empirical equations. Flow in the unsaturated zone is determined by soil moisture content and tension distributions. The SHE model assumes vertical unsaturated flow and therefore does not simulate interflow. A mass balance equation determines the exchange with the groundwater. Variations in the groundwater level of each grid are determined by the non-linear Boussinesq equation which combines Darcy's law and the mass conservation of two dimensional laminar flow in an isotropic heterogeneous aquifer (DWAF, 2004a). The model can simulate surface and groundwater interactions in gaining, losing and disconnected stream systems (a hydraulic conductivity is assigned to the stream bed and sides. One of the main advantages of the SHE model is its complete integration of both surface and groundwater processes with both water cycles represented with largely equal detail. However, it cannot separate groundwater baseflow from interflow, perched aquifers or spring flow and is highly data intensive.

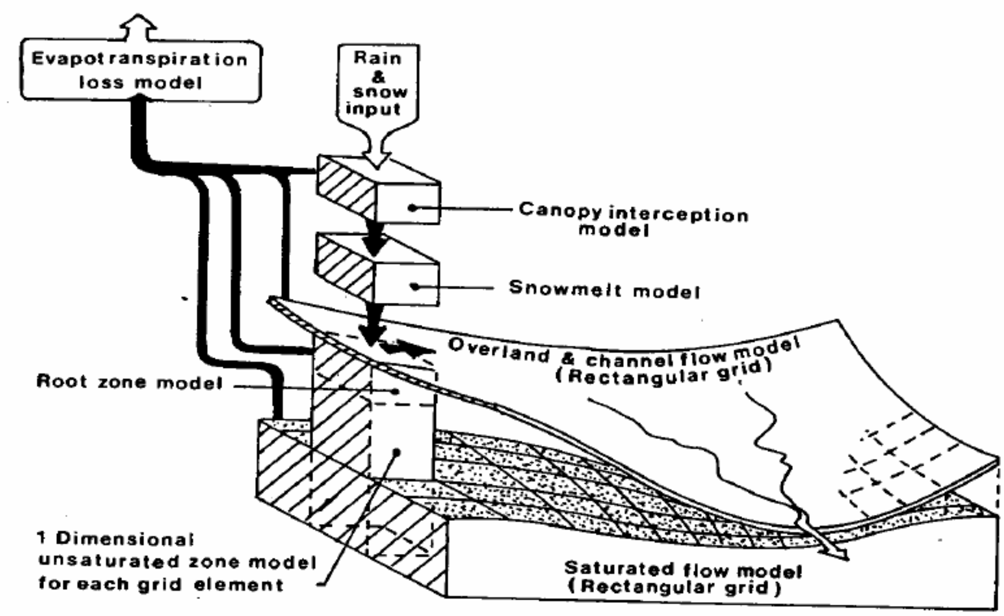


Figure 2-8 SHE model structure (Abbot *et al.*, 1986).

MIKE-SHE (DHI, 2004) is an extended version of the SHE model developed to be applicable at larger scales. The MIKE-SHE model can simulate a water budget for the full hydrological cycle, including baseflow and transmission losses and simulates groundwater levels directly. The model represents the catchment horizontally and vertically using a finite difference grid which facilitates the distribution of catchment parameters, rainfall input and the hydrological response. The extended version however, still does not model interflow and is still exceptionally data intensive.

2.6.5. Uncertainty

Historically, the reliability of model results have been evaluated using observed data (where available). More recent approaches to the application of hydrological and water resources estimation models have focused on the explicit quantification of the uncertainties in the outputs (Hughes and Mantel, 2010b). While a comprehensive review of uncertainty estimation methods will not be undertaken, the main concepts are outlined as the identification of the sources of uncertainty is essential for any integrated assessment. The science of hydrology is currently experiencing a shift from methods that focus on the identification of a single best model toward methods that attempt to reduce the uncertainty in the predictions of all possible models using ensemble methods. Instead of focusing on "optimisation", the focus is changing towards one of model "consistency" (Kapangaziwiri and Hughes, 2009). There are many hydrology and hydraulic researchers who now see uncertainty analysis as an important part of good scientific practice and a large range of papers are available on uncertainty estimation methods and applications (Beven, 2000; Vrugt *et al*., 2003; Pappenberger and Beven 2006; Renard *et al*., 2010). Beven (2001) states that it is impossible to prove a constructed water balance without allowing significant uncertainty in every input and output term. Similarly, Kapangaziwiri and Hughes (2009) point out that the notion of an optimal parameter set is considered both unwise and incorrect, especially when considered against uncertain model forcing data, model structures and incomplete and limited process understanding.

There will always be higher levels of risk associated with the management of water resources in South Africa due to the high degree of both spatial and temporal variability in available resources. In hydrological models the main sources of the uncertainty often include the reliability of the input hydro-meteorological data (typically rainfall and evaporation demand in rainfall-runoff models), unknown water use and return flows, the parameter values assigned to the algorithms used to represent the processes in a basin and model structural issues which includes spatial and temporal scale concerns as well as adequate process representation. The major sources of uncertainty are therefore common to any modelling approach. While structural uncertainties are often difficult to separate from parameter and data input uncertainties, they can be very important if one of the objectives is getting the right answer for the right reason (Kirchner, 2006). This is particularly relevant to simulating surface and groundwater interactions in hydrological models. It is important, however, not to confuse structural uncertainty and structural simplicity (or complexity) as they are not necessarily linked. Arguably, there are two levels of structural uncertainty that should be considered. The first is whether or not certain processes, known to exist in the real world, are represented in a model. The second is whether the algorithms used in the model can adequately represent the non-linearity's or thresholds that occur within the relationships between storages and fluxes, or one flux and another. Many of these uncertainties are difficult to resolve given the typically available data that can be used to define and quantify surface and groundwater interactions. Beven (2012) calls for improved methodological developments to test models and argues that the true level of uncertainty in model predictions is not widely appreciated. He identifies two common ways of assessing the 'correctness' of a model structure and setup. The first is to rely on expert opinion, but bias is inevitably introduced due to the difficulty of finding scientists not committed to one modelling paradigm or another. The second is to test the model against available data for a range of different circumstances. However, in fitting historical data is subject to the difficulties of differentiating between the various sources of uncertainty (structure, setup, input data). Beven suggests a "limit of acceptability" approach to model evaluation as a way of testing models, which would involve thought to define critical experiments that will allow models and their setups to be adequately differentiated. It should be possible to at least partly identify the uncertainty by a careful examination of the evidence for specific processes compared with the conceptual structure of a specific model. One of the advantages of uncertainty approaches to modelling is that the outputs of diverse model designs (based on different structures and/or different parameter sets) can be examined against the available evidence to identify those results that are generally behavioural (Beven, 2012).

2.6.6. 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, 2011). 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 conflict and 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 and it is beyond the scope of this thesis to review them all. The models reviewed in this thesis represent an overview of the types of models available. The reviewed models are outlined in Table 2-5 below.

Table 2-5 Overview of the processes simulated by the reviewed models.

Model details	IHACRES_ GW	AGES	PDM	Pitman w. mod.	SWAT	ACRU	VTI	MOD FLOW w. ext.	Stanford	SHE & MIKE SHE
Scale	Meso-macro	Meso-macro	Meso-macro	Meso-macro	Meso-macro	Micro-macro	Micro-macro	Micro-meso	Micro-macro	Micro-macro
Tested under South African conditions	No	Yes	No	Yes	No	Yes	Yes	Yes	No	No
Data intensive	No	No	No	Fair	Fair	Fair	Fair	Yes	Yes	Yes
Can determine positions of boreholes	No	No	No	No	No	No	No	Yes	Yes	Yes
Simulates: Interflow	No	Yes	No	Yes	Yes	Yes	Yes	No	Yes	No
Evapo-transpiration from shallow aquifers	Yes	Yes	Yes	Yes	Yes		Yes	Yes	Yes	Yes
Regional groundwater flow	No	No	No	No	Yes	No	Yes			
Transmission losses	Yes	Yes	Yes	Yes	No	No	Yes	Yes	Yes	Yes

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*, 2010). Unfortunately, despite the urgent need for integrated surface water and groundwater resource management, integrated models applicable to the real world are scarce (DWAF, 2004a).

The approach adopted in this thesis is to use the modified Pitman model (Hughes, 2004) to simulate the interactions between surface and groundwater. This model was selected as it is one of the most widely used and trusted surface water models in South Africa for water resource assessments. It was utilised in both the WR90 (Midgley *et al*, 1994) and WR2005 (Bailey and Pitman, 2005) national water resource assessments, and has been used extensively throughout southern Africa (Bullock *et al*., 1990; Hughes, 1997; IHP, 1997; Mazvimavi *et al*., 2004; Tshimanga *et al*., 2011). Recent additions to the model (Hughes, 2004) offer the possibilities of simulating the range of interaction environments found in South Africa, as the model has been developed to represent the dominant processes that have been identified in the country (DWAF, 2004a). It is a compromise model which falls midway between simple and complex models and that potentially can provide information for managing surface and groundwater interactions in an integrated manner. While there are other models available that can simulate many of the key processes satisfactorily (GR4J, Le Moine *et al*., 2008; IHACRES_GW, Ivkovic *et al*., 2005), the Hughes (2004) version of the Pitman model was selected for use due to its depth of experience in South Africa and in the Institute for Water Research at Rhodes University where the research was undertaken. Essentially this thesis focuses on improving the conceptual understanding of surface and groundwater interactions in South Africa and subsequently assessing the extent to which the model and the scale at which it is applied can represent the major processes. Although the research in this thesis is exclusively carried out using the modified Pitman model, the model outputs are compared to other model outputs in some of the case studies and the 'methodology' which focuses on the critical evaluation of the catchment processes can be applied to any model.

3. CONCEPTUALISING AND MODELLING SURFACE AND GROUNDWATER INTERACTION

The conceptual representations of the processes occurring in a catchment are translated into mathematical form in a model. The perceptual model is based on our understanding of the real-world catchment system, i.e. flow paths, number and location of variables, runoff production mechanisms, etc. If this understanding is poor, particularly for aspects relating to sub-surface system characteristics, the perceptual model and subsequent algorithms in the model will be highly uncertain (Wagener and Gupta, 2005). This chapter focuses on the dominant processes which typically occur in interaction environments. In the following sections an outline of the conceptual understanding of typical interaction environments in South Africa has been given. The second part of the chapter describes the Pitman model and provides details of the model algorithms. The model structure is critically discussed with respect to the major processes identified in the first part of the chapter.

The interaction environments found over much of South Africa can be categorised into four “types” of interaction environment. 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). The different aquifer environments and some of the associated major processes are illustrated in Figure 3-1.

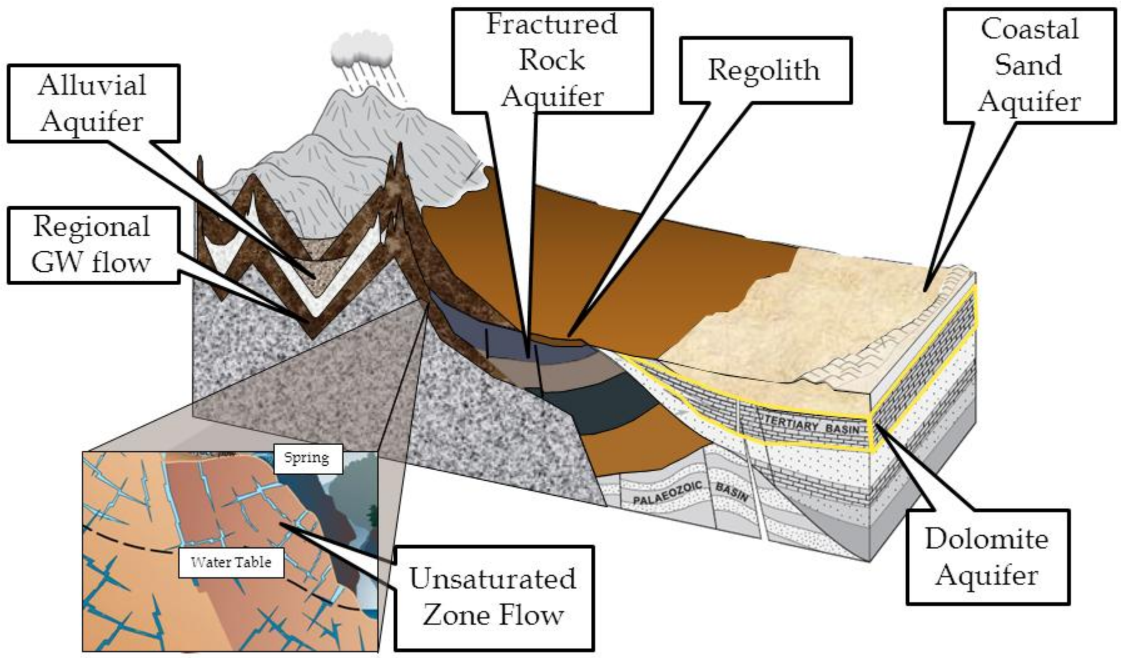


Figure 3-1 The major types of interaction environments encountered in South Africa.

3.1. Conceptualisation of fractured rock aquifers

Hard rock environments typically consist of three vertical zones, upper weathered, middle fractured and deep massive. Aquifers in the upper weathered zone (saprolite or regolith aquifers) often have similar properties to primary aquifers with intergranular porosity. These zones often correspond with the three vertical zones of flow which form part of general hydrogeological principles, schematised by 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 (upper) and fractured (middle) zones often form regional near-surface aquifers (up to hundreds of metres thick) which normally follow the surface topography. However, the local hydrogeological environment determines the geometry and structure of the aquifers and their parameters. For example, while it is often assumed that the shallow weathered zone will be more transmissive than the underlying fractured rock aquifer, Banks *et al*., (2009) found that the saprolite aquifer was less transmissive than the fractured rock aquifer beneath it. The extent and size of fractures generally decrease with depth. The deep or massive zone has very few fractures and faults. Deep fractures may act as isolated and more or less individual hydraulic bodies, although from a regional perspective can form interconnected networks which allow deep, regional groundwater flow reaching depths of hundreds or even thousands of metres (DWAF, 2008a). While the hydraulic conductivity within fractured rock aquifers follows well understood generalisations, the quantification of this hydraulic conductivity is not simple due to the geological heterogeneity of the fractures (Krásný and Sharp, Jr., 2003).

Banks *et al*., (2009) examined relative contributions to a gaining river from soil water, saprolite and a fractured rock aquifer in Australia. The authors found that although there was no hydraulic connection between the saprolite and fractured rock aquifers in the upslope parts of the catchment, all three sources contributed to the baseflow due to a convergence of the flow paths in the valley bottom prior to discharge within the river. Understanding flow in fractured rock has proven challenging. This is largely because the flow is controlled by high transmissivity fractures and we are currently unable to adequately predict where they occur and how they connect with other features, particularly other fractures. However, despite the high spatial variability in fractured rock properties, Krásný and Sharp, Jr. (2003) proposed that acceptable generalisations can be made about regional hydraulic and chemical properties and outline some of the generalisations they identified. Petrography can influence permeability and transmissivity in certain rock types. There are indications that relatively higher general transmissivity may be expected in basic igneous rocks. Quartzite hydrogeological properties differ depending on their extension. Where quartzite rocks are extended and exposed to intense fracturing they can form highly transmissive aquifer systems, with intensive and deep groundwater flow, due to their rigid geo-mechanical properties. Depth related decreasing permeability in granitic rock, orthogneisses and migmatities is more significant than in meta-sedimentary rocks. The regolith and weathered zone of these rocks (saprolite), often sandy and course grained, is permeable whereas meta-sediment regolith is typically more clayey and not as permeable. Fracturing in granites is usually not intense and reaches shallower depths than in other hard rock's.

Figures 3.2 and 3.3 illustrate some of the conditions encountered in typical fractured rock environments in South Africa. 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. Preferential flow in these environments can lead to numerous springs fed by unsaturated zone water, which means a large proportion of baseflow in these environments could be made up of interflow (Hughes, 2010a). High hydraulic gradients in mountainous headwater catchments can result in intensive groundwater flow in spite of the prevailing low transmissivity of rocks. Other processes which could be relevant in fractured rock environments include regional groundwater flows (in large folded systems and where large scale fault zones are present), preferential flow paths associated with dolerite intrusions and major fault zones, and confined aquifers (in the TMG of the Cape Fold Belt).

The volume of evapotranspiration from fractured rock aquifers will vary depending on the characteristics of the environment. In the Cape Fold Belt, recharge occurs during the winter rainfall season which means evaporation is low. Evapotranspiration can occur, however, directly from groundwater as groundwater levels in the area can be relatively shallow and the gaining nature of many of the streams suggests that riparian evapotranspiration can be important. In the semi-arid environments of the Karoo basin, rainfall rarely reaches the groundwater as recharge because it is removed by evapotranspiration during percolation in deep unsaturated zones.

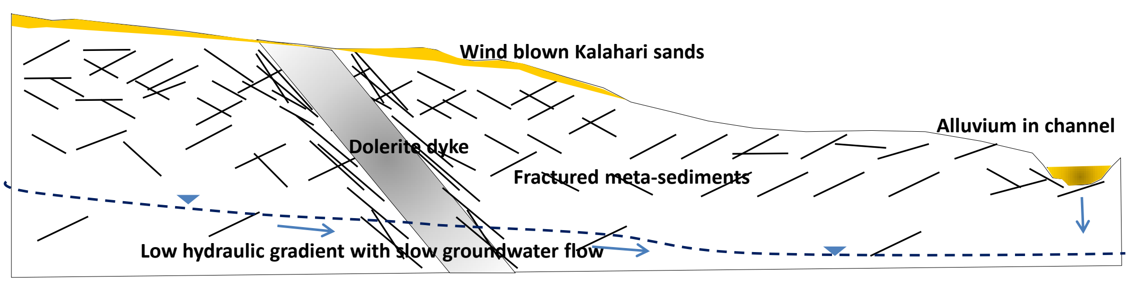


Figure 3-2 Groundwater flow in an unconfined aquifer in a semi-arid area of gentle slopes – losing system (e.g. Karoo Aquifer). The groundwater table reflects the topography, whilst very flat on a regional scale it conforms with the surface drainage showing an overall gradient towards the basin outlet. Locally, however, the groundwater flow directions are controlled by fracture orientations.

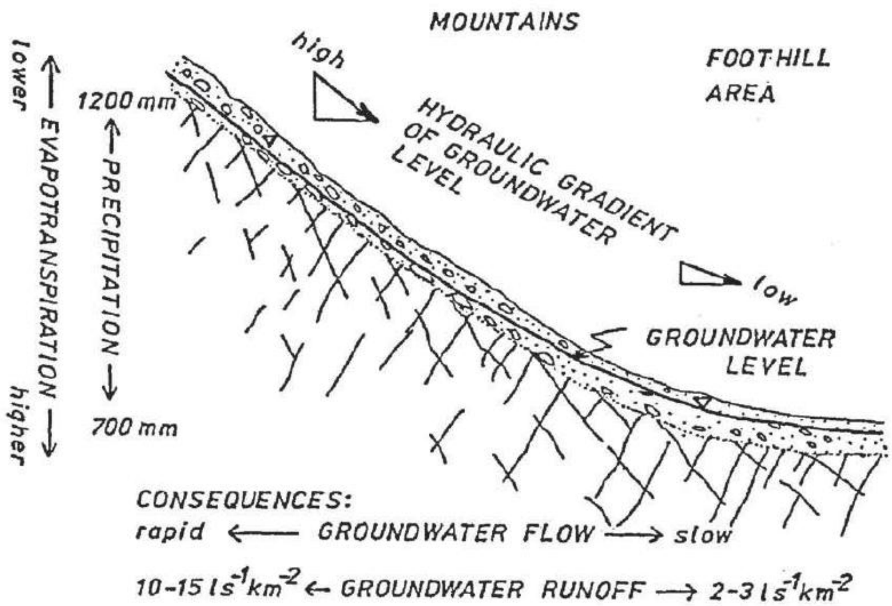


Figure 3-3 Groundwater flow in the near surface aquifer of mountainous hard rock terrains under temperate climatic conditions (Krásný and Sharp, Jr., 2003).

Without detailed field characterisations, any model applied in a fractured rock setting will be uncertain. Modelling of fractured rock aquifers often requires the scaling up of data from local site studies. Consequently, aquifer parameters might differ considerably depending on the methods of their determination. The comparison of regionally prevailing transmissivity indicates only small differences in distinct hard rock areas. The scale dependence of transmissivity is a well-known phenomenon due to heterogeneity (Sanchez-Vila *et al.*, 1996) and studies have shown that the scale dependence of transmissivity persists at the regional scale (Manga, 1997). As a result 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 (Shchoniger *et al*. 1997) and the effects of structures such as faults and fractures are averaged.

Cook (2003) identified an increase in average permeability of around three orders of magnitude from the laboratory scale to the borehole scale and estimated average transmissivity measured at the borehole scale between 10^{-2} and 10^{-7} m^2 day^{-1}. Singhal and Gupta (1999) suggest typical hydraulic conductivities of fractured siltstones and shales of 0.09 and 8.64 m day^{-1}. In South Africa, estimates of the hydraulic characteristics of the major fractured rock aquifers have included hydraulic conductivity values of 2.65 * 10^{-4} to 1.22 * 10^{-3} m^2 day^{-1} and a storativity of 0.001 for the TMG (Lin *et al.*, 2007).

Transmissivities of Karoo aquifers have been estimated at 15 m^2 day^{-1} (Sami and Hughes, 1996) and between 5 and 50 m^2 day^{-1} (Van Tonder and Kirchner, 1990), while storativity has been estimated at between 0.0001 (Botha and Cloot, 2004) and 0.001 (Sami and Hughes, 1996).

A summary of processes that could be important in these environments for large scale modelling includes:

- Perched aquifers and interflow from the unsaturated zone above the regional groundwater.
- Direct recharge into bare rock areas.
- Preferential recharge in highly fractured areas.
- Preferential flows and variable aquifer characteristics associated with regolith aquifers.
- Evapotranspiration directly from groundwater in riparian margins.
- Discrete groundwater discharge into surface water systems.
- Regional groundwater flows.
- Preferential flow through fault systems, dykes and sills.
- High spatial variability.
- Confined aquifers.

3.2. Conceptualisation of karst aquifers

The conceptual framework of karst aquifers seems to be well defined and generally accepted, while the modelling of karst aquifers is still highly uncertain. Surface and groundwater interact heavily in most karstic systems and it is not straightforward to delineate catchment boundaries which encompass both the surface water and groundwater system. Groundwater catchment divides have been shown to shift as storage volumes change, since groundwater spill into adjacent basins can occur during high recharge periods (White, 2002, Bailly-Comte *et al*., 2009). Larocque *et al*. (1998) documented large changes in aquifer hydraulic conductivity and storage properties with changing water levels in a karst aquifer. 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/km2) and can be used to estimate the approximate recharge area of springs. Figure 3-4 illustrates some of the typical karst environments found in South Africa.

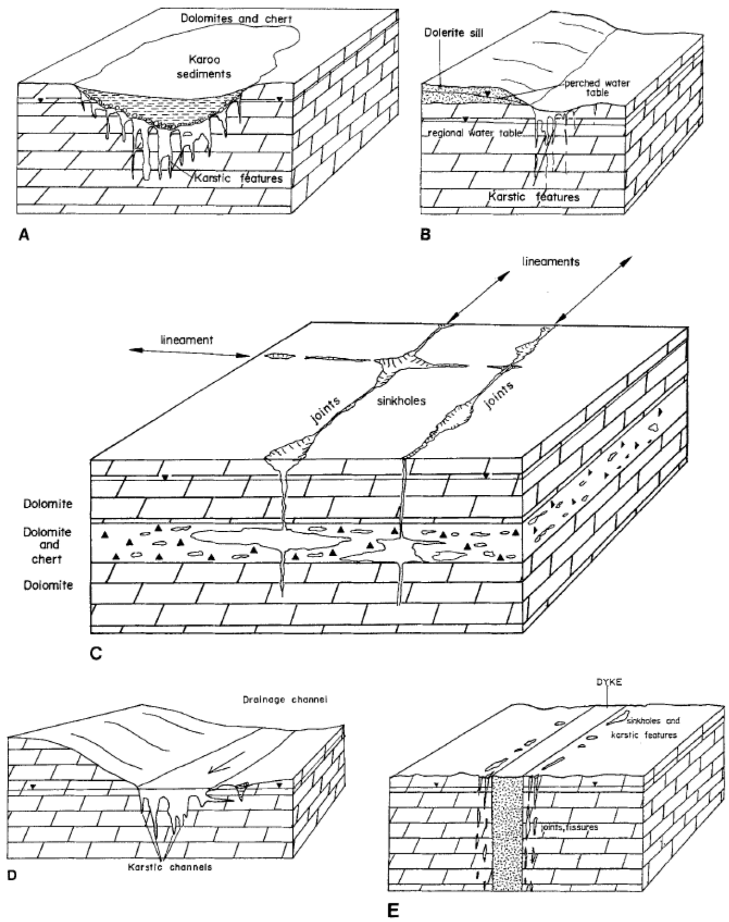


Figure 3-4 Common types of karstification in South Africa: (a) underneath Karoo Sequence outliers, (b) on sill margins, (c) along lineaments and joints and within chert rich dolomites, (d) underneath drainage channels, (e) on dyke margins (Kafri and Foster, 1989).

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). While it is generally accepted that groundwater recharge is very high in karst terrain due to precipitation infiltrating readily through rock openings that intersect the land surface, White (2002) identified four specific types of recharge that occur in karst environments. These include (1) allogenic recharge from upstream non-carbonate parts of a catchment, (2) diffuse infiltration which includes direct infiltration from precipitation into the soil and carbonate aquifer, (3) internal runoff, where surface runoff flows into closed depressions and enters the aquifer through sinkhole drains, and lastly, (4) overflow from perched aquifers which can be common in certain karst environments. There are three different types of flow in karst aquifers which vary depending on the specific karst system. These include, intergranular permeability (often, but not always small in relation to the other permeability types), fracture permeability (similar properties to fractured rock aquifers) and conduit permeability. The onset of non-Darcian behaviour occurs when the aperture exceeds about 1 cm (White, 2002). Conduits in karst aquifers frequently make up only a very small percentage of the aquifer but completely dominate the flow behaviour where they are present. As a result, discharge from the aquifer may be made up of relatively slow moving water draining from pores and rapidly derived storm water (Wong *et al*., 2012). 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).

Determining the hydraulic conductivity of a karst system is difficult if not impossible. . Similar to fractured rock systems, the measurement of effective permeability is completely scale dependant (Sauter, 1992). While karst aquifers often have a high storage capacity, their capacity for water retention is low. Monitoring groundwater levels can give some indication of storage levels as the more variable a water level, the lower the storage capacity (Bonacci, 1993). Few measurements of karst conduit permeability are available in the literature, especially in South Africa. According to Guyot (1985), Josnin *et al*. (2000), and Kiraly (2003), the order of magnitude of conductivity for a karst conduit network should be 86 m day^{-1} to 864 m day^{-1}. Other estimates of hydraulic conductivity include 0.3 (matrix) to 305 m day^{-1} (Scanlon *et al*., 2003); 20 to 100 m day^{-1} (Bishop and Lloyd, 1990) and a mean value of 7 m day^{-1} with a matrix conductivity of 0.001 m day^{-1} for a thoroughly tested karst aquifer in Texas, USA. Transmissivity estimates in the USA have ranged from 10 m^2 day^{-1} in poorly fractured areas to 3000 m^2

day^{-1} for boreholes that tap solution channels (Cook, 2003). In Australia transmissivity determined from pumping tests ranged from 200 m^2 day^{-1} to more than 10 000 m^2 day^{-1}. In contrast a much less developed karst aquifer in Australia had a mean hydraulic conductivity of 1 m day^{-1} (Cook, 2003). In South Africa, Bredenkamp (2007) estimated the storativity of a typical karst aquifer at 0.05 but stated that transmissivity values were too variable to give reliable estimates.

Seeps and springs of all sizes are characteristic features of karst terrains. Large spring inflows to streams in karst terrain contrast sharply with the generally more diffuse groundwater inflow characteristics of streams flowing across other types of aquifers. A number of springs can sometimes drain a single groundwater catchment, with rates of discharge varying by several orders of magnitude. 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). If the groundwater catchment size is known, a relatively reliable characterisation of the aquifer can be undertaken. . Bonacci (1993) quantified the storage properties and hydraulic conductivity of a karst aquifer using only observed hydrographs (Figure 3-5) and stage discharge curves.

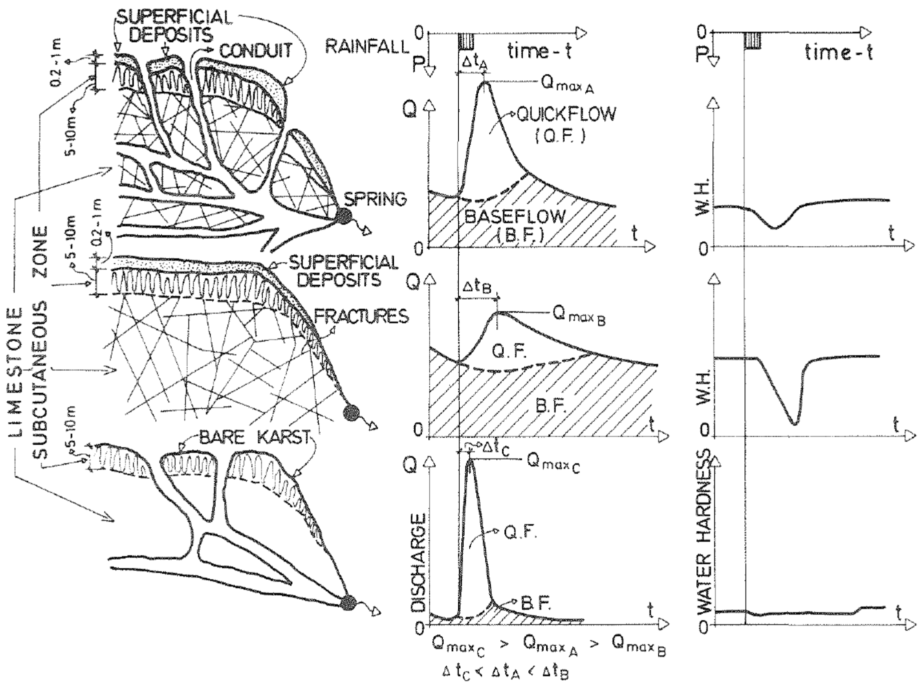


Figure 3-5 Various forms of discharge hydrographs for (a) combined, (b) diffuse and (c) conduit type karst springs (Bonacci, 1993).

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 (Katz *et al*., 1998). Numerical modelling has been used (Eisenlohr, 1997) to characterise karst aquifers. However, due to the lack of necessary data required by these models (aquifer geometry, position and direction of fractures and open channels, hydraulic properties), they are often used in a simplified mode, with large simplifications of the physical and hydrogeological structures that these methods aim to represent. Conceptual models can be based on a simplified but efficient representation of these systems using interconnected reservoirs and flow hydrographs (Jukic and Denic-Jukic, 2006; Rimmer and Salingar, 2006; Le Moine *et al*., 2008). These models are easier to implement and can be used for hypothesis testing to explore the complexity of such hydrological systems. Le Moine *et al*. (2008) set up a rainfall-runoff model for the Touvre karstic spring in western France (a tributary for a major river), and found that a lag and route model with a linear store was the best option. However, many investigations carried out using rainfall-runoff models have been less successful with particular challenges including temporal uncertainy, non-linearity (non-proportional effects) and non-stationarity (lack of time invariance). Labat *et al*. (2000) reviewed rainfall-runoff models for the purpose of characterising karst aquifers and concluded that these models can successfully assess karst aquifers if they are physically based and include non-linear and non-stationary functions.

A summary of processes that could be important in these environments for large scale modelling includes:

- Different surface and groundwater catchment divides.
- The presence of dykes and sills which can compartmentalise the karst aquifer.
- Highly concentrated recharge.
- Highly concentrated (springs) discharge into surface water systems.
- Possible lag effects between recharge and discharge.
- Perched aquifers and interflow.
- Extremely high transmissivity and storativity.

3.3. Conceptualisation of alluvial aquifers

Alluvial valleys comprise river deposits (called alluvium) ranging in size from clay to gravel. Larger particles are deposited in faster flowing water typically in a river channel, and smaller particles are deposited in slower moving water, such as water inundating a flood plain. Topographic slope exerts a major control on water velocity and particle size. A river flowing down a mountain and onto a gently sloping plain would deposit large amounts of sediment at the base of the mountain, where the slope decreased abruptly. Large accumulations of sand and gravel in alluvial fans can be prolific aquifers. Alluvium can also fill valley bottoms in large basins to varying widths and depths depending on the type of environment, and can form extensive aquifers surrounding a river channel for many kilometres (Figure 3-6). Over time, river channels move both laterally and vertically which often results in a complex sequence of sedimentary deposits. Alluvial aquifers are often made up of multiple sand and gravel lenses which represent positions of the former river channel, and the finer deposits represent ancient flood plains. The result is variable permeability associated with coarse sands and gravels to finer material and clays. Stream alluvium is usually more 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.

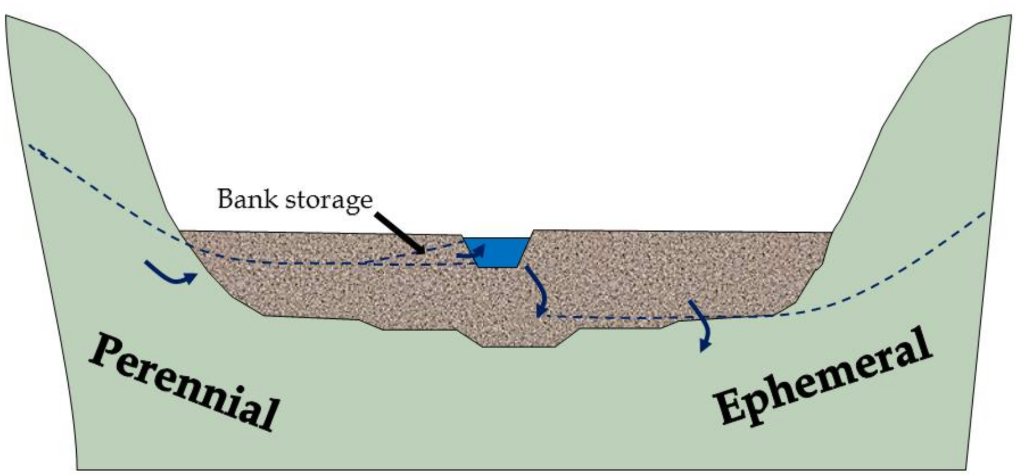


Figure 3-6 Simple conceptualisation of an alluvial aquifer in both a perennial (left bank) and ephemeral (right bank) environment.

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 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).

Larger scale investigations mostly involve water balance estimations. The resulting regression equations or simple hydrological flow routing approaches are often site specific and blurred by unknown lateral inflows and require further verification and modification before being applied to other arid channels. However, a recent investigation by Dahan *et al*. (2008) carried out on the Kuiseb River in Namibia, contradicts many of the arguments against a generally applicable perceptual model. The authors examined infiltration controls and recorded similar fluxes for both small and large floods. Ultimately the study 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's stage. The storage capacity is determined by the width of the aquifer and the thickness of the unsaturated zone.

While water losses in flood plain areas might not contribute significantly to groundwater recharge of the surrounding aquifer, they are important in that they can significantly increase transmission losses and decrease flood discharge downstream. Both Knighton and Nanson (1994) and Lange (2005) identified that significant transmission losses only occurred when the discharge exceeded a certain limit (100-120 $m^3\ s^{-1}$; Lange, 2005), with losses being minor during small to medium flows. This was attributed to enhanced water losses by infiltration in flooded overbank areas. These conclusions are fairly typical in many alluvial environments, where small to medium flows are concentrated in inner, confined channels where transmission losses are restricted by the limited area available for infiltration and presumably by clogging layers on and within the channel alluvium. The degree to which the flood plain infiltration recharges groundwater seems to vary between locations, perhaps connected to the aridity of the environment. Lange (2005) concluded that high magnitude events are decisive for recharging groundwater under the flood plains along large desert streams. Dahan *et al*. (2008), however, concluded that these flooding zones beyond the width of the active channel contribute very little to groundwater recharge although the losses will result in lower flood discharge downstream. The inundation of floodplains can, however, exert a considerable influence on flow patterns, by influencing the return flows from the surrounding floodplains back into the channel. These hysteretic effects can be quite significant especially for floodplains with closed depressions below banktop. Hughes (1980) showed how the movement of inundating water depends on the shape of the flood hydrograph and the prevailing surface form of the floodplain and these determine the balance between inflow and outflow and hence the volume at any one time on the floodplain. Hysteretic relationships between stored volume and channel discharge result, the degree of hysteresis depending upon the conductivity of the floodplain/flood event combination.

A general conceptual understanding of the processes occurring during a recharge event in an alluvial aquifer in a dry environment include; (1) propagation of a wetting front from the stream channel down through the unsaturated zone toward the groundwater; (2) a rise in water table due to the recharge process; and (3) water level relaxation and stabilisation at the new water level as the surrounding aquifer and the newly recharged alluvial aquifer adjust. At a detailed scale, the water content variation in the sediments during percolation is a function of the flow conditions above and below each layer, as well as the physical characteristics of the layer itself, such as porosity and grain-size distribution. While the flow conditions above and below each layer control inflow and outflow, the porosity and grain size distribution control the layer's water retention, field capacity, and degree of saturation.

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. The main Limpopo River however, is predominantly a perennial system along with 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. These alluvial fans which together make up the Breede River Alluvial aquifer are recharged by rain, by surface water runoff, perennial spring flow and by transmission losses along the Breede channel (DWAF, 2008b). The majority of the alluvial aquifers in South Africa, however, lie along ephemeral streams in the Kalahari and Limpopo Basins where transmission losses make up the majority of the groundwater recharge. 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 flood plain 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.

Interactions between the alluvium and surrounding aquifer can be important in many environments. In South Africa, the surrounding aquifers are mostly fractured rock with a few karst aquifers in some regions. In areas such as the Cape Fold Belt, the surrounding TMG provides an important source of recharge to the alluvial aquifers (this connection can reverse with seasonality). In the drier parts of the country, transmission losses are often the only substantial form of recharge for the alluvial aquifers beneath stream channels and the aquifers connected to them. Transmissivity rates in alluvial aquifers are generally high and have been reported at between 86 to 4000 m^2 day^{-1} (Razack and Huntley, 1991) and from 1 to 1800 m^2 day^{-1} in Colorado (Konikow, 1977).

A summary of processes that could be important in these environments for large scale modelling includes:

- Connections with underlying aquifers which can recharge the alluvium or be recharged by the alluvium.
- Transmission losses, both channel losses and flood plain losses.
- Diffuse discharge into surface water systems.
- Characteristics of flood events (duration and intensity).
- Hysteretic effects.

- Bank storage.
- High evapotranspiration from lowland flood plains and riparian margins.
- High transmissivity and storativity.

3.4. Conceptualisation of primary aquifers

Primary aquifers can exist in a variety of different geological environments, such as glacial sediments, chalk sediments or deep sand basins. The focus in this thesis will be on coastal and dune aquifers as they are the principal primary aquifer types in South Africa. Coastal terrain is often characterised by streams, estuaries, wetlands and lagoons that are affected by tides, as well as ponds commonly associated with coastal sand dunes (Figure 3-7). Dune terrain is characterised by hills and depressions in the landscape. These depressions commonly form lakes or wetlands which can interact with groundwater in a variety of ways depending on the local setting. While the landscape may be undulating, the regional land surface is generally flat with the water table close to the surface. The saturated zone groundwater flow obeys Darcy's Law and therefore this type of aquifer is unlikely to have a pronounced or prolonged interflow component in the stream discharge. 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. These losses can form cones of depression around the perimeters of lakes and wetlands in some environments (Winter *et al.*, 1998). Recharge and discharge within primary aquifers can be described as diffuse in comparison to the preferential nature of fractured rock environments.

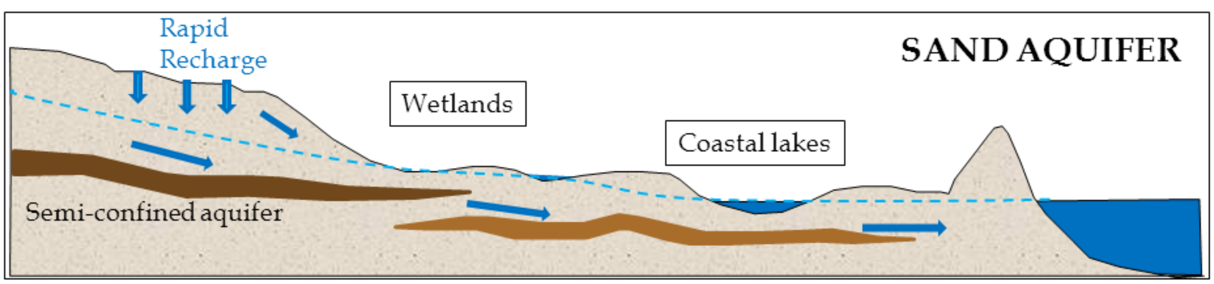


Figure 3-7 Simple conceptualisation of a typical coastal primary aquifer.

Coastal and dune aquifers can be further sub-divided based on their relative heterogeneity. Aquifers can be generally classified as homogeneous, uniform and porous, or heterogeneous, non-uniform and porous (Kelbe and Germishuyse, 2010). The more homogeneous aquifer type is often a sand aquifer with little clay or silt. These types of "ideal" systems do display spatial variability in flow (fingering) at the local and macroscopic scales. Heterogeneous systems will have a degree of lateral flow distinguishing them from the purely vertical flow system in homogeneous aquifers, although the degree of heterogeneity would need to be large to cause lateral flow. Heterogeneous porous aquifers often have significant amounts of clay and silt, which can lead to perched aquifer conditions and confined conditions. There can be large variability in hydraulic properties such as porosity and permeability. However, at the larger scale, even a largely heterogeneous porous aquifer is relatively homogenous compared to a fractured rock or karst environment, Transmissivities of sand, silt and clay have been reported at between 0.09 and 8000 m^2 day^{-1}.

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. Groundwater flow is relatively uniform and therefore relatively easy to characterise as they are more homogeneous than secondary aquifers. Often the sediments are highly permeable which promotes rapid recharge to the aquifers. In all of these primary aquifers, aquifer thickness varies considerably. 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 an reliable water balance for the region as a whole (Munch *et al*., in press) .

Rainfall-runoff models with groundwater components have often been described as assuming a homogeneous uniform aquifer (DWAF, 2004) which would seem to be ideal for application in an environment with a primary aquifer. While these types of models have been applied extensively in heterogeneous fractured rock (Manga, 1997), karst (Le Moine *et al*., 2008) and alluvial (Lange *et al*., 1999) environments, their application in porous aquifer settings often turns out to be just as complicated. Often it is necessary to incorporate additional components into a model, such as wetland and/or reservoir (lake) functions. In addition, coastal aquifers are often situated in highly populated areas so there is large disruption to the natural flow regime. Consequently, the use of models in these more uniform environments often has the same level of complexity as the more heterogeneous environments.

A summary of processes that could be important in these environments for large scale modelling includes:

- Low probability of unsaturated zone flow.
- Relatively diffuse recharge.
- Shallow groundwater tables.
- High evapotranspiration losses directly from shallow groundwater and the riparian margin.
- Diffuse discharge into surface water systems.
- The presence of lakes, wetlands and estuaries.
- Perched aquifers, confined and semi-confined aquifers on a small scale.
- Connections with underlying aquifers.
- High transmissivity and storativity.

3.5. The Pitman Model

3.5.1. Introduction

The monthly time-step Pitman rainfall-runoff model (Pitman, 1973) has been used for water resource availability estimation in South Africa for many years. It has also been used in investigations in many other countries in southern Africa (IHP, 1997; Mazvimavi *et al*., 2004; Tshimanga *et al*., 2011). It is a conceptual type model with parameters representing the main storages and fluxes that constitute the natural water balance of river basins. The model originally incorporated groundwater components in a highly simplified way which rendered it unable to simulate many important processes necessary for IWRM. As a result the model was modified by both Hughes (2004) and Sami (2006) by incorporating more explicit groundwater routines in an attempt to create a more integrated tool for use in IWRM. The modification or improvement of an existing, widely trusted model was deemed preferable to the development of a new model as the model foundations are already established and robust and only the new portions would need to be tested and debugged. The model (Hughes, 2004) was developed with a conscious awareness of the lack of data that plagues the application of models in the southern African region, which meant it had to be based on a robust conceptual understanding of surface and groundwater interaction processes, had to be somewhat physically based and had to be able to operate within an uncertainty framework. The model represents a compromise between complex physically based distributed models and simple empirical approaches. While the model has been designed to utilise existing databases of regional information for the purposes of national water resource assessments and many of the input parameters can be derived from maps or by field work, calibration is still typically necessary. The model outputs are therefore assessed based on either comparison with real data (if available), outputs from other models that have already been tested or, at the very least, against our conceptual interpretation of reality.

The model is applied using an integrated modelling software package called SPATSIM (Spatial and Time Series Information Modelling; Hughes, 2005) that links spatial data with other data types (parameter tables and time series, for example) and includes a variety of data input, output and analysis routines as well as links to hydrological and water resource simulation models. Current versions of the model also include components that allow artificial impacts such as small farm dams, larger dams, abstractions and return flows (Hughes and Mantel, 2010b; Hughes *et al*., 2010) to be included in the modelling scheme. In addition, a wetland component has been developed (Hughes *et al*., 2013) which can account for the impacts of both wetlands and lakes. The model can be applied within an uncertainty framework (Kapangaziwiri and Hughes, 2009) while detailed physically based parameter estimation routines have been developed to fit into this framework (Kapangaziwiri and Hughes 2008). These routines attempt to translate uncertainty in the available physical basin property data into uncertainty in the resulting estimates of parameter values.

Both the Hughes (2004) and Sami (2006) approaches have been incorporated into the Department of Water Affairs (DWA) official version of the model that forms part of the WR2005 national database and analysis tools (Bailey, 2007). The groundwater parameter values can be obtained from a groundwater database generated during the Groundwater Assessment Phase II (GRAII) project (Conrad, 2005) in South Africa and provide estimates of storativity, transmissivity, maximum depth to groundwater and annual recharge at a large spatial scale (approximately 100 to 1000 km^2). Much of the information contained in this section is drawn from documents, published and unpublished, that have described modifications to the original model (Hughes, 1997; Hughes, 2004; Hughes *et al*., 2007; Kapangaziwiri and Hughes, 2008; Hughes *et al*., 2010 and others). Figure 3-8 illustrates the main components of the revised model.

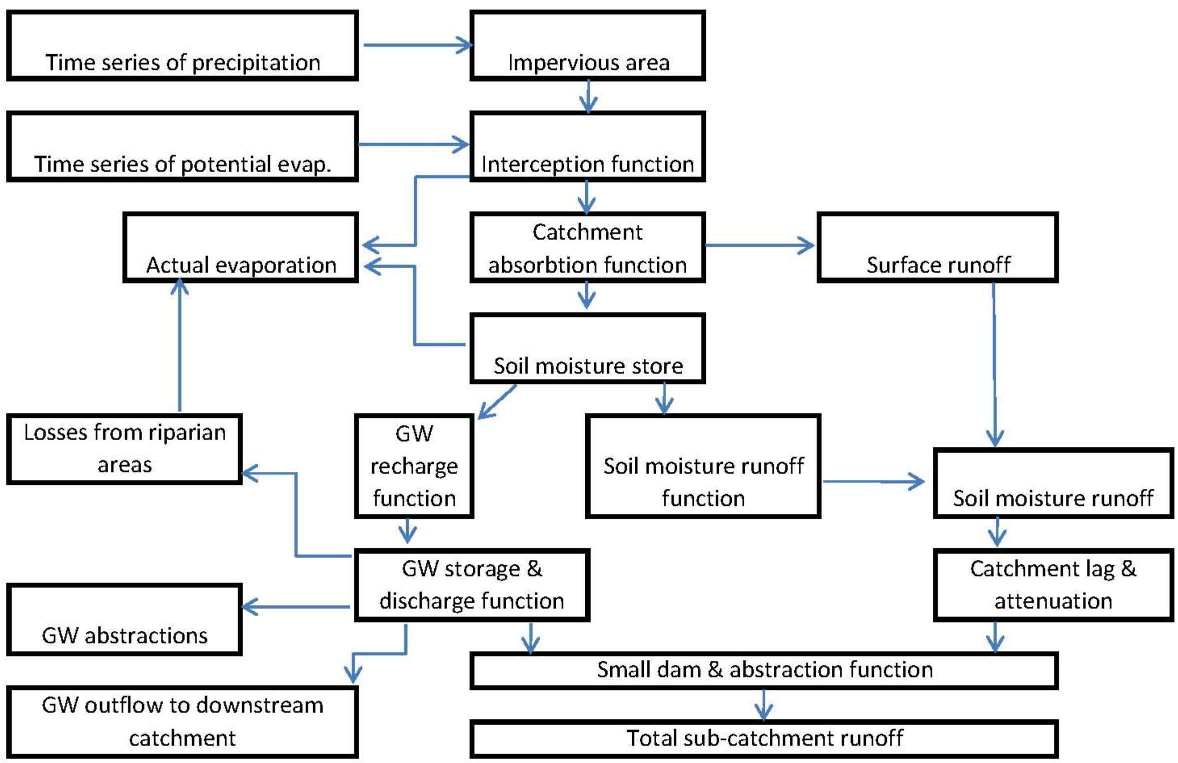


Figure 3-8 Conceptual structure of the revised Pitman Model (adapted from Hughes, 2004).

3.5.2. The surface water components of the Pitman model

The Pitman model operates at a catchment scale, uses a monthly time step, and the parameters represent spatial averages rather than direct point values from field measurements. It is assumed that sub-catchment effects can be accounted for through probability distribution functions which represent the largely unknown variations in process functioning within a spatial unit. This approach considers the heterogeneity of the catchment response (sub-grid effects) without mapping the heterogeneity onto specific locations in the catchment. Geology, topography, vegetation cover, soil type and texture will all influence patterns of moisture distribution within a catchment. It is therefore reasonable to suggest that, for any given mean basin moisture content, the spatial variation could be represented by a frequency distribution. This would mean variability would be low when the catchment is either very dry or close to saturation (extreme ends of moisture content range) and highly variable at moderate moisture contents. However, the integration of the variability of these properties on a catchment wide scale is not straightforward and the degree to which frequency distributions can represent the variability on such large scales is uncertain.

Table 3-1 provides a list of the Pitman model surface and groundwater parameters. Additional compulsory data requirements include basin area, a time series of basin average rainfall, seasonal distributions of evaporation (fraction) and monthly parameter distribution factors. The model operates over four equal time steps within a month in order to approximately represent variations in the rainfall (the distribution is controlled by parameter RDF), and to avoid excessively large changes in any one component of the water balance. This is considered necessary in a coarse time-step model given that natural processes occur concurrently, while the model simulates them sequentially. The model simulates upstream and downstream sub-catchments simultaneously with a routing of the surface water and groundwater between sub-catchments. A brief explanation of all the model components is provided below; however the groundwater functions are reviewed in more detail.

Table 3-1 **The modified Pitman model surface and groundwater parameters.**

Parameters	Units	Description
Surface water parameters		
RDF		Rainfall distribution factor. Controls the distribution of total monthly rainfall over four model iterations
AI	fraction	Impervious fraction of sub-basin
PI1s	mm	Summer Interception storage for vegetation type 1
PI1w	mm	Winter Interception storage for vegetation type 1
PI2s	mm	Summer Interception storage for vegetation type 2
PI2w	mm	Winter Interception storage for vegetation type 2
AFOR	%	% area of sub-basin under vegetation type 2
FF		Ratio of potential evaporation rate for Veg2 relative to Veg1
PEVAP	mm	Annual basin potential evaporation
ZMINs	mm month^{-1}	Summer minimum basin absorption rate
ZMINw	mm month^{-1}	Winter minimum basin absorption rate
ZAVE	mm month^{-1}	Mode of distribution of absorption rates
ZMAX	mm month^{-1}	Maximum basin absorption rate
ST	mm	Maximum moisture storage capacity
SL	mm	Soil moisture below which there is no recharge
POW		Power of the moisture storage runoff equation
FT	mm month^{-1}	Runoff from moisture storage at full capacity
R		Evaporation-moisture storage relationship parameter
TL	months	Lag of surface runoff
Groundwater parameters		
GW	mm month^{-1}	Maximum recharge depth at maximum moisture capacity
TLGMax	mm	Maximum channel loss
GPOW		Power of the moisture storage recharge equation
DD	km km^{-2}	Effective drainage density
T	m^{2} day^{-1}	Transmissivity
S		Storativity
RG		Regional groundwater drainage slope
Rest RWL	m below surface	Aquifer depth
RSF	% slope width	Riparian Strip Factor

Interception: A proportion of any precipitation input does not reach the basin surface because it is intercepted by the vegetation cover. This function is based on the interception parameter PI, which can vary seasonally and have values for two different vegetation types (typically but not necessarily, natural vegetation and plantation forest). The model assumes that: (1) the total rainfall on any rain day is concentrated in one storm event only, and (2) all the intercepted rainfall is evaporated (at the potential rate, PEMAX) before the next rain day. Total monthly interception loss is assumed to be determined by interception storage capacity (PI) and the total rainfall.

Infiltration: Figure 3-9 illustrates the approach to surface runoff generation that uses a triangular distribution (parameters ZMIN, ZAVE and ZMAX) to define the frequency of catchment absorption rates. The rainfall rate can then be used to calculate what proportion of the catchment will exceed the absorption capacities as well as the volume of surface runoff.

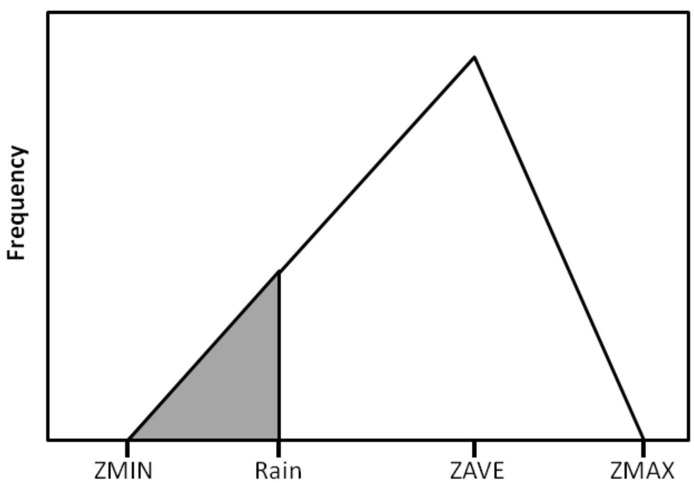


Figure 3-9 Format of the Pitman model algorithm for surface runoff (Hughes *et al.*, 2007).

Evapotranspiration: Figure 3-10 illustrates the simple linear relationship used to estimate actual evapotranspiration loss from the main (soil) moisture store. The evaporation function depends on the current month's potential evaporation value relative to the month with the highest potential evaporation together with the values of parameters R and FF. R defines the relationship between the ratio of actual evaporation to potential evaporation and the level of the soil moisture store (S). As the potential evaporation (PE) rate reduces, the parameter ($0<R<1$) defines the rate at which actual evapotranspiration declines as the relative moisture content decreases. A further parameter (FF) represents an evaporation scaling factor for a second vegetation type, frequently used to represent plantation forestry. The proportion of the basin area covered by the second vegetation type is given by the parameter AFOR.

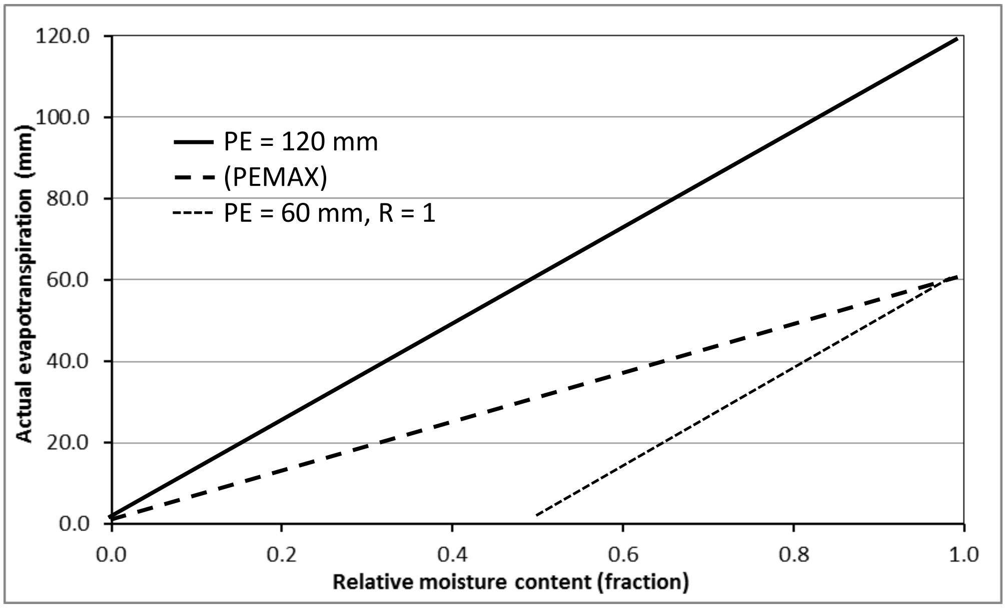


Figure 3-10 Relationship between catchment evapotranspiration (ET) and soil moisture (S) for R = 0 (A) and R = 1 (B) (Hughes *et al.*, 2007).

Equation 3 defines the algorithm for the estimation of actual evapotranspiration (all values in equations 3 and 4 are in either mm (for storages) or mm per month for moisture fluxes.

$$ET = PE * [1 - \{1 - R * (1 - PE/PEMAX)\}^{-1} * (1 - S/ST)] \qquad \textbf{Equation 3}$$

Including the effect of the second vegetation type the total evapotranspiration is ET_{total} is given by:

$$ET_{Total} = ET * FF * AFOR + ET * (1 - AFOR) \qquad \textbf{Equation 4}$$

Where: ET = Actual evapotranspiration

PE = Potential evaporative demand

PEMAX = Maximum potential evaporative demand

S = Current unsaturated zone moisture storage

ST	=	Maximum capacity of the unsaturated zone moisture storage
R	=	Evaporation parameter
FF	=	Evaporation scaling factor for the second vegetation type
AFOR	=	Proportion of basin area covered by the second vegetation

3.5.3. The surface and groundwater interaction components of the Pitman model

3.5.3.1. Unsaturated zone storage and runoff

The model assumes that water draining from the soil or unsaturated zone would have two directional components, a vertical one contributing directly to groundwater recharge and a lateral one as soil interflow or re-emergence as springs or seeps, which occur at elevations above the groundwater level. Horizontally aligned fractures and perched aquifers associated with layers of lower permeability (which are possible in most interaction environments) can account for some of the lateral water movement. Infiltrating water is added to the soil moisture storage (S) which has a maximum of ST (mm). This storage is depleted by evaporative losses, runoff and recharge to the groundwater store. The model assumes that S can represent moisture stored not only in the soil but also within the unsaturated zone above the groundwater table. While, the parameter estimation techniques developed for the model (Kapangaziwiri and Hughes, 2008) estimate ST_{soil} and ST_{unsat} values separately, both are represented in the model under the one parameter (ST).

The soil moisture (ST_{soil}) storage can be estimated from some knowledge of the soils or of the geology and climate of a catchment. Soils developed from geological formations that are easily weathered would be expected to have deeper soils and higher values of ST, as would areas where the prevailing climate promotes deep weathering. Deeper soils would also be expected in flat landscapes and on valley floors unlike steep slopes and mountainous areas. The unsaturated zone component (ST_{unsat}) will be influenced by the storativity (S) of the underlying geological formation and the depth to the water table. Storativity refers to the volume of water that a permeable geological formation will absorb or expel from storage per unit surface area per unit change in head and is equal to the product of specific storage and aquifer thickness. The storativity of the unsaturated zone will determine the volume of unsaturated zone storage that could potentially contribute to runoff, which will consist largely of fracture zones. ST_{unsat} is expected to be zero in primary aquifers.

Runoff from the soil/unsaturated zone moisture store is simulated using a non-linear relationship between discharge and soil moisture content (Figure 3-11). FT refers to the interflow generated when the moisture level (S) is at its maximum value (ST). FT must therefore represent the maximum possible runoff (mm) per month from both the soil moisture and unsaturated zone storage. The power (POW) parameter represents the shape of the relationship that determines reduced runoff (relative to the maximum) as the moisture contents of the soil zone and unsaturated zone decrease.

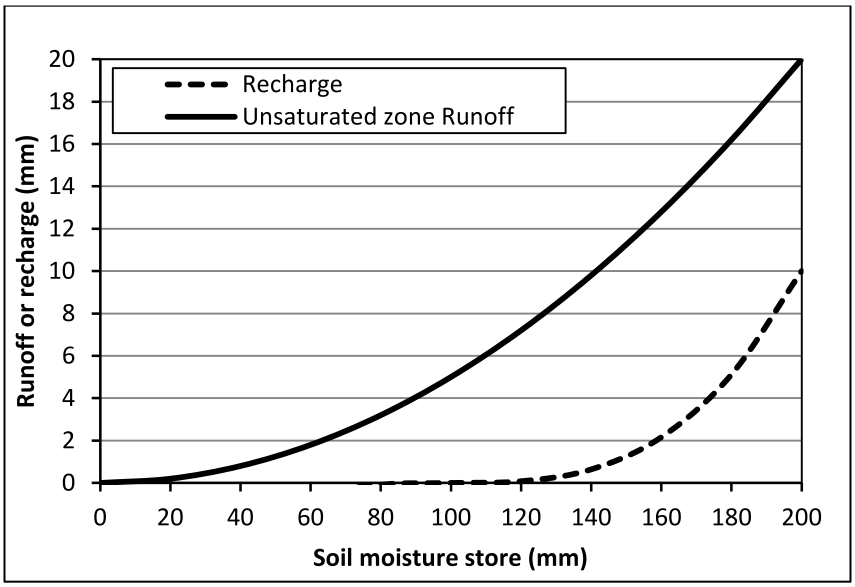


Figure 3-11 Illustration of the soil moisture runoff function (with parameter ST=200, SL=0, FT=20 and POW=2) and the recharge-moisture state relationship with parameter SL=100, GW=10 and GPOW=3 (Hughes *et al*., 2007).

Runoff from the soil moisture store is determined by the following equation;

$$Q = FT*(S/ST)^{POW} \qquad \textbf{Equation 5}$$

where S is the current soil moisture store.

Beyond ST all excess rainfall is converted directly to runoff. In reality therefore FT cannot be greater than ST as this would mean runoff greater than the amount of moisture available to generate that runoff. All other factors being constant an increase of POW will result in an increase in discharge. POW can be assumed to represent the relationship between total basin moisture status and the spatial distribution of this moisture. While it is not straightforward to determine unsaturated zone parameter values from physical characteristics at the basin scale, characteristics such as porosity, topography, soil types and depths can give indications of the expected range of values. Similarly, the generation of interflow will be dependent upon characteristics such as vertical variations in permeability, fracture orientation, as well as the degree of interconnectivity and connectivity with the surface channel network.

3.5.3.2. Recharge

There are four parameters which directly influence the volume of groundwater recharge. These are ST, SL, GW and GPOW. SL, represents the lower limit of soil moisture storage below which no recharge is expected to occur. GW is defined as the maximum amount of recharge (at a moisture status equal to ST) and GPOW determines the form of the relationship between the volume of recharge and moisture stored in the unsaturated zone S (Figure 3-11). The depth of recharge can then be estimated as a non-linear relationship with the ratio of current storage to the maximum:

$$\boldsymbol{RE = GW\ \frac{S-SL}{ST-SL}^{GPOW}}$$ **Equation 6**

Where RE is the monthly recharge rate in mm and S is the current soil moisture storage level in mm.
GPOW is very similar to POW and expected to reflect similar physical relationships. As the moisture status in the entire catchment declines, the proportion of the basin with soil moisture states above field capacity will decrease. During a dry period in a catchment, it is possible that there will be no parts of the basin that have soil moisture states above field capacity and no unsaturated zone fractures which contain enough water to generate vertical drainage. At this point no groundwater recharge will be generated. The type of data typically required to define the vertical structure of the unsaturated zone and its relationship with surface topography is rarely available. This is further complicated by the high non-linearity and spatial variability of the recharge process and its close association with the other outputs (interflow and evapotranspiration losses) from the soil water storage (S).

The value of SL is normally set to zero because the rates of recharge at low soil moisture are small and have little influence on the total water balance of the basin (Hughes and Parsons, 2005). A general conceptual understanding of large scale recharge processes can however, assist with parameter estimation to a degree. Topography is expected to have a large impact on local and regional flow paths and recharge is thus generally expected to occur in topographic highs and discharge in topographic lows in the more humid environments. While the identification of active recharge areas is important, it is the quantification of the recharge rates needed for the model that will always be difficult without data obtained via a field investigation. In the absence of relationships between recharge rates and basin properties, it should be possible to use annual or monthly recharge estimates against which the values of GW and GPOW could be calibrated. Unfortunately, the literature rarely contains information on annual variations or seasonal distributions of recharge, both of which could be very different for similar annual means.

3.5.3.3. Groundwater discharge to stream flow

The approach used to define the groundwater components of the model is a compromise between representing the real processes of sub-surface flow and using simple geometry to represent the aquifer. Each catchment is represented as a rectangle and the channels as parallel lines, separated by drainage slopes. The drainage slopes consist of the two areas between the edges of the rectangle and the outermost 'channels', plus two between each 'channel' line (Figure 3-12). Drainage is assumed to be one-dimensional for simplicity. The number, length and width of each drainage slope are determined from the catchment area and the effective *drainage density*. The assumption in the model is that the *drainage density* includes only those channels that are likely to receive groundwater discharge. Smaller tributary channels are often only flowing during storm events and therefore only receive surface runoff or interflow. While all of the groundwater parameters detailed below have direct physical meaning, the way in which they are quantified may depend upon the particular circumstances in any one region or basin. A clear understanding of the perceptual model in a region will help the parameterisation process in the absence of field data. Many of the parameters detailed below can be quantified using the GRA II database (Conrad, 2005), however, these values are fraught with uncertainty as they are based on scarce and uncertain data.

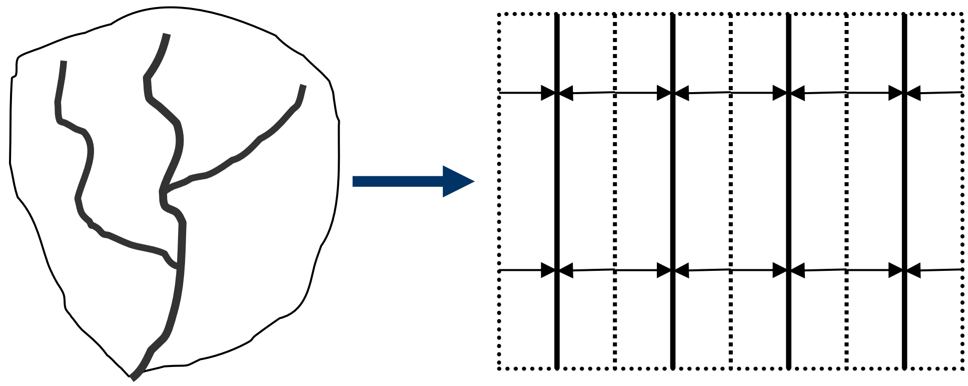


Figure 3-12 **Conceptual simplification of drainage in a basin for a drainage density of 4/SQRT(Area) (solid lines are channels, dashed lines are drainage divides and arrows show drainage directions). There are 8 drainage slopes (Hughes *et al*., 2007).**

The number of channel lines can be calculated from:

Total channel length = Drainage density * Area **Equation 7**

The ratio of catchment width/length is assumed to be related to *drainage density* as follows:

Width = Length * 2.0 * Drainage density **Equation 8**

Therefore:

Length = SQRT(Area / (2 * Drainage density)) **Equation 9**

By definition (and from Figure 3-12):

No. drainage slopes = 2.0 * Drainage density * Area / Length **Equation 10**

The number of drainage slopes is equal to 2 * number of channels, however Equation 10 has to be corrected to generate an even integer number of drainage slopes, each of which has a width given by:

Drainage width = Width / No. of drainage slopes **Equation 11**

Figure 3-13 illustrates the concept behind a single drainage slope and the volume of the groundwater 'wedge' stored under that drainage slope (assuming that the lower boundary is the channel at the bottom of the slope) can be calculated as:

'Wedge' volume = (Drainage width)^2 * Gradient * Drainage length / 2 **Equation 12**

Where 'gradient' is the hydraulic gradient of the groundwater flowing toward the river channel or away from the channel when the groundwater is below the channel.

Volume of water in 'wedge' = 'wedge' Volume * Storativity **Equation 13**

Outflows from this wedge to the river channel, within a single slope element can be calculated by:

Discharge = Transmissivity * Gradient * Time step * Channel length **Equation 14**

The groundwater slope or gradient is divided into two parts; the upper (or remote from the channel) and the lower (or near to the channel) (Figures 3-13 and 3-14). The near channel line segment is arbitrarily set at 40% of the slope element width and the remote segment as the remaining 60% with the gradient for each part calculated separately. The principles are that the water balance calculations are first performed on the lower slope component and the lower slope and position of the junction point fixed (the near channel groundwater slope is always attached to the channel). The water balance calculations are then performed on the upper slope element and the gradient of the upper slope fixed for the start of the next time interval. The recharge input and downstream catchment outflow (see later) are proportionally divided (i.e. 60:40) up for the two slope components.

To be able to quantify the groundwater discharge to stream flow, the following parameters need to be estimated, (1) *transmissivity*, (2) *storativity*, (3) *drainage density*, (4) *regional groundwater drainage slope* (see later), (5) *rest water level* (see later) and (4) the *riparian strip factor* (see later). *Transmissivity* (m^2 d^{-1}) is a product of the permeability and saturated aquifer thickness, while storativity (S) is a measure of the capacity of the aquifer to store water. Ideally the *storativity* and *transmissivity* parameters should be quantified on the basis of rock type and its degree of deformation or fracturing.

Both transmissivity and storativity are included in the national GRA II database (Conrad, 2005) but in certain circumstances need to be adjusted if the individual model user considers the database values to be incorrect or inappropriate for the specific study. The *drainage density* is expressed as a ratio of the total channel length expected to receive groundwater inputs to the basin area given in km km^2. *Drainage density* can be roughly inferred from maps and an approximate understanding of the basin characteristics. Lower *drainage densities* result in fewer slope elements and therefore lower rates of groundwater discharge per month. Lower values also result in smaller total sub-catchment outflow widths and therefore lower rates of downstream groundwater drainage. It is not always straightforward to determine the proportion of channels in a catchment that receive groundwater baseflow and the density of major channels in the catchment is usually taken as a rough estimate. In the model, the d*rainage density* parameter can be set at an initial value of 0.4 for most headwater catchments that do not have any specific shape characteristics. If they are elongated and the *transmissivity* parameter is high, it is appropriate to reduce the *drainage density* (to 0.3 or even 0.2) to ensure that outflow volumes to the downstream catchment are not excessive. Reducing the *drainage density* can also be used to smooth the variations in groundwater discharge to surface water (as can increasing the *storativity*). For downstream catchments, lower *drainage densities* appear to be appropriate (0.2 to 0.3). Note that *drainage densities* higher than about 0.5 should not be used unless there is extremely good justification.

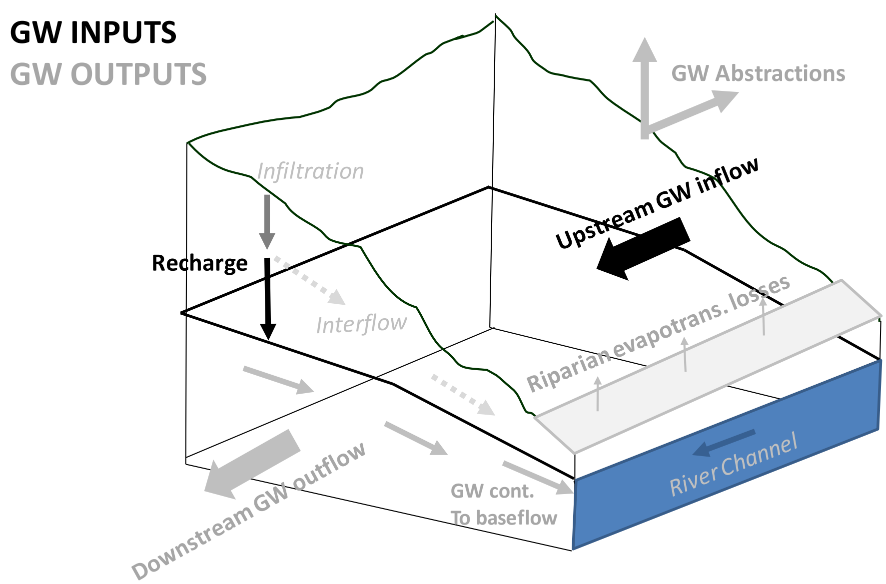


Figure 3-13 Diagram illustrating the main interaction components of the modified Pitman Model (scenario where groundwater is contributing to surface water). The 'wedge' represents the part of the groundwater body that is above the conceptual river channel and can contribute to discharge (adapted from Hughes *et al*., 2007).

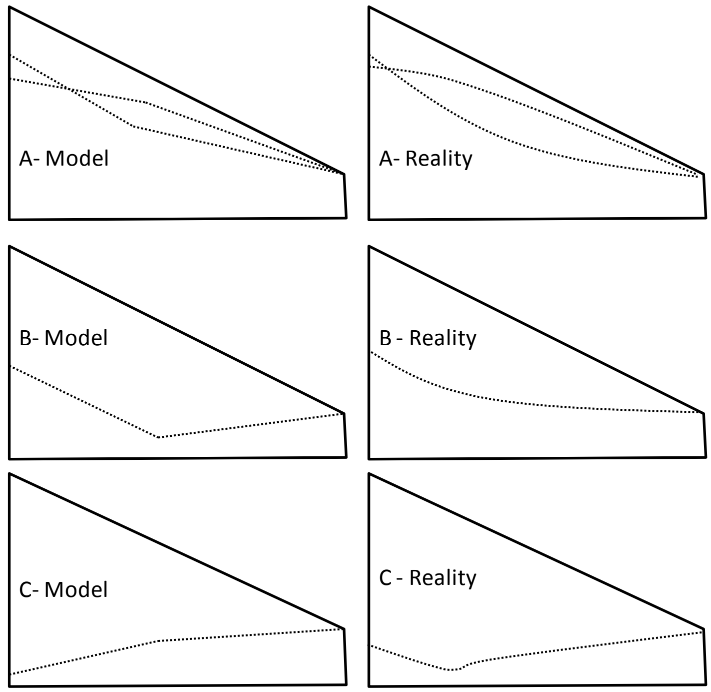


Figure 3-14 Modelled versus real groundwater conditions in a single hillslope element. Dashed lines represent the groundwater levels, the solid upper line represents the surface and the solid triangle represents the river channel (adapted from Hughes *et al*., 2010).

Figure 3-14 illustrates the conceptual representation of the two slope elements in the model. The right hand side of Figure 3-14 interprets the modelled groundwater slope geometry (left hand side) into the actual groundwater conditions that the model is attempting to conceptually represent. Figure 3-14A illustrates two examples where both line segments have positive gradients and groundwater will contribute to flows in the channel (unless riparian evapotranspiration losses are higher that the lateral contributions). Figure 3-14B illustrates a situation where the groundwater level is below the channel and no contributions to the channel are possible. This is reflected in the model as a negative near channel gradient. For this scenario, the connecting point between the two groundwater segments will be lower than the channel and this level is used to reduce both riparian evapotranspiration losses and downstream groundwater outflow. If this point reaches the *rest water level*, riparian losses and downstream groundwater outflow cease. Figure 3-14C illustrates a scenario where both gradients are negative. This situation could develop if groundwater abstractions occurred in the remote groundwater segment at a rate that exceeded the recharge inputs. In reality, the result would be an extensive cone of depression around a borehole field, at a distance away from the channel and no movement of groundwater toward the channel.

While the geometric representation of groundwater flow is lumped, very simplistic and ignores many of the realities of groundwater movement, it does incorporate the major groundwater principles and is useful as most of the calculations are simple geometric equations. While it may not be the hydraulic gradient that changes as groundwater contributions to surface flow vary (it could be contributing area or other factors), nevertheless changing the gradient has the desired effects:

- More recharge, increased gradients and more drainage to the channel in the future.
- If drainage is greater than recharge the outflow will gradually decline.
- Lower *drainage density*, less outflow.

3.5.3.4. Riparian losses to evapotranspiration

The effect of riparian vegetation on groundwater is well known. In many environments, groundwater can be subject to evapotranspiration losses close to the channel margin through transpiration by riparian vegetation and through evaporation from channel beds and banks. This is represented by a model parameter referred to as the r*iparian strip factor* (RSF) which represents the percentage of the total slope element width over which evapotranspiration losses are assumed to occur. This defines the volume of water loss through evaporation close to the channel margin, in the lower slope element and the losses are assumed to occur at the potential evaporation rate when the gradient is greater than zero. Any groundwater discharge to the channel is reduced by evapotranspiration losses. A further parameter (*rest water level - RWL*) refers to the maximum depth below the channel that an aquifer is assumed to reach and defines the depth of the connecting point between the upper and lower slope elements at which the groundwater is considered to be inaccessible to abstractions and evapotranspiration. This depth is translated into a gradient (necessarily negative) that can be used to estimate a depletion factor, when the current lower slope element gradient is less than zero:

GW depletion factor = (gradient at RWL – current gradient) / gradient at RWL **Equation 15**

Evapotranspiration losses are reduced by this depletion factor:

Evap. losses = Drainage Width * Net Evap. * Riparian strip factor * Depletion Factor **Equation 16**

Net evaporation refers to the difference between potential evaporation demand and rainfall. If the net evaporation is negative (i.e. where rainfall exceeds potential evaporation), the value is corrected to zero (to avoid duplicating the recharge function over the riparian strip).

The *riparian strip factor* parameter should reflect the areal extent and type of riparian vegetation which is likely to use near surface groundwater directly or intercept groundwater contributions to stream flow. Aerial photographs or google earth can be useful in delineating the average riparian strip width along a channel. The *rest water level* parameter can be taken from the existing database of groundwater information (using the variable 'median saturated thickness', Conrad, 2005) if no data on aquifer depths are available. This parameter is mostly relevant to semi-arid basins where the groundwater table is consistently below the channel bed. While it is not a very sensitive parameter in the model, it could impact on the extent to which large abstractions (relative to mean annual recharge) from groundwater can be maintained. However, extreme parameter values should be avoided (i.e. less than 10m and greater than about 50m) to avoid problems with the variation in the groundwater depletion factor calculation.

3.5.3.5. Channel transmission losses

Transmission losses are not straightforward to conceptualise and the process in the model assumes that the rate of loss will be due to characteristics of the channel, the head difference between the channel and the groundwater and the transmissivity of the material under the channel. If the near channel gradient is negative (the groundwater level drops below the level of the channel), losses will occur from the channel to the aquifer. In dry regions these transmission losses can be an important source of groundwater recharge, and therefore able to substantially alter the water balance in the model.

In downstream sub-catchments receiving inflows from an upstream sub-catchment, there are two components of channel loss calculated by the model. The first component is channel losses from the incremental runoff generated within a sub-catchment, while the second component is channel losses from flow in the main channel. Although they are treated separately in the model, the same algorithm is used.

The following scheme has been adopted for the channel losses to flow generated within the catchment (the incremental runoff). Three other variables are required which include MAXQ, TLQ and TLG. MAXQ is the maximum runoff (in mm) for the sub-basin being modelled and is estimated during the first run of the model. The variable is set to a default value of 20 mm at the start of the first run and a further variable TLQ estimated from the current months runoff (Q) and its value calculated using the following equations (see Figure 3-15):

If Q/MAXQ < 0.3

TLQ = 0.5 * (tanh(10 * (Q / MAXQ – 0.25)) + 1.0) **Equation 17**

If Q/MAXQ > 0.25

TLQ = 0.5 * (tanh(6 * (Temp – 0.625) + 1.0) **Equation 18**

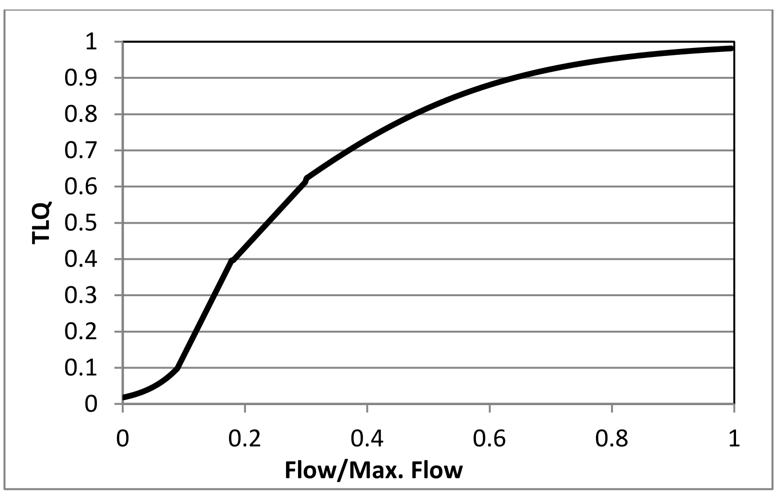


Figure 3-15 Shape of the power relationship between current month discharge (mm), relative to a maximum value (20mm in this case) and a model variable, TLQ (Hughes *et al.*, 2007).

A further variable (TLG) is estimated from the current gradient relative to a maximum gradient defined by 0.7 of the gradient at the '*Rest Water Level*'. It is therefore a measure of the head difference between the channel and the groundwater (i.e. groundwater gradient of the near channel slope element) and they are related to each other by a power function (Figure 3-16).

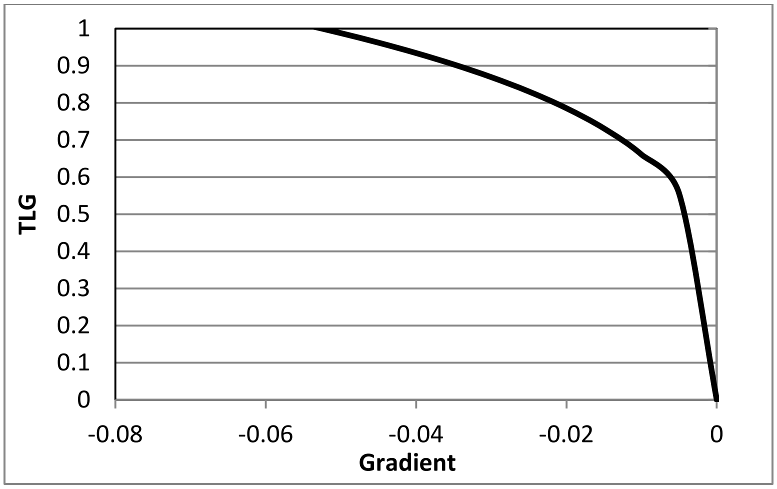


Figure 3-16 Shape of the power relationship between current downslope gradient and a model variable, TLG. The maximum value of TLG is defined by a model parameter (Hughes *et al.*, 2007).

TLG is estimated as follows: If Gradient < 0.7 * RWLGrad then TLG = 1, otherwise:

$$TLG = Gradient / (0.7 * RWLGrad)^{0.25} \qquad \textbf{Equation 19}$$

Channel loss (mm) is then the product of TLQ * TLG * TLGMax, which is removed from any available runoff and added to the lower slope component. The two exponents (0.4 and 0.25) have been fixed in the current version of the model to avoid introducing additional parameters that will be very difficult to quantify. The only additional parameter is therefore *TLGMax*, which refers to the maximum channel loss from the whole sub-catchment in mm month^{-1}. This maximum loss will occur when the lower slope gradient is lower than 70% of the gradient at the rest water level and when the sub-catchment runoff is at its maximum value.

As already noted the first channel loss routine only applies to incremental runoff generated within the sub-catchment of the distribution system and not to upstream runoff that passes through that sub-catchment (allogenic). To manage cumulative flow channel losses without adding additional parameters, the same functions as described above for sub-catchment channel losses have been used, but applied to the upstream inflow to the sub-catchment. The groundwater gradient component of the function remains the same (equations 18 and 19), except that *TLGMax* now represents a maximum channel loss from upstream inflow (in $m^3 * 10^6$). *TLGmax*_Inflow is calculated from the *TLGmax* parameter for incremental flow using the following scheme:

$$TLGmax\_Inflow = TLGMax * (MAXQ\_Inflow / MAXQ) \qquad \textbf{Equation 20}$$

Where MAXQ is defined previously as the maximum sub-area runoff (mm) and MAXQ_Inflow is the maximum upstream inflow. Both of these are set to initial values in the first run of the model (MAXQ = 20 mm, MAXQ_Inflow = 20 mm * cumulative upstream catchment area) and are then re-calculated for the second run from the data simulated during the first run.

Equations 17 and 18 are also used to estimate the TLQ component, but with MAXQ replaced by MAXQ_Inflow and Q defined as the upstream inflow in any one month. The cumulative inflow channel losses are estimated at the start of a single months simulation and reduce the upstream inflow (there is no iteration of this calculation). The additional volume is then added to the near channel (or lower element) groundwater storage in equal amounts over the model iteration steps (fixed at 4).

The TLGMax parameter represents the maximum possible channel loss and is used for both the loss routines. The estimation of this parameter value will never be simple, largely because of the highly non-linear nature of the channel loss process. The estimated channel losses in the model are also dependent upon the current months flow rate and near channel groundwater compartment gradient. It is important that this parameter is not ignored in dry regions where the groundwater lower slope element gradient will be nearly always negative. The only possibility for calibration would occur where nested gauged basins exist and where the downstream basin contributes little in terms of incremental runoff. If the TLGMax parameter is set too high relative to simulated runoff depths it is possible that a large part of the runoff generated from other model components could be lost to groundwater. The use of TLGMax for both loss functions might be considered problematic. However, where there are major losses from upstream runoff, there is likely to be very little incremental flow within the sub-catchment. The value of TLGMax will therefore be dominated by the range of values of upstream inflow, rather than local runoff.

3.5.3.6. Discharge to downstream catchments

A *regional groundwater gradient* parameter is included that refers to the gradient appropriate for estimating outflows from one sub-catchment to the next one downstream. While large scale groundwater flow can move laterally relative to a surface water sub-catchment, the model assumes that the groundwater flow follows the topography and therefore moves in a downstream direction. The same basic flow equation as Equation 15 is used:

Downstream outflow = Transmissivity * Regional gradient * Time step * slope width **Equation 21**

The total outflow for a sub-catchment would then be the result of equation 16 multiplied by the number of slope elements. Clearly the influence of the *drainage density* on the catchment width/length ratio will have a major impact on the volume of downstream outflow. The outflow is reduced by the groundwater depletion factor (Equation 15) when the lower slope element gradient is negative.

The *regional groundwater slope* does not seem to need to vary very much between catchments and provisional estimates suggest that a value of close to 0.01 will be satisfactory in most catchments. This parameter only affects the groundwater drainage to downstream sub-catchments which can be important in some sub-catchments. There is very little information available on this process at the scale of quaternary catchments and the *drainage density* parameter is likely to influence the volumes as much as any other parameter.

3.5.3.7. Groundwater abstractions

The response to abstractions will be different in the near channel areas (lower slope) to those that occur in areas remote from the channel (upper slope). Groundwater abstraction parameters (annual volume and seasonal distributions) can be applied to the two groundwater segments independently. There are therefore two water use parameters which represent the abstraction volumes in m^3 * 1000 from all the upper and lower slope elements. The seasonal distribution of groundwater abstractions (the same distribution is applied to both abstractions) is included as part of the monthly distribution data requirement. Simulated abstractions are not limited while the point joining the two line segments is above the *rest water level*, at which abstractions are assumed to cease. The model does not allow for water to be transferred from the near channel groundwater segment to the remote segment. Realistically, this is unlikely to happen as the groundwater gradients close to the channel would probably be very low.

3.5.3.8. Summary of the groundwater components of the model

The basic assumptions of the model are that the groundwater storage water balance is determined by inputs of recharge and outputs of flow to the river, riparian losses in the channel margins, drainage to a downstream sub-catchment and abstractions from boreholes. A further input of channel transmission loss is added if the groundwater level is below the channel and there is flow in the channel generated from the surface water components of the model. Therefore, through straightforward geometry calculations and the storativity parameter (based on aquifer porosity), the groundwater storage variations are translated into variations in the gradients of the two groundwater segments. The main reason for adopting this approach, rather than directly simulating a groundwater level, is the large spatial scale the model is typically applied at within which depth variations could be substantial. This approach avoids the complexity and incorporation of additional factors that would accompany a direct simulation of groundwater depth. The order of all the processes detailed above is given by Hughes *et al*. (2007) who outlined the progression within each model iteration step (4 per month) which is as follows:

- The recharge is calculated and the associated volume of water added to the upper and lower wedge storage volumes.
- The groundwater gradient from the previous iteration step is used to estimate outflow from the upper slope component to the lower slope component, the outflow from the lower slope component to the channel and the regional groundwater gradient (RG) is used to calculate the groundwater outflow to the downstream catchment. The riparian evapotranspiration losses are calculated, as are any channel transmission loss inputs to groundwater and any abstraction losses from groundwater.
- The new volumes of groundwater in the two slope elements are then used to estimate the slope gradients for the next time step.
- It is assumed that the lower slope end point is fixed at the river channel and the gradient calculated from 40% of the width and the volume.

- From the previous calculations, the upper slope end point, where it joins the lower slope element can be determined and therefore so can the gradient of the upper slope element from the upper slope volume and simple geometry.

3.5.4. Water use components of the Pitman model

The model simulates the influence of water resources developments on the natural stream flow. These can be grouped into those relating to reservoirs (both small farm dams and large dams) and those relating to water use (both surface water and groundwater). The parameters used for all the water use components are given in Table 3-2.

Table 3-2 The modified Pitman model water use and wetland sub-model parameters.

Parameters	Units	Description
Surface water use parameters		
AIRR	km^2	Irrigation area
IWR	fraction	Irrigation water return flow fraction
EFFECT	fraction	Effective rainfall fraction
RUSE	Ml/year	Non-irrigation demand from the river
Groundwater use parameters		
GWA (Upper slopes)	Ml $year^{-1}$	Groundwater abstraction far from the channel
GWA (Lower slopes)	Ml $year^{-1}$	Groundwater abstraction near to the channel
Small farm dam parameters		
MDAM	Ml	Small dam storage capacity
DAREA	%	% of sub-basin above dams
A		Parameter in non-linear dam area-volume relationship
B		As above
IRRIG	km^2	Irrigation area from small dams
Large reservoir sub-model parameters		
Reservoir Capacity	MCM	Reservoir storage capacity
Dead Storage	% Capacity	Dead storage of the reservoir
Initial Storage	% Capacity	Reservoir magnitude at the beginning of the simulation period
A in $Area(m^2) = A^* Volume(m^3)^B$		

B in Area(m^2) = A * Volume(m^3)^B		
Reserve level 1-5	% Capacity	5 levels of operating rules used to reduce abstraction of reduced storage.
Annual Abstraction	MCM	Demand from the reservoir
Annual Compensation Flow	MCM	Downstream compensation flow released into the river
Wetland and lake sub-model parameters		
MaxWA	Km2	Maximum wetland area.
RWV	$m^3 * 10^6$	Residual wetland storage volume below which there are no return flows to the river channel.
IWV	$m^3 * 10^6$	Initial wetland storage volume at the start of the simulation.
AVC	m^{-1}	Constant in the WA = AVC * WV^{AVP} relationship, where WA (m^2) and WV (m^3) are the current wetland area (limited to MaxWA) and volume, respectively.
AVP		Power in the WA = AVC * WV^{AVP} relationship.
QCap	$m^3 * 10^6$	Channel capacity below which there is no spill from the channel to the wetland.
QSF		Channel spill factor in SPILL = QSF * (Q – QCAP), where Q is the upstream flow and SPILL is the volume added to wetland storage.
RFC		Return flow constant in the RFF = RFC * (WV / RWV)^{RFP} relationship. RFF is a fraction limited to a maximum of 0.95 and then adjusted when Q is greater than QCap (RFF = RFF * QCap / Q). The return flow volume is calculated from RFLOW = RFF * (WV – RWV).
RFP		Return flow power in the RFF = RFC * (WV / RWV)^{RFP} relationship.
EVAP	mm	Annual evaporation from the wetland (distributed into monthly values using a table of calendar month percentages).
ABS	$m^3 * 10^6$	Annual water abstractions from the wetland (distributed into monthly values using a table of calendar month percentages).

The surface water use accounts for direct river abstractions for agricultural, domestic and industrial purposes. The model has routines for differentiating direct abstractions from the river for irrigation and non-irrigation purposes. The irrigated area satisfied by run of river abstractions is controlled by parameter AIRR and a model attribute which describes the monthly distribution of irrigation depth (in mm) required. The effective rainfall parameter (EFFECT) reduces the irrigation depth requirement by this proportion of the rainfall occurring within a month. RUSE refers to the annual volume of non-irrigation demand and is used together with a model attribute that fixes the monthly distribution of demand. The model attribute is made up of 12 rows (months) and 4 columns (Monthly Distribution Weights, Monthly Irrigation Demand (mm), Monthly Water Demand (fraction) and Groundwater Demand (fraction)). The first column of data is used to distribute seasonally different parameter values for all the months, the second to determine the depth of monthly irrigation water demand, the third to distribute the annual non-irrigation water use value and the fourth to distribute the annual groundwater abstractions. The groundwater use parameters are discussed in more detail in section 3.5.3.7.

The farm dam component of the model has five parameters which account for their impact in a catchment. DAREA (%) relates to the proportion of a sub-catchment contributing runoff to the small dams while MDAM refers to the capacity of the small dams' storage. The runoff generated in this area is assumed to initially satisfy the available reservoir storage before being able to contribute to flow at the outlet of the sub-catchment. Water in the dams is subject to loss through abstraction for irrigation purposes (parameter IRRIG) and a model attribute of monthly distribution of depth of irrigation water demand, and evaporation which is controlled by a non-linear relationship between area and volume and the monthly potential evaporation demand.

Large dams are simulated using a reservoir sub-model of the Pitman model. Inflows to the reservoir include flow generated within the sub-catchment and from all upstream sub-catchments. The reservoir parameters with brief descriptions are given in Table 3-2, while the seasonal distribution of abstraction and compensation flow (for all five operating rule levels) comprise a further model input.

A wetland sub-model was developed during the course of this project as an optional extra in parallel with large dams (Hughes *et al*., 2013). The model was developed to improve simulations where the impacts of wetlands or natural lakes on stream flow, which are found in many river basins in southern Africa, are evident. Processes associated with wetlands and lakes can exert a considerable influence on downstream flow regimes through attenuation, storage and slow release processes that occur within the water bodies. These processes are critical in understanding the general patterns of runoff generation at the basin scale. Table 3-2 lists the input parameters for the wetland model, most of which are physically based. The approach is based on relatively flexible functions that account for the input-output relationships between the river channel and the wetland or lake. Incorporation of the wetland sub-model will attenuate stream flows which can change the baseflow response. As both the wetland sub-model and the surface and groundwater interaction components of the model will affect low flows, it is essential to ensure there is no conflict between the two components.

3.5.5. Uncertainties in the representation of processes

It is accepted that the way in which the groundwater components have been conceptualised within the model will not be appropriate for all situations. Current uncertainties associated with the revised model include: (1) Potential structural uncertainties associated with the spatial scales used; (2) uncertainties associated with the temporal scales used; (3) uncertainty associated with the groundwater parameters utilised and (4) processes which are not represented in the model, such as regional groundwater flows. These uncertainties can affect the outputs of the model in different environments in diverse ways. Therefore, it is important to identify in which environments the model is applicable and in which it is not, as well as to isolate specific uncertainties that any model user should be aware of (Table 3-3). It is important to note that the model is not designed to replace detailed groundwater investigations, either based on field observations or through the use of specialised groundwater models but was developed to complement detailed surface and groundwater models by providing broad catchment scale estimates of the main interaction variables (recharge and groundwater contribution to stream flow). Some of the more specific uncertainties are outlined below.

Table 3-3 Dominant processes in 'typical' interaction environments found in South Africa and the ability of the Pitman model to represent them.

Process	Simulated
Fractured rock aquifers	
Perched aquifers and interflow	Yes
Direct recharge into bare rock areas	Not explicitly
Preferential recharge in highly fractured areas	Not explicitly
Preferential flows and variable aquifer characteristics associated with regolith aquifers	Not explicitly
Evapotranspiration directly from groundwater in riparian margin	Yes
Discrete groundwater discharge into surface water systems	No

Regional groundwater flows	No
Preferential flow through fault systems and dykes and sills	No
High spatial variability	Yes
Confined aquifers	Not explicitly
Karst aquifers	
Different surface and groundwater catchment divides	Not explicitly
The presence of dykes and sills which can compartmentalise the karst aquifer	Not explicitly
High concentrated recharge	Yes
High concentrated discharge into surface water systems	Yes
Groundwater lag	No
Perched aquifers and interflow	Yes
Extremely high transmissivity and storativity	Yes
Alluvial aquifers	
Connections with underlying aquifers which can recharge the alluvium or be recharged by the alluvium	Not explicitly
Transmission losses, both channel losses and flood plain losses	Yes
Diffuse discharge into surface water systems	Yes
Different infiltration rates dependent on characteristics of flood events (duration and intensity)	Yes
Hysteretic inundation effects	Yes
Hysteretic recharge effects	No
Bank storage	Not explicitly
High evapotranspiration from lowland flood plains and riparian margins	Yes
High transmissivity and storativity	Yes
Primary aquifers	
Low probability of unsaturated zone flow	Yes
Relatively diffuse recharge	Yes
Shallow groundwater tables	Yes
High evapotranspiration losses directly from shallow groundwater and the riparian margin	Yes
Diffuse discharge into surface water systems	Yes
The presence of lakes, wetlands and estuaries	Yes
Perched aquifers, confined and semi-confined aquifers on a small scale	Not explicitly
Connections with underlying aquifers	Not explicitly
High transmissivity and storativity	Yes

Direct recharge (unrelated to soil moisture storage) into bare rock areas is not accounted for in the model and can be important in fractured rock settings particularly in high relief areas. The assumption is that direct recharge would occur in areas with thin soils. This scenario is simulated by the model and thought to sufficiently replicate direct recharge in bare rock areas. High rainfall in an area with thin soils would lead to rapid saturation of the soil moisture store inducing high groundwater recharge, increased groundwater gradients and subsequent increased discharge to the channel. Making use of a high value for GPOW in Equation 6 would simulate these conditions on substantial non-linearity in the relationship between recharge and soil moisture storage.

There is no groundwater or recharge lag routine, as the groundwater function acts as a routing reservoir. Studies show groundwater discharge hydrographs respond very quickly to rainfall events, and therefore the assumption is that the displacement effect negates the need for a lag routine. The groundwater displacement effect assumes that when the aquifer is recharged, it will discharge fairly quickly, however, the discharge water will be older groundwater, unrelated to the new water that has just recharged the aquifer. However, in some environments a more substantial groundwater lag can be important and it is uncertain whether the groundwater reservoir is sufficient as a routing reservoir. This is especially true in karst environments which can have non-linear responses to inputs and are often reliant on thresholds.

The effect of large scale groundwater abstraction on groundwater discharge to stream flow is a process that is not yet well understood. The ability of the model to acceptably simulate this process is therefore questionable as there is no way to clearly validate the simulations. Figure 3-14 illustrates the perceived conceptual process that the model attempts to recreate using both the remote and near channel gradients.

While the assumption is that the majority of groundwater discharge to surface water in South Africa is from unconfined aquifers, it is accepted that in some settings confined aquifers play an important role. While the Pitman model does not explicitly simulate confined conditions, it can represent them implicitly. The model assumes that the only way a confined aquifer contributes to groundwater baseflow is via fractures in the aquitard which would connect it to the overlying unconfined aquifer. The volume of the contributions to groundwater baseflow would depend on the hydraulic head of the confined aquifer which would need to be higher than the river channel. This scenario would be simulated in the model by reducing the transmissivity which would considerably limit groundwater discharge and cause a buildup of the groundwater gradient in effect replicating the hydraulic head and permeability controls on confined aquifers. The uncertainty in the model consists of interpreting the final discharge pathway of the aquifer.

The Pitman model formulation does not currently account for situations where surface and groundwater catchment divides do not coincide. A model component that allows for the movement of groundwater out of the system would be relatively simple to incorporate in the model at the expense of additional parameters. This feature has not been developed as yet largely due to the lack of conceptual and quantitative information on this process in South Africa.

The use of the Pitman model was preferred for the following reasons, including the requirement for an integrated model that can support regional water resource assessments, the ability to parameterise the model in data scarce areas (using the regional surface and groundwater databases as well as physical data from generally available maps), and the presence of uncertainty functions within the model. The model is suited for use in semi-arid and fractured rock environments which cover the majority of South Africa, and can represent the various sources of low flow in rivers. However, the limitations and uncertainties in both the model and the conceptual hypothesis need to be recognised to prevent the potential problem of a false belief in model results. To a large extent it is apparent that the structure should be able to handle most South African environments. However, there clearly remain some real-world processes that are either absent or are not represented explicitly. Whether these structural uncertainties can be differentiated from the other dominant issues associated with the form of the model algorithms, uncertain input data and quantifying appropriate and representative values for the parameters remains to be seen. This thesis focuses on the improvement of the conceptual understanding within a catchment, analysing the model's ability to represent this conceptual understanding (structural uncertainty), then assessing the outputs from the model based upon the processes simulated to achieve those outputs (parameter equifinality uncertainty). While the Pitman model is well suited to modelling in South African environments due to its extensive experience in the region, it is important to maintain awareness of the model assumptions when using the model and in the assessment of modelled catchments and their outputs.

3.5.6. Methodology

The methodology followed for each of the following case studies can be summarised as follows:

1. Collection of all qualitative and quantitative data.
2. Formulation of conceptual model.
3. Critically analyse available data and current conceptual understanding.
4. Set up model to represent conceptual understanding.
5. Calibrate model while ensuring that the parameter values remain sensible.
6. Evaluate model outputs based on:
   - Ability to represent conceptual understanding sufficiently.
   - Comparisons with alternative model outputs.
   - Comparisons with observed data.

4. RESULTS

There are seven example catchments outlined in this chapter, which represent a range of environment 'types' found in South Africa. A conceptualisation of the dominant processes found in each catchment is provided before the model setup and results are detailed. While the study sites often explore a number of issues, they are all focused on one main topic of interest. Unsaturated zone processes are explored in both the Sabie River and Grahamstown study sites, while the Gamagara River explores processes associated with an alluvial environment. The Molopo dolomitic eye explores a karst environment and the Upper Breede River, Buffelsjag River and Elands River are all within fractured rock environments.

4.1. Unsaturated zone processes – the Sabie River

4.1.1. Introduction

While the Grahamstown site examined a setting dominated exclusively by soil moisture and unsaturated zone processes, the Sabie catchment represents a much larger area with contributions to the low flows expected from both interflow and groundwater. It is not straightforward to determine the relative contributions to the low flows and in this scenario some simple observations of the observed flow coupled with knowledge of the geological and topographical setting enabled a perceptual model to be formulated. While the sub-catchment is gauged there are no other data available with which to confirm the detailed outputs of the model. However, there are some signals within the observed stream flow data that can be used to assess the likely contribution of interflow and groundwater.

4.1.2. Description of the study area

The Sabie River drains the Eastern Escarpment of South Africa within Mpumalanga Province (Figure 4-1). It consists of one sub-catchment designated X31A (Figure 4-2). The area experiences summer rainfall with high summer temperatures. The mean annual precipitation is between 1200 and 1300 mm y^{-1}, while mean annual potential evaporation is around 1300 to 1400 mm y^{-1} (Midgley *et al*., 1994). Previous investigations carried out include King and Louw (1998), Moon *et al*. (1997) and Rivers-Moore and Jewitt (2004). The long-term mean annual recharge of this area has been estimated to be 8 to 9 % of mean annual rainfall, or a depth of approximately 110 mm (DWAF, 2005a). It is an area of steep topography (elevation range of 308 m) dominantly underlain by quartzite's, but with a band of dolomite traversing the catchment from north to south (Figure 4-3). The catchment is gauged in the upstream reaches at X3H001 with a catchment area of 173 km^2.

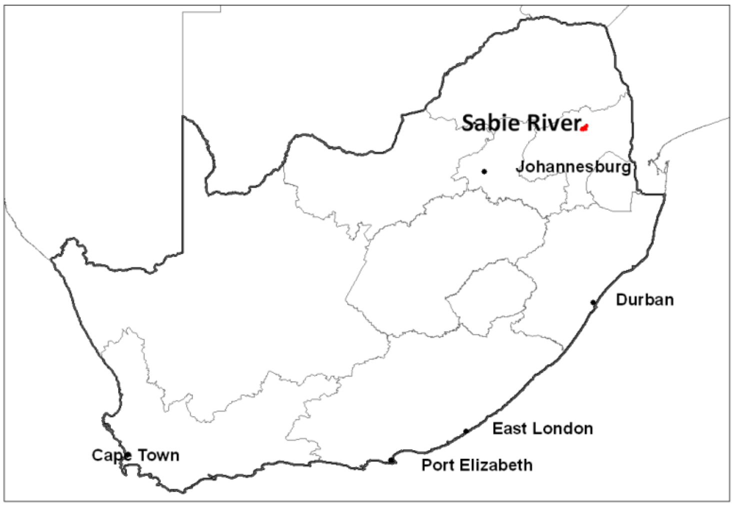


Figure 4-1 Location of the sub-catchment X31A in South Africa.

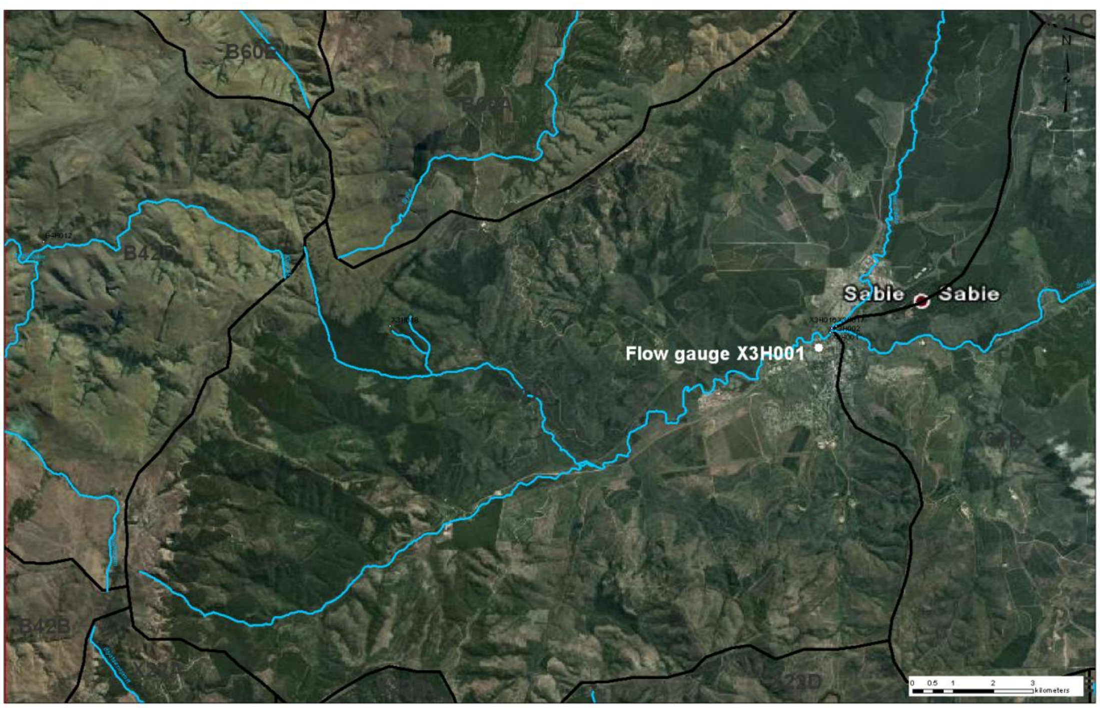


Figure 4-2 Sub-catchment X31A showing the river network and location of the flow gauge.

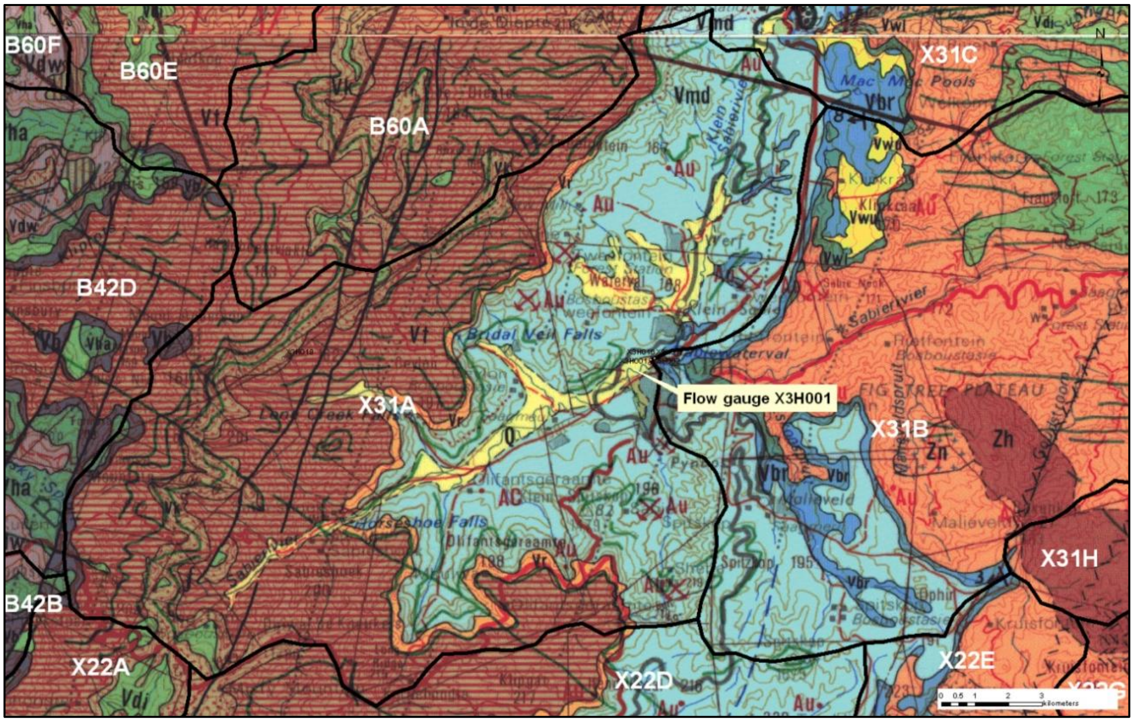


Figure 4-3 Sub-catchment D41A showing the geology of the region. Underlying the dolomite is quartzite shown in dark blue (east), the band of dolomite is represented by the light blue layer, overlying quartzite is shown in red (west) and the recent quaternary sand and alluvium is shown in yellow (Geological survey, 1986).

4.1.3. Conceptualisation of the interaction processes

The conceptualisation of the processes occurring in the catchment are based on a visual interpretation of the daily stream flow time series as well the application of a digital filtering baseflow separation technique (Hughes *et al*., 2003). From analyses of the observed flow hydrograph, it is apparent that during dry years the minimum baseflows are maintained (although at lower levels), while the seasonal 'baseflow' response is largely absent. This suggests a relatively continuous drainage system with an additional baseflow response superimposed. The latter responds rapidly to seasonal variations in rainfall and is therefore poorly buffered. From these observations the baseflow, or low flow component of stream flow, is conceptualised to be made up of a groundwater component which is fairly stable and fluctuates around the estimated long term mean annual recharge of 110 mm. In addition, there seems to be an interflow component of the baseflow which responds to seasonal rainfall totals more rapidly and is likely to originate from lateral flow in the macro-pores of the dolomite unsaturated zone. This component is expected to be much more variable and dependent upon annual fluctuations in rainfall. Two dry and one wet periods were selected to attempt to approximately quantify the groundwater discharge to stream flow and interflow contributions. Figure 4-4 shows monthly rainfall and mean daily flow time series for the three 5-year periods. The first period (October 1973–September 1978) is generally wet with a mean rainfall of 1468 mm y^{-1}, the second covers a drier period with mean rainfall of 980 mm y^{-1}, while the third has a mean rainfall of 1037 mm y^{-1}.

The volumes of baseflow or low flow during both the wet and dry period are given in Table 4-1, based on commonly used hydrograph separation procedures (Nathan and McMahon, 1992; Hughes *et al.*, 2003). During the dry periods, the baseflow component is still apparent but the volume of baseflow or low flow is lower. Removing the last two years of the dry period, which are similar to the wet period hydrograph, provides even lower baseflow figures. Additionally, the daily hydrograph indicates a very quick response with baseflow responding fast to rainfall and decreasing in the first year of lower rainfall (i.e. 1981), therefore there is no delay. A similar pattern was observed during the second drought sequence in the 1990s.

Table 4-1 Proportion of baseflow or low flow in both wet and dry periods.

Water balance component	Wet period (1973-1978)	Dry period (1981-1986)	Shortened dry period (1981-1984)
Rainfall	1470 mm	980 mm	1037 mm
Baseflow % of rainfall	377 mm (26%)	137 mm (14%)	120 mm (12%)
Total flow	585 mm	292 mm	239 mm
Baseflow % of total flow	380 mm (65%)	143 mm (49%)	112 mm (47%)

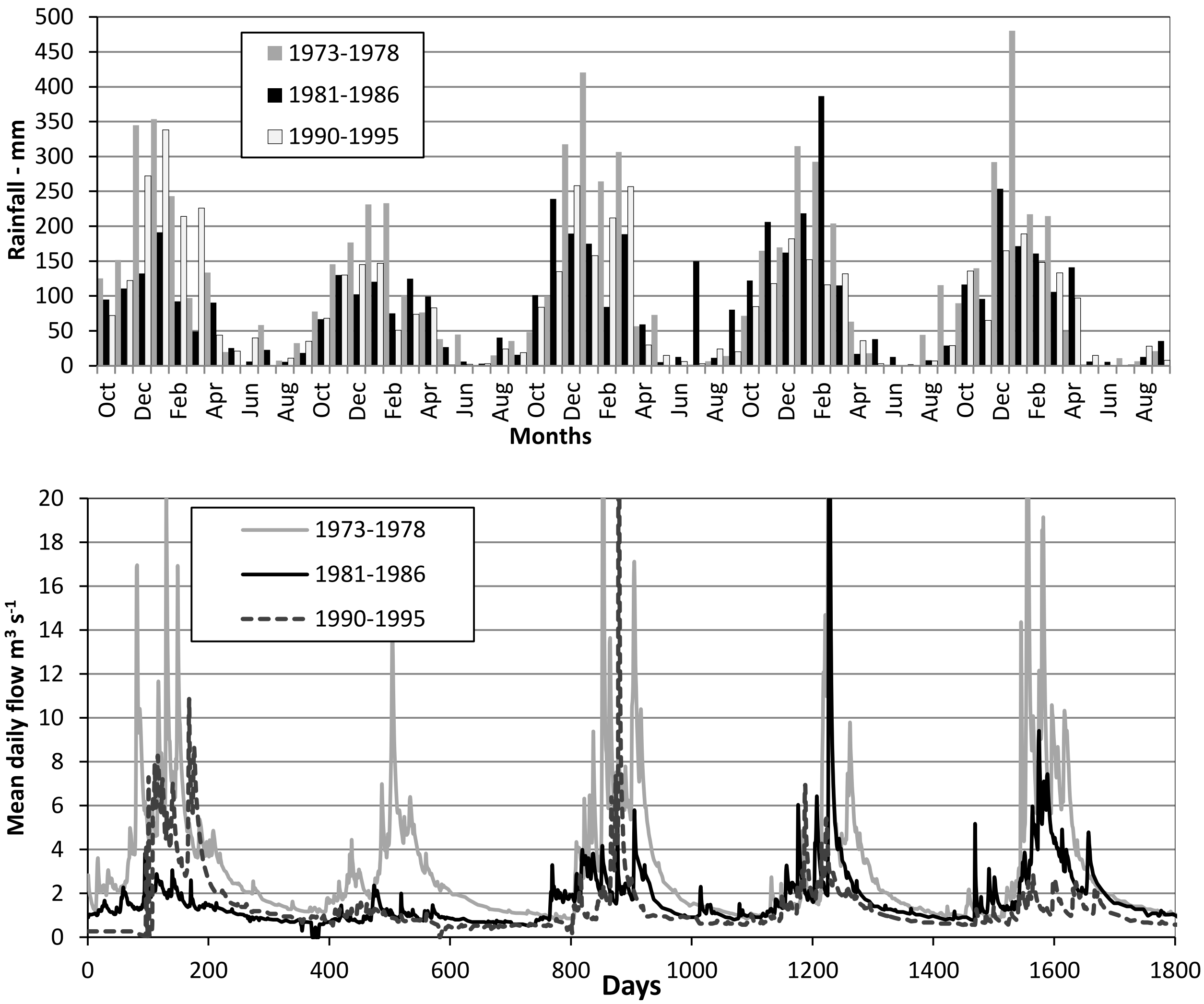


Figure 4-4 **Rainfall and mean daily stream flow for the Sabie River (gauging station X3H001) for three 5-year periods (01 October 1973 to 30 September 1978, 01 October 1981 to 30 September 1986 and 01 October 1990 to 30 September 1995).**

4.1.4. Model setup and results

The Pitman model was established with the regional rainfall and evaporation demand data available in WR90 (Midgley *et al*., 1994), and was calibrated against the observed data. The model parameters are given in Table 4-2; the GW parameter was calibrated to match the published long term recharge (DWAF, 2005a) and the unsaturated zone parameters set relatively high to represent the volume of interflow. Table 4-2 also includes a summary of the simulated water balance components during dry and wet periods. The resulting simulation is illustrated in Figure 4-5 together with the observed flows, while the objective function results are given in Table 4-2. Figure 4-6 illustrates the three simulated runoff components which constitute the stream flow during the wet period and the first dry period.

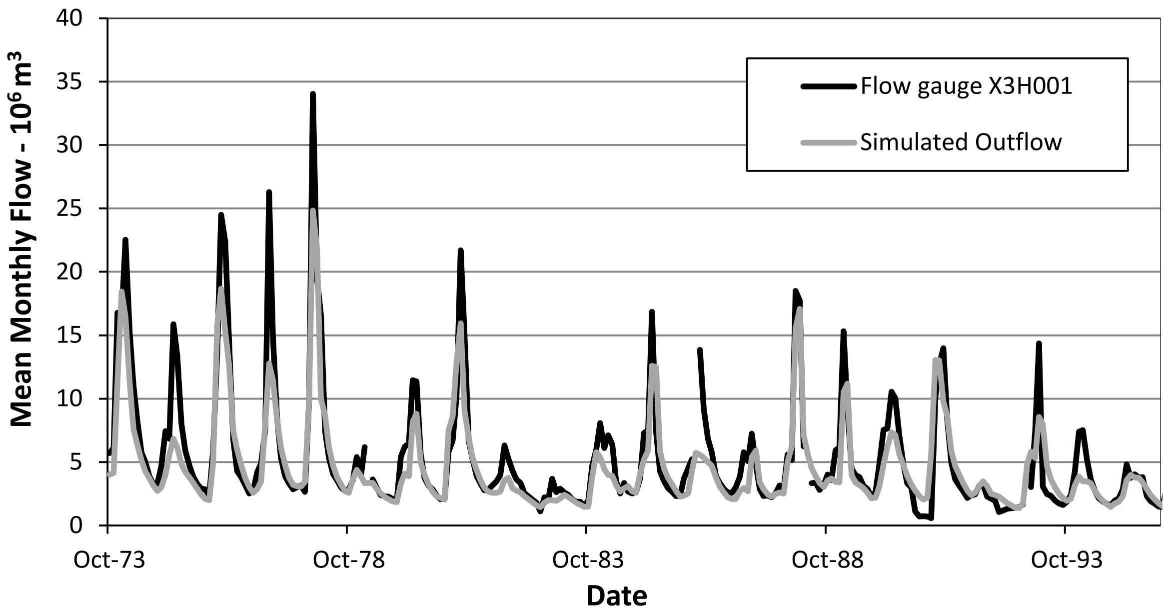


Figure 4-5 **Comparison of observed monthly stream flow for the Sabie River (gauging station X3H001) and simulated stream flow (Pitman Model).**

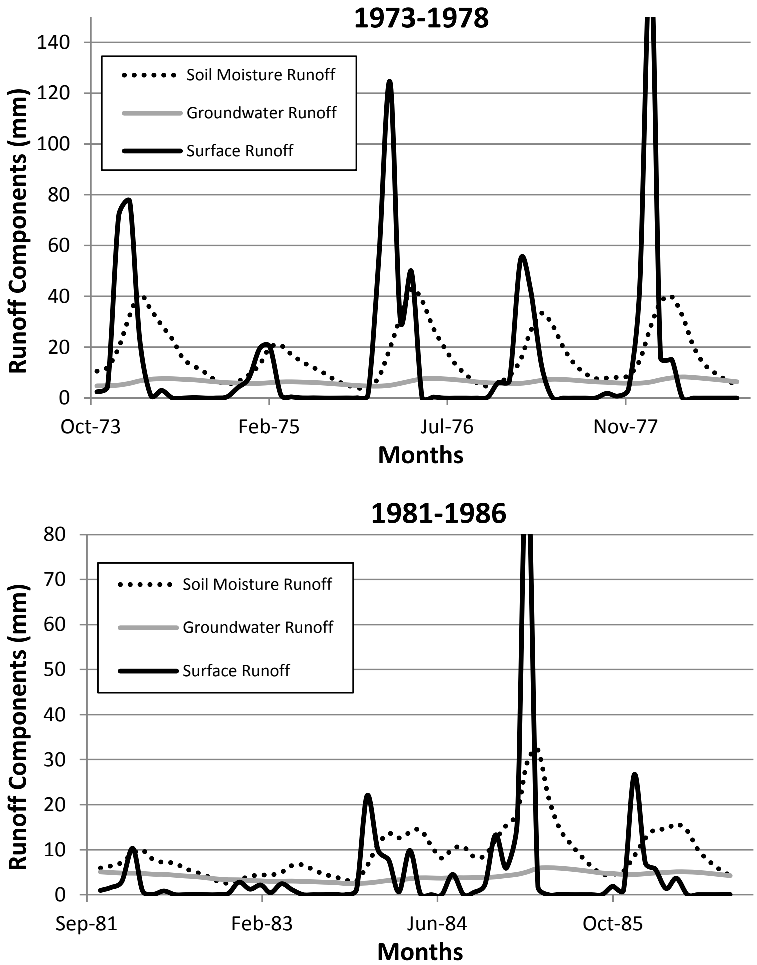


Figure 4-6 **Runoff components (mm) for the Sabie River for two 5-year periods, a wet period (01 October 1973 to 30 September 1978) and a dry period (01 October 1981 to 30 September 1986).**

Table 4-2 Parameter values, approximate water balance and model statistics for the Sabie River example during wet and dry periods.

Parameter values	Entire period			
ST	1000			
FT	125			
POW	3			
GW	70			
GPOW	3			
Drainage density	0.4			
Transmissivity (m^2 d^{-1})	9.4			
Storativity	0.004			
ZMIN (mm m^{-1})	0			
ZMAX (mm m^{-1})	800			
Riparian Strip Factor (% slope width)	0.2			
Water balance component	Wet period (1973-1978)	Dry period (1981-1986)	Dry period (1990-1995)	
Rainfall (mm y^{-1})	1470 (100%)	980 (100%)	1037 (100%)	
Surface runoff (mm y^{-1})	197 (13%)	116 (12%)	135 (13%)	
Unsaturated zone macro-pore flow (mm y^{-1})	263 (18%)	30 (3%)	42 (4%)	
Groundwater flow (mm y^{-1})	128 (9%)	109 (11%)	103 (10%)	
Evaporative losses (mm y^{-1})	882 (60%)	725 (74%)	757 (73%)	
Model performance	Wet period (1973-1978)	Dry period (1981-1986)	Dry period (1990-1995)	Entire period (1948-2005)
Nash Coefficient (Untransformed)	0.78	0.57	0.67	0.79
Nash Coefficient (Ln transformed)	0.86	0.67	0.52	0.74
Nash Coefficient (Inverse transformed)	0.83	0.70	0.15	0.11
% Bias in simulated monthly flows	-11.37	-11.49	4.80	-5.58
% Bias in simulated monthly ln(flows)	-3.56	-8.74	14.75	-1.71

While the objective functions of the model performance are better for the wet period than for the dry periods, the percentage bias in the simulated flows for all the high and lows flows remain within 15% of the observed data. Figure 4-7 compares the observed and simulated flow duration curves and illustrates that the model successfully simulates the stream flow volumes throughout the range of flows.

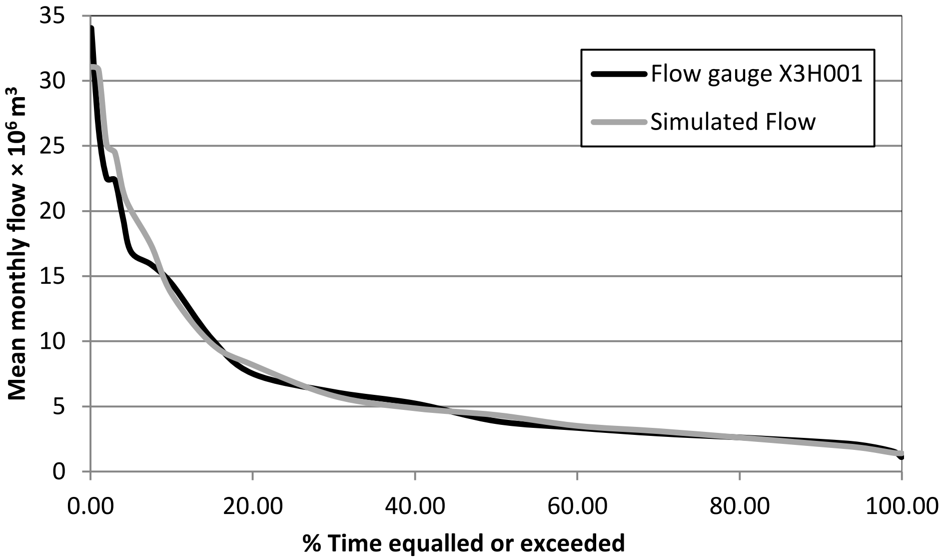


Figure 4-7 Comparison of observed monthly stream flow (10^6 m^3) for the Sabie River (gauging station X3H001) and simulated stream flow (Pitman Model).

A further analysis was carried out using the uncertainty version of the model to:

1. Assess whether the model achieves better results if the baseflows are simulated with either interflow or groundwater.
2. Assess whether the calibration of the model is affected by the time period of observed data used.

The uncertainty version of the model was run using three sets of data (Table 4-3), a wet period, a dry period and the complete record. The 'best' ensembles were selected based on the objective functions the coefficient of efficiency for the normal (CE N), natural logarithm transformed (CE L) and inverse values (CE 1/d) and the mean monthly runoff error for both normal (% Diff N) and natural logarithm transformed values (% Diff L). The remaining ensembles were constrained using the recharge values from the GRA II database (DWAF, 2005a). The results of the analysis are shown in Table 4-3 and Figure 4-8 below.

Table 4-3 Minimum and maximum parameter values of the constrained ensembles.

		Uncertain parameters				Objective functions	
Years	Parameter	ST	FT	GW	RSF	CE N	CE L
1973 - 1978	Minimum	649	190	47	0.36	0.845	0.885
	Maximum	786	199	52	0.639	0.856	0.893
1981 - 1995	Minimum	766	115	40	0.352	0.727	0.625
	Maximum	1111	141	45	0.799	0.749	0.661
1966 - 2005	Minimum	908	137	42	0.312	0.795	0.762
	Maximum	1147	160	48	0.75	0.808	0.78

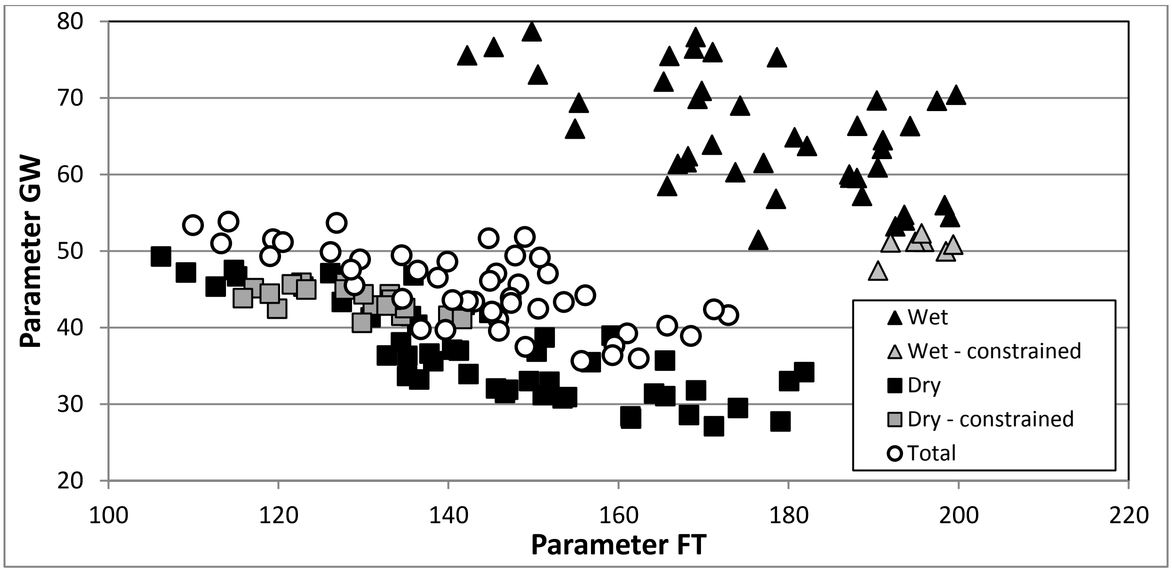


Figure 4-8 The 'best' ensembles (according to objective functions) both unconstrained (shown in black) and constrained (shown in grey) using available recharge values. The circles represent the unconstrained values from the simulation using the total period of observed data.

The results of the analysis indicated that the model could not clearly determine whether the baseflows were comprised predominantly of groundwater or interflow contributions using the observed data alone. The conceptual understanding is thus important to ensure the model is correctly set up. In addition, the parameters are sensitive to the calibration period used, highlighting the importance of ensuring, as far as possible, that the observed data used in calibration is representative of a range of conditions.

4.1.5. Conclusion

This example sub-catchment highlights the importance of representing unsaturated zone flow processes in any integrated model developed for South African conditions. The reliable simulation of such a non-stationary response in the hydrograph requires three runoff components in a model in order to generate a conceptually realistic representation of the runoff response. A large volume of unsaturated zone flow is only likely to be important in catchments that have steep enough topography for the lateral flow vector to intersect the ground surface. The evidence in this setting includes conceptual interpretation of stream flow records combined with some knowledge and perceived understanding of the catchment physical properties. While it is possible that there are other interpretations of the patterns of observed flow, the catchment conditions certainly seem to support the hypothesis. Until more detailed field measurements can be obtained, the interpretations will always be open to question. Quantifying the characteristics of fractured or dolomitic unsaturated zones, in terms of the available fracture storage, transmissivity rates and the relative importance of the vertical and lateral flow components will always be difficult and represent a limitation to the success of any studies. Chemical or isotopic tracer based studies should be able to resolve questions about the sources of the components of stream flow. However there have been very few isotope tracer analyses undertaken for South African rivers. This type of interaction environment emphasises the care that must be taken in inferring sub-surface processes from an interpretation of stream flow data such as baseflow separation methods, particularly if it is important to get the right results for the right reason.

4.2. Unsaturated zone processes: Grahamstown tributary

4.2.1. Introduction

The Grahamstown study site was selected as a small experimental catchment to attempt to assess the low flow components of the Pitman model. As data from long-term experimental catchments are not generally available in developing countries such as South Africa, an alternative is to initiate targeted short-term investigations to resolve specific uncertainties in either the structure of hydrological models or the way in which their parameters are quantified. It was planned to use some short-term simple periodic stream flow observations in a small grassland catchment to assess the validity of the low flow algorithms of the Pitman Model. The Grahamstown study site was selected as interflow processes clearly play a large role in the generation of low flows in the catchment. One of the objectives of the study was to identify if there are differences in low flow response between the very small headwater catchment and the lower catchment and between wet and dry periods. It was initially considered possible that during dry conditions the flows in the lower part of the catchment might be derived from groundwater seepage, however the results of the study detailed below do not support this concept. Given the small scale of the catchment and the short period of observations (< 2 years), the Pitman model algorithms were applied in a daily time step version of the model. This work forms part of a study which aimed to assess the model structure and performance using more detailed (in space and time) data than would be available under typical water resource assessment applications of the model (Hughes *et al*., 2013).

4.2.2. Description of the study area

The Grahamstown site is located in the Eastern Cape Province (Figure 4-9) and consists of a first order, steep, grassland catchment underlain by quartzites of the Witteberg Group. Grahamstown is located in a summer rainfall area, with two dominant rainfall seasons (September, October and November, and February, March, April). However, the area is located between the winter rainfall region of the Western Cape and the country's remaining summer rainfall region, which results in high rainfall variability and in reality it can rain at any time of the year. Winter rainfall is typically due to cold fronts moving eastwards, with mean annual precipitation at between 600 and 700 mm. Mean annual evaporation is estimated at between 1500 and 1600 mm (WR90 – Midgley *et al*., 1994). There are several perennial and seasonal springs that have been identified within the area of Grahamstown and there is evidence to suggest that these are fracture flow springs fed by interflow through the quartzite high relief ridges (Hughes, 2010a). They are normally found at topographically lower positions than the catchments used in this study and therefore there is some uncertainty about the source of low flows in the small headwater catchments draining the quartzite ridges. The small catchment (0.23 km^2) has a single incised river channel (Figure 4-10) with slopes varying from about 17% in the headwaters to 40% at the point where the channel enters the incised gulley. The majority of the catchment slopes are approximately 18 to 22%, while the channel has a slope of 13%. The vegetation cover consists of quite dense grassland that has not experienced burning during the study period.

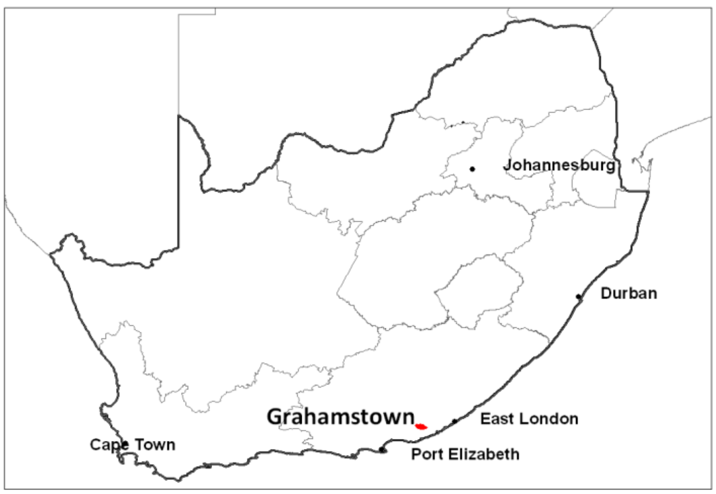


Figure 4-9 Location of the Grahamstown site in South Africa.

Figure 4-10 Aerial image of the Grahamstown site showing the topography and soil depths.

4.2.3. Conceptualisation of the interaction processes

To obtain a clearer understanding of the volume of storage available in the soil moisture store, a series of representative soil depth measurements were obtained from different parts of the catchment. A metal probe was hammered into the ground and the depth to the rock interface measured. In areas where the soil depth was deeper than the length of the probe (1.2 m), an auger was used (the auger could reach depths of 5 m). The measurements were taken within a series of cross sections over the catchment to get the best representation possible. Soils vary in depth from very shallow near some rock outcrops, through shallow (< 300 mm) on the main slopes and up to 700 mm in the flatter headwater areas. In the valley bottom, surrounding the incised channel, the colluvial deposits can be over 3 m deep and there is plenty of evidence of preferential sub-surface pathways (pipes). These appear to occur at the interface of the sandy loam (with a high organic content) top soils and deeper clay soils of the valley bottom colluvium or saprolite and hard rock on the slopes. While the depths were reasonably consistent, anomalies were measured which were attributed to soft weathered rock which the probe was able to penetrate at weak points. The information on the average soil depths was used to estimate the available soil moisture storage. In addition, groundwater levels were measured in May 2012 from three boreholes located near the study site (Figure 4-11). A fourth borehole drilled during 2010 at the top of the hill near the head of the catchment failed to strike water and reached 150 m before the drilling was abandoned. The borehole information suggests that the groundwater table is likely to be well below the catchment outlet. Similarly, isotope samples collected for both the upper and lower sites during a wet and dry period (Figure 4-12), support the concept that the low flow response of the very small headwater catchment and the lower catchment are similar. Both sites show similar degrees of enrichment, with the dry season samples showing greater evaporative enrichment, consistent with slower drainage through the soil profile. The groundwater storage and drainage components were therefore not included in the model setup, and the model assessments focused on unsaturated zone water balance components.

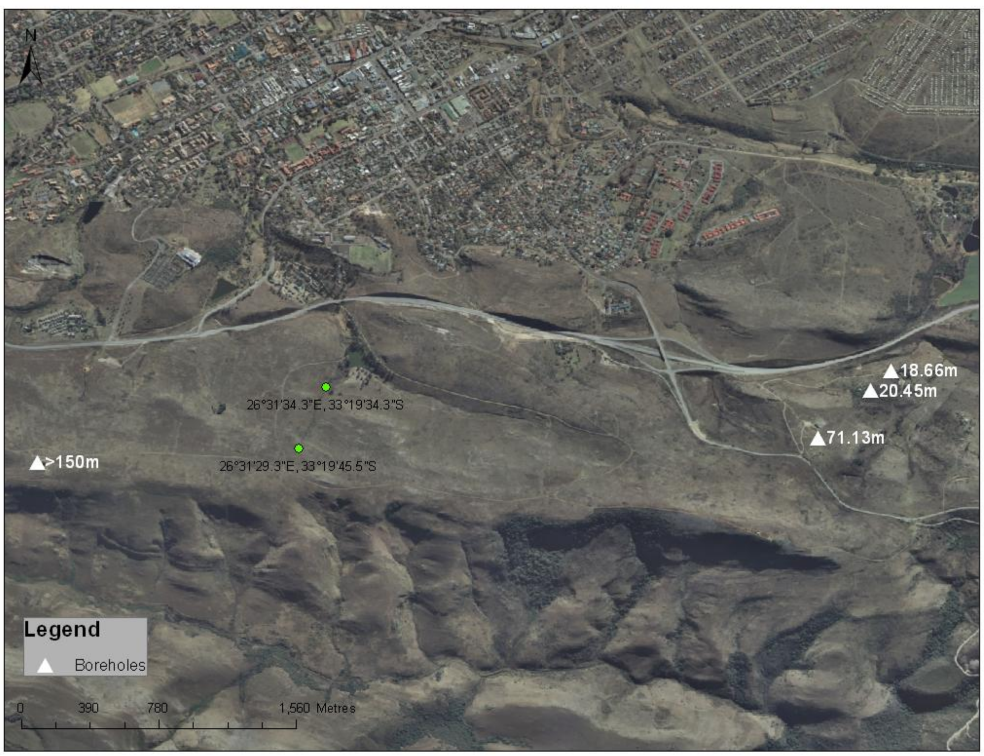


Figure 4-11 Grahamstown catchment showing location of boreholes with groundwater level depths.

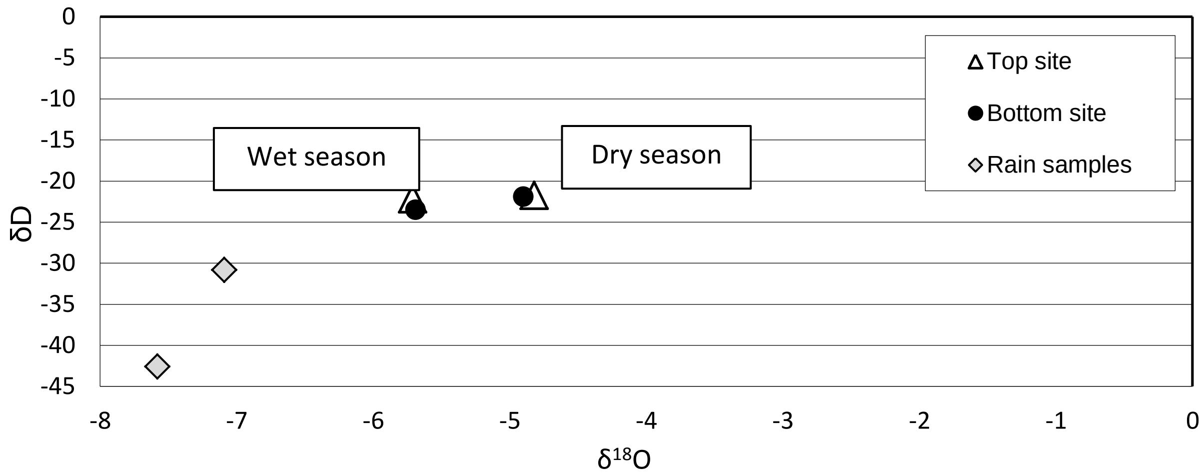


Figure 4-12 Plot of δD versus δ^{18}O for isotope samples taken from runoff at the upper and lower (outlet) sites during a dry and wet season.

4.2.4. Model setup and results

In this study the monthly water balance algorithms of the Pitman model have been applied in a daily version of the model so that the outputs can be more directly compared to the field data collected. Exactly the same algorithms that are used in the monthly model can be applied in the daily version, although the parameter values typically applied to the monthly model are not appropriate at a daily time scale. In this case, the FT and POW parameter values, which represent the monthly variation in soil moisture, were much higher and therefore more variable than would be expected at the monthly time step. While it is clear that the typically applied parameter values need to be adjusted in the daily time step version of the model, it is still uncertain exactly how different the revised parameter values need to be. Further implementation of the daily version together with the monthly version would enable new parameter relationships to be established. Stream flow measurements at the outlet (catchment area of 0.2 km^2) and at a point close to the start of the channel incision (catchment area of 0.03 km^2) were started in January 2011. A simple measurement system was used based on the time taken to fill a bucket of a known volume (repeated 5 times for each measurement). There are 60 observations available up to August 2012 with an approximate weekly interval, together with several more intense sampling periods immediately after heavy rainfall events. Daily weather station data are available from the Rhodes University campus located 2 km from the site and include estimates of ET_0 as part of the data logger outputs (Rhodes University weather station, 2012). Daily total rainfall observations were sourced from a site approximately 1.2 km away, but it should be recognised that rainfall can be quite spatially variable in this hilly terrain. The evaporation was simulated using an annual potential evaporation value (1552 mm) with a seasonal distribution. Two alternative seasonal distributions were used. The first was a fixed monthly seasonal distribution based on regional values taken from WR90 (Midgley *et al*., 1994) and this results in fixed daily values within any calendar month. In the second case the same annual value was distributed using the daily ET_0 data calculated from the weather station. The main differences in the two seasonal distributions of evaporation demand are during the winter months, with the ET_0 based estimates suggesting a much more evenly distributed demand than the regional values from the WR90 database.

The daily model was manually calibrated against the observed flow; however there was a degree of equifinality in the parameters due to some uncertainty in assigning parameter values. The following calibration guidelines were followed to ensure that the parameter sets were constrained to appropriate values as far as possible:

- ZMIN and ZMAX values were set to generate surface runoff only during rainfalls of greater than 50 mm day^{-1} and have very little influence on the simulations.
- Interception storage was set to 0.6 mm resulting in mean losses of approximately 11% of rainfall, which is considered to be consistent with the dense and relatively tall grass cover.
- GW, GPOW and SL_{GR} values were established to ensure that the mean recharge lay within expected values (DWAF, 2005a) of approximately 4% of rainfall.
- ST values were constrained to be within values expected from the knowledge of soil depths and texture (150 to 300 mm).
- FT values were calibrated to ensure a good fit to the maximum observed discharge values under high moisture storage conditions.
- POW, SL_{QI} and R values were calibrated against the observed recessions and the runoff observations under relatively dry conditions.

The parameter space was explored in the following sequence:

- ST was varied over the range 150 to 300 and the GW parameter adjusted to get a mean recharge of 4% of rainfall (GPOW and SL_{GR} values were fixed at 2.5 and 0).

- For each ST, R was varied from 0 upwards and the best-fit values for FT, POW and SL_{QI} were identified.
- While several objective functions were calculated, the Nash Coefficient based on untransformed (CE) and natural logarithmic transformed data (CE{ln}) were found to be the best measure of performance across the full range of flows and these were used together with visual assessments of the total time series graphs to guide the calibration process.

While the calibration of the simulated outputs compared reasonably well with the observed data (Figure 4-13), there were many different parameter sets that generated similar solutions based on the combinations of the two objective functions. However, no parameter set was able to achieve a balance between the relatively dry periods during September and November (days 280 to 300 and 320 to 350) with those during the mid-winter months of May to July (days 540 to 600). The observed flows during the recession within the former periods are mostly under-simulated, while those during the winter period are over-simulated. It was not found to be possible to derive an alternative parameter set that would retain the generally good simulations as well as fitting to the observed flows at the end of the record. During the manual calibration process, this effect is evidenced by a deterioration of the CE{ln} values as the CE values improve, and vice-versa. Uncertainties within the climate input data were thus examined to attempt to improve the simulation.

It is possible that some of the periods of poor model fit are related to inadequate representation of the site-specific rainfall inputs, but no additional information is available to check this. Uncertainties within the evapotranspiration data were examined by comparing the simulations using both available seasonal distributions. Replacing the regional evaporation demand data with a daily distribution based on the weather station ET_0 estimates improved the visual fit (Figure 4-13B), but many of the observed flows at the end of the record were still over-simulated. Tables 4-4 and 4-5 provide the parameter values and water balance components for one possible parameter set (used to generate the Figure 4-13B simulated flows), but there are other sets which give similar objective function values and similar visual fits.

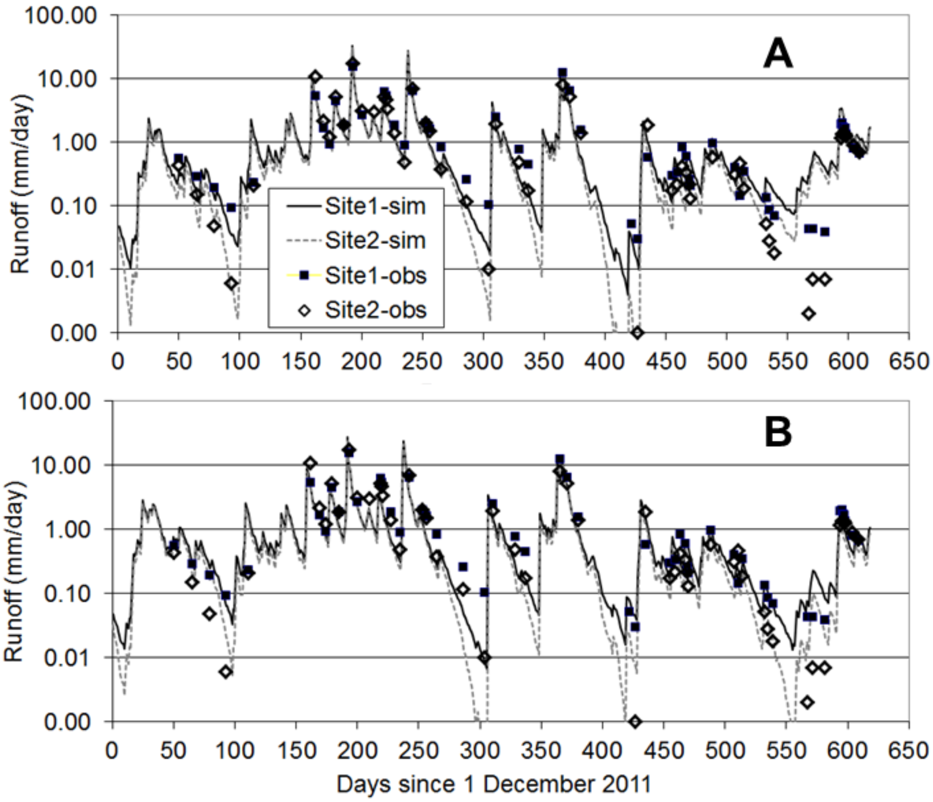


Figure 4-13 Simulations of stream flow for the Grahamstown sites (Site 1 = total catchment; Site 2 = lower catchment). A is based on fixed monthly regional potential evaporation (PE) estimates, while B is based on the regional mean annual PE distributed using ET_0 observations from the weather station.

Table 4-4 **Pitman model parameter values used in the Grahamstown site simulations.**

Parameter	Upper (site 2) catchment	Total (site 1) catchment
ST (mm)	220	220
Evap. Parameter R	0.0	0.0
FT (mm d^{-1})	60.0	70.0
POW	3.5	3.5
SL_{QI} (mm)	35	10
GW (mm d^{-1})	1.2	1.9
GPOW	2.5	2.5
SL_{GR} (mm)	0.0	0.0
CE{ln}	0.690	0.732
CE	0.801	0.817

Table 4-5 **Grahamstown water balance results for the simulations based on the annual potential evaporation distributed using daily ET_0 estimates (all values are annualised and rounded to nearest mm).**

Component	Pitman Model simulations	
	Upper catchment	Total catchment
Potential or reference evap. (ET_0 mm)	1 576	1 576
Rainfall (mm)	985	985
Interception loss (mm)	112	112
Soil evapotranspiration (mm)	479	397
Total actual evap. (mm)	591	509
Surface runoff (mm)	7	7
Interflow runoff (mm)	338	421
GW recharge (mm)	39	40
Change in soil storage (mm)	10	8

The main problem seems to be associated with insufficient moisture losses during the dry second winter. As this catchment is north facing it was postulated that the winter solar radiation input would be enhanced. Scott Munro and Huang (1997) investigated the differences in evaporative loss between north and south facing slopes in a Chinese catchment close to the Tropic of Cancer. Their results indicated that south facing slopes have approximately 40% higher evaporative losses compared to north facing slopes during winter, with no differences in summer. The effects for the Grahamstown catchment could be even greater, given its higher latitude. If the weather station potential evaporation estimates are increased relative to the summer values (by a factor of 20% for June and July, 10% for May and August and 5% for April and September), the visual fits to the period between days 520 ad 580 and the CE{ln} values are slightly improved but the CE objective function values are decreased. Increasing the potential evaporation estimates to greater than 20% starts to affect the other components of the model and results in poor correlation with the observed flows.

The parameter values for the maximum soil store are rather higher than those obtained from the field survey of soil depths but it is very difficult to estimate a representative value given the large variations between the slopes and the incised channel area. High maximum soil moisture runoff parameter values (60 to 70 mm day^{-1}) lead to rapid reductions in simulated runoff as the soil moisture content reduces, consistent with field observations of rapid sub-surface pipe flow during wet conditions throughout the length of the incised channel banks.

4.2.5. Conclusion

This investigation highlights the importance of getting the other components of the water balance correct, if a reliable assessment of the unsaturated zone algorithms is to be undertaken. The models inability to simulate both a wet and a dry winter satisfactorily could be due to issues with the evapotranspiration component of the model and not due to the unsaturated zone drainage components. Without more detailed information, it has not been possible to conclusively attribute these effects to any specific component of the model, a weakness in any specific model algorithm or a lack of representative climate inputs. It is also difficult to conclude whether these effects have identified a possible weakness in the overall model concepts, or whether they might be specific to the daily version that has been applied in this study. Improvements in the simulations were obtained by using daily estimates of ET_0 based on weather station data and the Penman-Monteith estimation equation. In addition, there is the possibility than the sub-catchment experiences increased evapotranspiration (relative to the estimates provided from the weather station) during winter due to its north facing position, as the weather station is situated in a part of Grahamstown not influenced by slope aspect. Increasing the winter potential evapotranspiration estimates relative to the summer values did improve the visual fit to the observed data although affected other parts of the model negatively when increased too much. A further possibility is that the rainfall data used are not adequately representative of the rainfall at the study site. The area experiences quite high spatial variability in rainfall due to the hilly terrain and there could have been differences in total daily rainfall depths between the measurement site and the catchment. Hughes *et al*. (2013) incorporated an uncertainty assessment into the simulation which included the generation of 1000 flow ensembles of the Grahamstown site. This work further confirmed the models inability to satisfactorily simulate both the wet and dry periods.

The results have quite clearly demonstrated that the standard monthly distributions of evaporation demand that have been published nationally (WR90 - Midgley *et al*., 1994) are not appropriate, at least at the small scale of the catchment studied. It is possible that this issue is related to catchment scale and that the published values are appropriate for much larger catchments. Although much of the data that are available to quantify the expected ranges of some of the water balance components (e.g. interception and groundwater recharge) are somewhat uncertain, the simulations appear to be behavioural, despite the fact that there remain some unresolved uncertainties about the operation of the model under dry winter conditions. This suggests that the main water balance algorithms of the Pitman Model are generally acceptable even for small catchments, given appropriate input data. The study highlights the importance of covering a range of wet and dry conditions in short term studies before conclusions are made about behavioural parameter sets and the most appropriate input data. The overall conclusion of the study is that even the simple type of field data collected for the Grahamstown site has been valuable in calibrating and assessing the structure of a daily implementation of the Pitman Model.

4.3. Alluvial environment – Gamagara River

4.3.1. Introduction

This environment consists of a large alluvial aquifer with a strong connection to the underlying aquifer which has been extensively dewatered as part of an open cast mining operation. The only available data are a series of groundwater levels which have been used to assess the model by comparing them to the simulated groundwater slope. In many parts of South Africa, mine dewatering has lowered groundwater levels by hundreds of metres. These environments are fairly complex to characterise given the lack of information generally available on the volumes and rates of dewatering.

A reliable simulation of the response of an alluvial aquifer is not straightforward as the aquifers are often only present within small parts of a basin (although have a large effect on the interaction processes) and are parameterised quite differently to the surrounding catchment. The parameters of conceptual type models are typically designed to represent catchment scale processes, but there remain questions about spatial scale effects on the estimation of parameter values, especially in these types of environments.

4.3.2. Description of the study area

The study site is located in the Northern Cape province of South Africa, adjacent to the town of Kathu (Figure 4-14). An iron ore mine is situated on a portion of the catchment (D41J), which forms part of the Molopo Drainage region in the Lower Vaal Water Management Area. The region is semi-arid and experiences hot summer months and cold dry winters. The MAP is 400 to 600 mm y^{-1} and the mean annual potential evaporation is 2200 to 2600 mm y^{-1} (Midgley *et al.*,1994). Rainfall data collected by the iron ore mine indicates a mean annual rainfall for the period between 1963 and 2009 of 379 mm y^{-1}, however, rainfall in this region is highly spatially variable. The topography is characterised by a relatively flat plain and a gentle gradient with an elevation range of around 100 m.

The flow in the Gamagara River is ephemeral responding to high rainfall events over the catchment but without sustained baseflows. However, there are no stream flow gauges on the river that can provide quantitative information about historical patterns of stream flow. Any knowledge of the frequency of stream flow events is therefore based on anecdotal evidence from local landowners (Bredenkamp and De Jager, 2011). There are two main tributaries in the upper reaches of the Gamagara River viz. the Daniëlskuil branch (1451 km^2) in the east and the Postmasburg branch (356 km^2) in the south (Figure 4-15). Even under natural conditions transmission losses are expected to be high with stream flow generated in the headwaters being lost to the alluvial material along the length of the main channel. The alluvial aquifer is also assumed to be hydraulically connected to the regional aquifer within the underlying fractured rock and karst (Bredenkamp and De Jager, 2011).

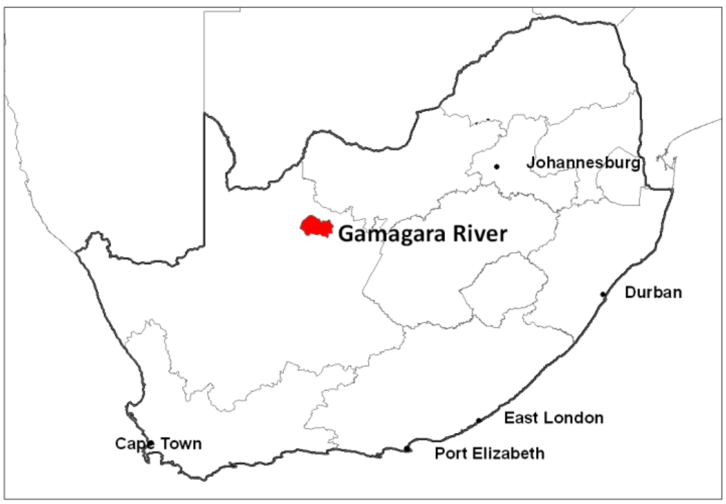


Figure 4-14 Location of the Gamagara River in South Africa.

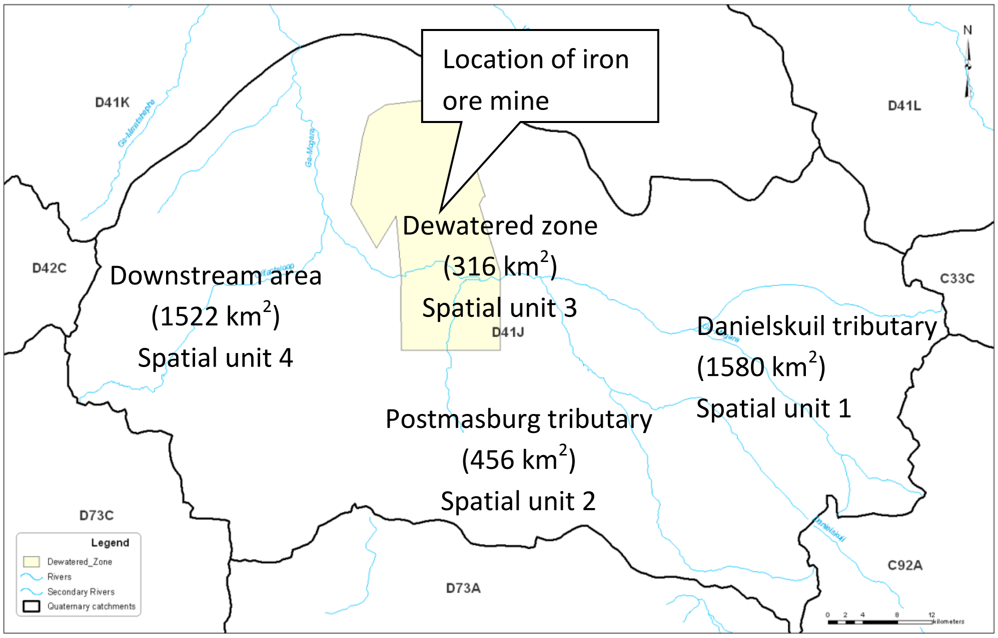


Figure 4-15 Gamagara River with dewatered zone (shaded).

The geology of the area has been well documented in studies such as Friese (2007) and Meyer (2009). The bedrock geology of the Postmasburg section (spatial unit 2) is dominated by dolomite, chert and dolomitic limestone. The Danielskuil section (spatial unit 1) originates on jasperlite, banded iron stone and volcanic rocks but flows mainly over Kalahari Group sediments. The bedrock geology of the dewatered section is dominated by surface limestone (recent) and quartzite, while the downstream section flows mainly over Kalahari windblown sands (Figure 4-16).

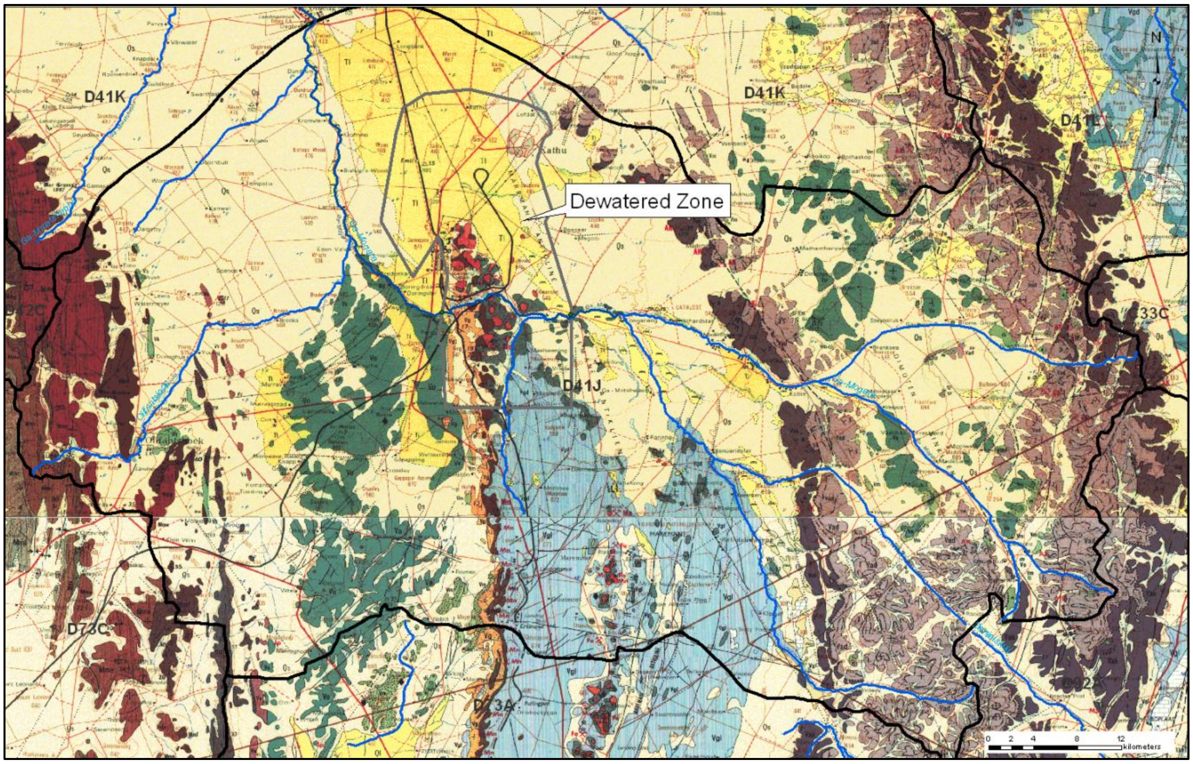


Figure 4-16 **Sub-catchment D41J showing the geology of the region. Dolomite is shown in blue over the south of the sub-catchment, overlain by jasperlite shown in brown over the east of the sub-catchment. The remainder of the catchment is predominantly wind-blown Kalahari Sands (light yellow) with alluvium shown along the river course (Geological survey, 1979).**

4.3.3. Conceptualisation of the interaction processes

The main channel valley bottom is filled with up to 10 m of fine quartzitic sand underlain by a competent and low permeability calcretised pebble layer. The recharge to the alluvial aquifer is expected to occur partly from channel transmission losses during flow events as well as from the surrounding aquifers that will be recharged from high rainfall events over the catchment as a whole. The recharge mechanisms are therefore expected to be complex and highly variable over time. VSA (2008) concluded that monthly rainfalls below about 80-100 mm do not recharge the regional aquifer significantly, while Bredenkamp and De Jager (2011) estimated that recharge to the alluvium could be between 5 and 25% of MAP or even higher. The GRA II database estimates recharge over the entire catchment at between 0.4 and 5% of MAP (DWAF, 2005a).

Dewatering of the aquifer (for open cast mining) started in 1967 and pumping volumes increased from 1970 to 1976 reaching a maximum during the period 1977 and 1981 (25×10^6 m^3 y^{-1}). The volume of groundwater abstracted has varied each year between 10 and 15×10^6 m^3 y^{-1} since. Dewatering started impacting on groundwater levels between 1974 and 1980. The mine currently abstracts approximately 13×10^6 m^3 y^{-1}. The comparisons between the observed and simulated groundwater levels within the dewatered zone are somewhat confused by the lack of detailed information on the history of pumping rates. The dewatered portion of the aquifer (5 to 6 km wide and 30 km long) is enclosed by a series of dolerite dykes which have effectively compartmentalised the area and hydraulically disconnected it from the remainder of the aquifer in catchment D41J. Figure 4-17 illustrates observed groundwater levels from a series of boreholes located within and just outside the boundary of the dewatered zone. The positions of the boreholes used in this study are shown in Figure 4-18.

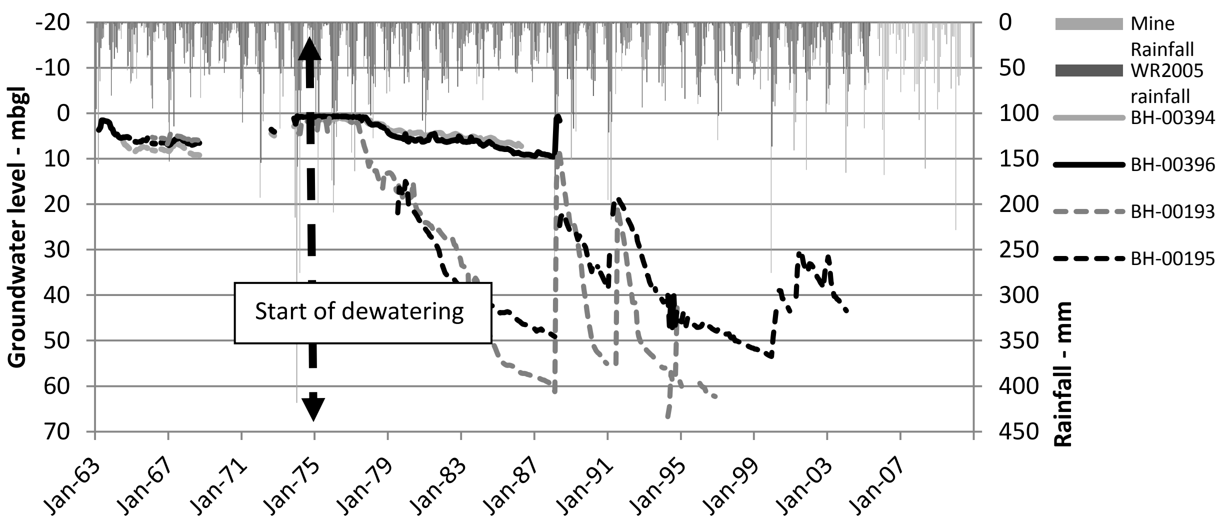


Figure 4-17 **Observed groundwater levels in and near the dewatered zone represented as metres below ground level (mbgl) for the entire period of observed data available, with both WR2005 and the mine rainfall included.**

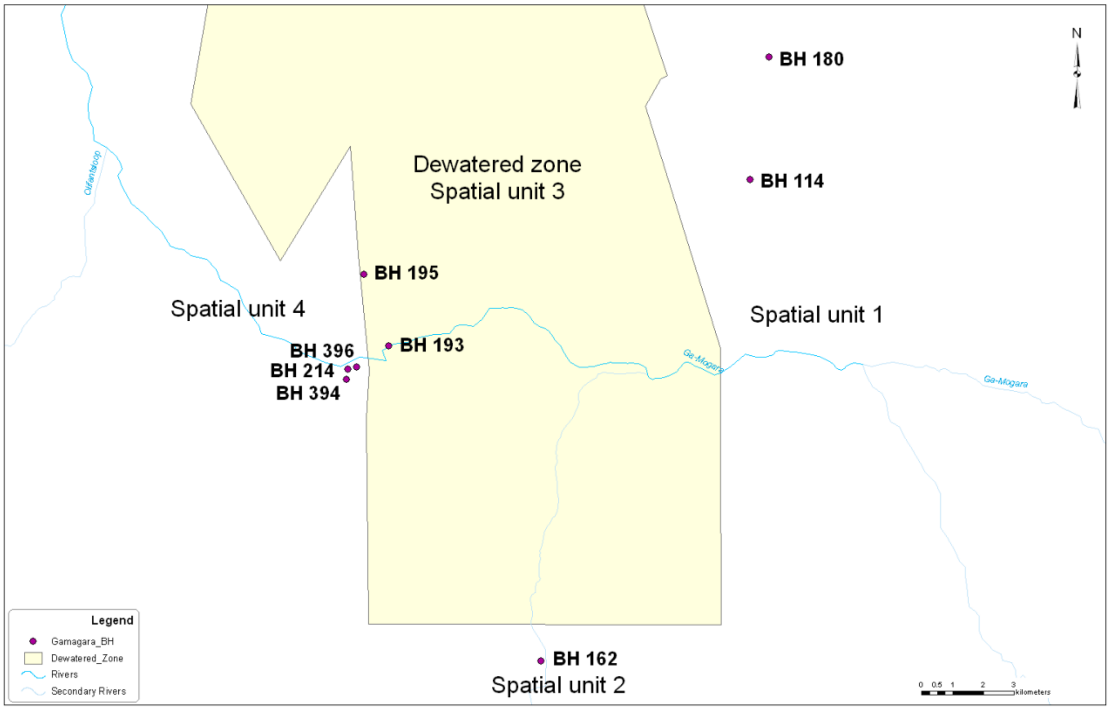


Figure 4-18 Gamagara River showing the location of boreholes used in the investigation.

Boreholes BH 394 and BH 396 located immediately outside of the dewatered zone illustrate the strong hydraulic disconnection of the dewatered area from the surrounding sub-catchment. The hydraulic connection between the alluvium and surrounding aquifer is demonstrated by the similarity in the water level variations of boreholes BH 193 (thought to be located in the alluvium) and BH 195 (located some way from the channel). The observed groundwater level data is not referenced to a common datum and has been reported in metres below ground level (mbgl), therefore it is the relative level variations and the range over which the depths vary that are examined. The groundwater level data is assessed based on characteristics such as their response to likely flow events (evidenced by high rainfalls given no measured flow data). Boreholes BH 114, BH 162, BH 180 and BH 195 seem to be located in the surrounding regional aquifer while, the location of the remaining boreholes is uncertain and could include the alluvium, the surrounding regional aquifer or both aquifers.

The water levels of boreholes adjacent to the river are relatively dynamic and are expected to be influenced by rainfall and flow events. However, water levels in boreholes near the Gamagara River outside the dewatering zone show minor ranges in water levels when compared to boreholes in the dewatered zone. The large variations in water levels of the boreholes along the Gamagara River in the dewatered zone could be explained by either recharge emanating from surface flow in the river or by variations in groundwater pumping. Bredenkamp and De Jager (2011) concluded that water levels in the different aquifers (alluvial and bedrock aquifers) outside the dewatered zone seem to be in equilibrium with each other and have the same piezometric surface. However, without the borehole data being referenced to a common datum, this cannot be confirmed. The authors also stated that the groundwater levels emulated the topography in 1963-1973 in the area of the current dewatered zone and that the groundwater levels in the Gamagara alluvial aquifer were relatively shallow and likely to be un-impacted during 1963 to 1973 (Figure 4-18 - BH's 394 and 396).

4.3.4. Model setup and results

Quaternary catchment D41J was divided up into four spatial units consisting of two tributaries upstream of the dewatered zone (Daniëlskuil and Postmasburg branches), the dewatered zone and the area downstream of the dewatered zone. The two upstream tributaries were designated spatial units 1 and 2; the dewatered zone was designated spatial unit 3 and the downstream area designated spatial unit 4 (Figure 4-18). These areas were modelled as individual but connected sub-catchments. With no gauged flow data in the catchment, the results of the simulation were compared to groundwater level data. Data from boreholes were available in all four parts of the catchment both before and after dewatering.

The Pitman model does not explicitly simulate groundwater levels; rather a groundwater gradient is simulated (Section 3.5.3.3 - Figure 3-14). For a more direct comparison between the simulated and observed groundwater levels, the near channel groundwater gradient was converted to a groundwater level in metres below the channel by using the width of the near channel gradient (m) in a simple trigonometric function (Figure 4-19):

Level = (Simulated gradient × width of near channel slope / 100) × 1000 **Equation 22**

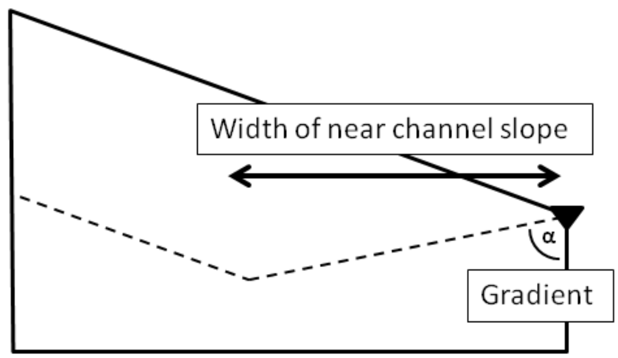


Figure 4-19 Conceptualisation of model gradient and its conversion to a groundwater level. Gradient α represents the complementary angle.

It is not possible to explicitly compare the simulated water levels with any of the observed groundwater levels. The latter will vary depending on the location of the boreholes, while there is a single simulated level. Therefore the simulated groundwater levels are largely assessed based on comparisons of the temporal patterns of variation of the observed and simulated groundwater levels. Since the entire compartment is contained by dykes and groundwater levels throughout the unit are lowered, the groundwater abstractions were applied to both the near (40% of abstraction volume) and far (60% of abstraction volume) groundwater gradients.

The model was established with standard inputs (rainfall and evaporation demand) based on the regional information available in WR90 (Midgley *et al*., 1994), as well as the WR90 recommended parameter values for the surface water components of the model. The groundwater parameters were set up for alluvial aquifer conditions and therefore many of the values are higher than those estimated by GRA II (DWAF, 2005a) which represent the entire quaternary catchment. The GRA II parameters (for the entire quaternary catchment) and the Pitman model parameters for each spatial unit in the sub-catchment are given in Table 4-6.

Table 4-6 Parameter values used in the model simulation.

Parameter	GRA II (DWA, 2005a)	Spatial unit 1	Spatial unit 2	Spatial unit 3	Spatial unit 4
Recharge (mm y^{-1})	1.6 – 18.0	3.38	3.38	3.38	3.38
GW		92	92	92	92
GPOW		4.0	4.0	4.0	4.0
ST		150	130	150	130
Channel loss TLGMax (mm)		2.0	2.0	1.8	1.8
Transmissivity (m^2 d^{-1})	32.14	30	30	50	50
Storativity before dewatering	0.001	0.006	0.002	0.015	0.015
Storativity after dewatering		0.006	0.002	0.002	0.008
Drainage density	-	0.2	0.2	0.2	0.2
Rest water level	10	40	40	55	30
Regional groundwater drainage slope	0.002	0.002	0.002	0.002	0.002
Riparian strip factor (% slope width)		0.30	0.37	0.37	0.27

Figure 4-20 illustrates the effects or sensitivity of the different parameters on the simulated groundwater level and shows spatial unit 3 before dewatering with the groundwater levels each representing various parameter combinations.

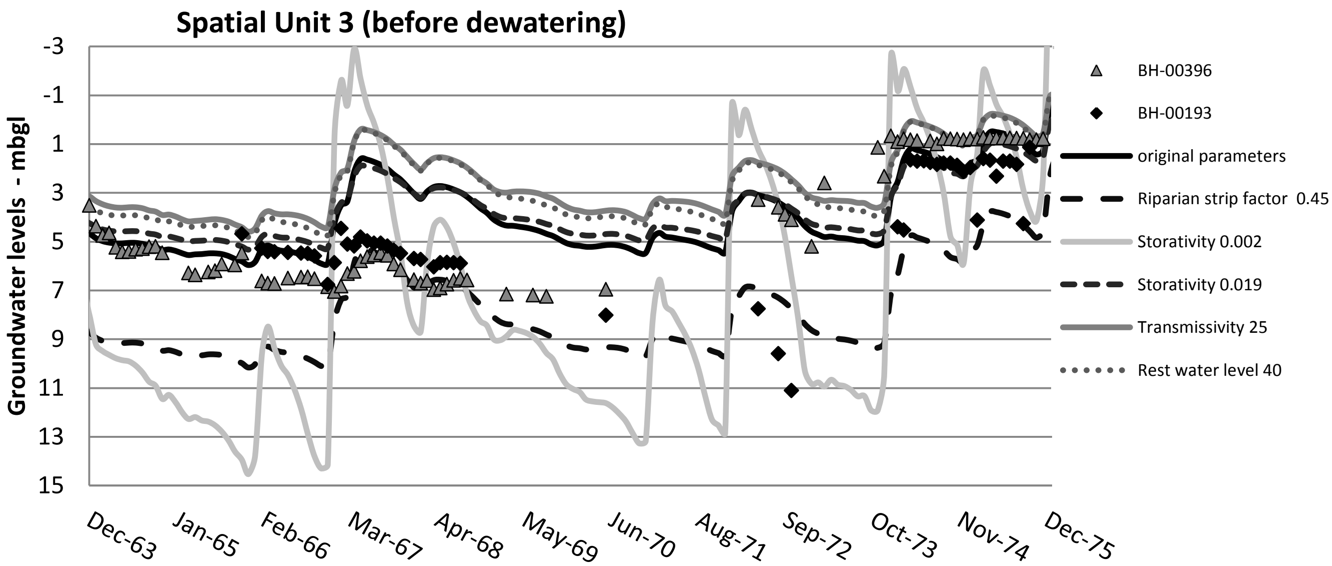


Figure 4-20 Spatial unit 3 (dewatered zone)with groundwater levels represented as metres below ground level (mbgl) before the start of dewatering showing the effects of different parameter combinations on the simulated groundwater levels.

The dark solid line represents the original parameters used in this study and given in Table 4-6. It seems the two most sensitive parameters include the riparian strip factor which affects the riparian evapotranspiration volume and the storativity parameter which has a large effect on the shape of the groundwater slope with a low value leading to a more variable groundwater level. The model was initially run without the groundwater abstraction (dewatering) incorporated into the simulation and the outflow from the most downstream spatial unit (4) compared with the national 'naturalised' or synthetic stream flow data (Midgley *et al*, 1994) for the entire quaternary catchment (Figure 4-21). The simulated flows which incorporate the groundwater abstraction are also shown to demonstrate the impact that dewatering has on the simulated flows. Table 4-7 provides comparisons of the flow volumes for three percentage points on the flow duration curves.

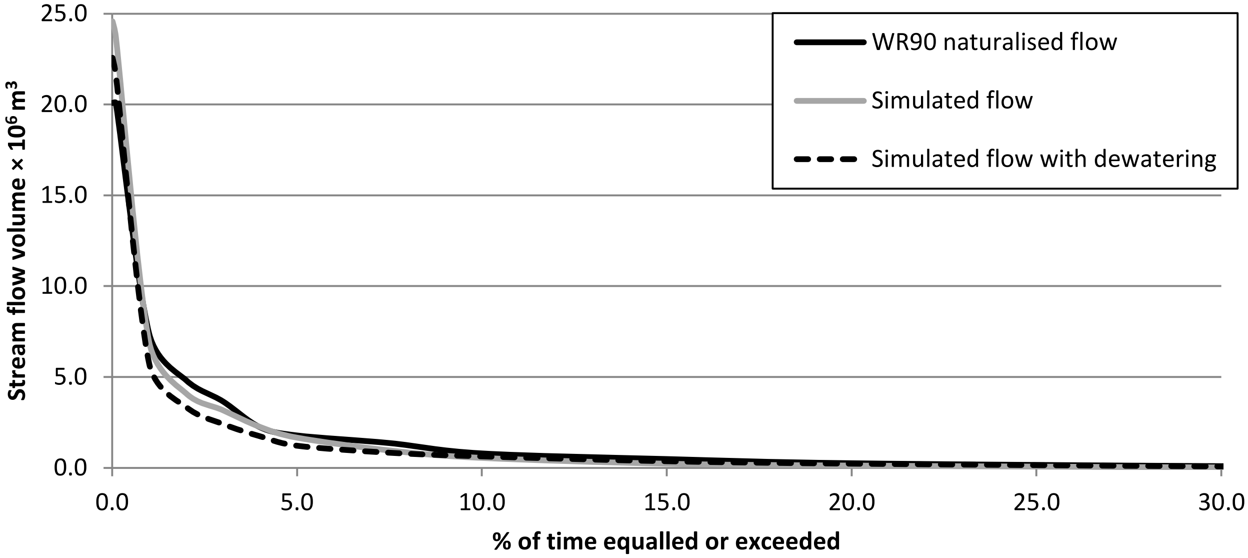


Figure 4-21 Flow duration curve comparing the WR90 naturalised flow with the Pitman model simulated flow (without groundwater abstraction).

Table 4-7 Comparison of flow volumes (10^6 m^3) at selected flow duration curve % points.

% of time equalled or exceeded	1%	5%	10%
WR90 naturalised flow	7.39	1.78	0.79
Simulated flow under natural conditions	6.71	1.55	0.65
Simulated flow with groundwater abstraction	5.75	1.21	0.63
% reduction due to dewatering	14%	22%	3%

The simulated flow from both natural time series ceased at 40% on the flow duration curve and were well matched in terms of flow volumes, although the simulated natural outputs were slightly less than the WR90 naturalised flows in terms of medium and high flows. Subsequently, the groundwater abstraction was incorporated in the dewatered zone (spatial unit 3) and the Pitman model rerun. While the simulated flow with groundwater abstraction incorporated also ceased at 40% of time on the flow duration curve, the volumes for all the flows (high to low) were obviously less than the natural flows. The results of the comparisons between the simulated groundwater 'levels' and the observed groundwater levels are given in Figures 4-21 to 4-26.

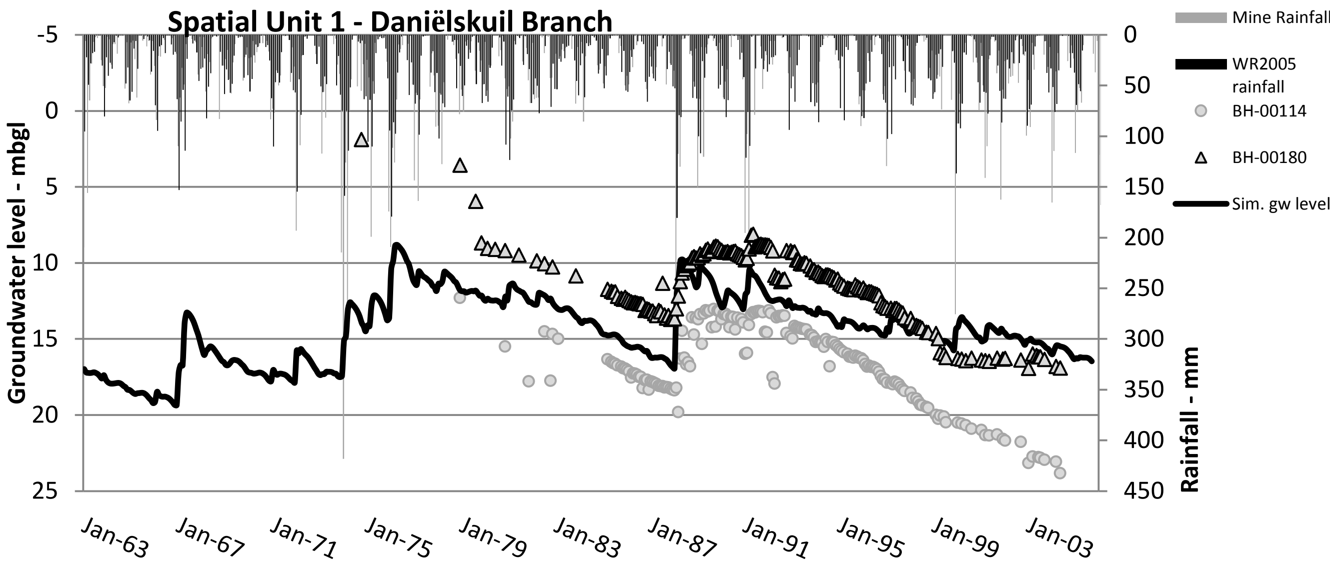


Figure 4-22 **Spatial unit 1 (Daniëlskuil branch – upstream of dewatered zone) groundwater levels represented as metres below ground level (mbgl) with both WR2005 and the mine rainfall included.**

The boreholes located in spatial unit 1 (Figure 4-22) are situated approximately 6 km (BH 114) and 10 km (BH 180) from the main channel and are therefore assumed to be positioned within the regional aquifer. The groundwater levels recorded in these boreholes therefore rely exclusively on rainfall for recharge. The simulated groundwater level is probably too dynamic or not as smooth as it should be when compared to the observed water levels, however, the model does capture the general water level patterns.

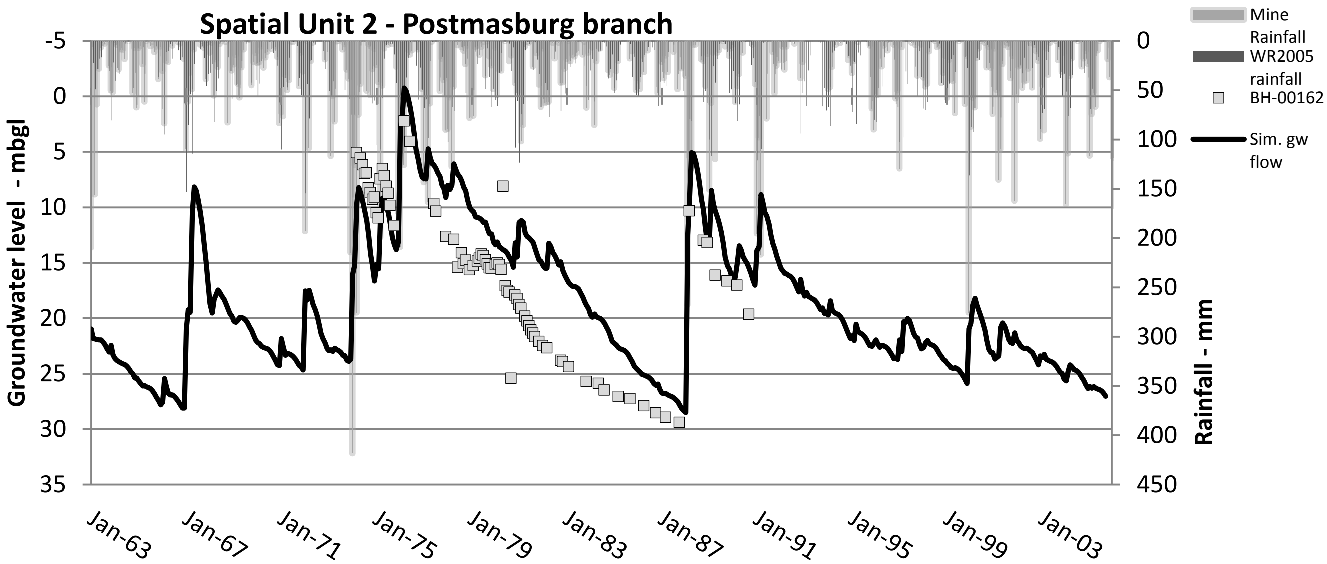


Figure 4-23 Spatial unit 2 (Postmasburg branch – upstream of dewatered zone) groundwater levels represented as metres below ground level (mbgl) with both WR2005 and the mine rainfall included.

Borehole BH 162 is located near the channel and is therefore assumed to be located within or at least partly within the alluvium. The geology of this spatial unit, however, largely comprises dolomite with little alluvium observed on the geological map. It is clear, however, that the response of the observed groundwater levels are fairly dynamic, indicating that channel transmission losses probably influence the levels. Perhaps this is associated with a narrow channel of alluvium not evidenced at the scale of the geological map, or a well-developed karst aquifer. The model was set up to reflect a more dynamic response of the groundwater level and respond more quickly to recharge events. The simulated increases in water levels shown in Figure 4-23 are associated with major rainfall recharge events and local transmission losses during those periods when the model generates stream flow. The decreases are mostly associated with simulated riparian evapotranspiration losses as downstream groundwater flows will be small because of the low gradients.

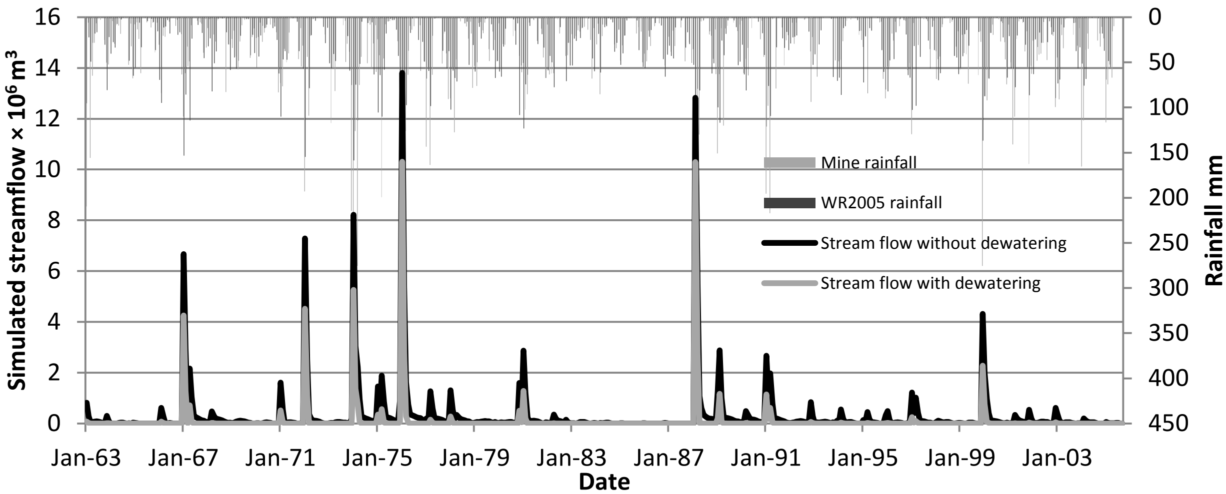


Figure 4-24 **Simulated stream flow and rainfall data for spatial unit 3.**

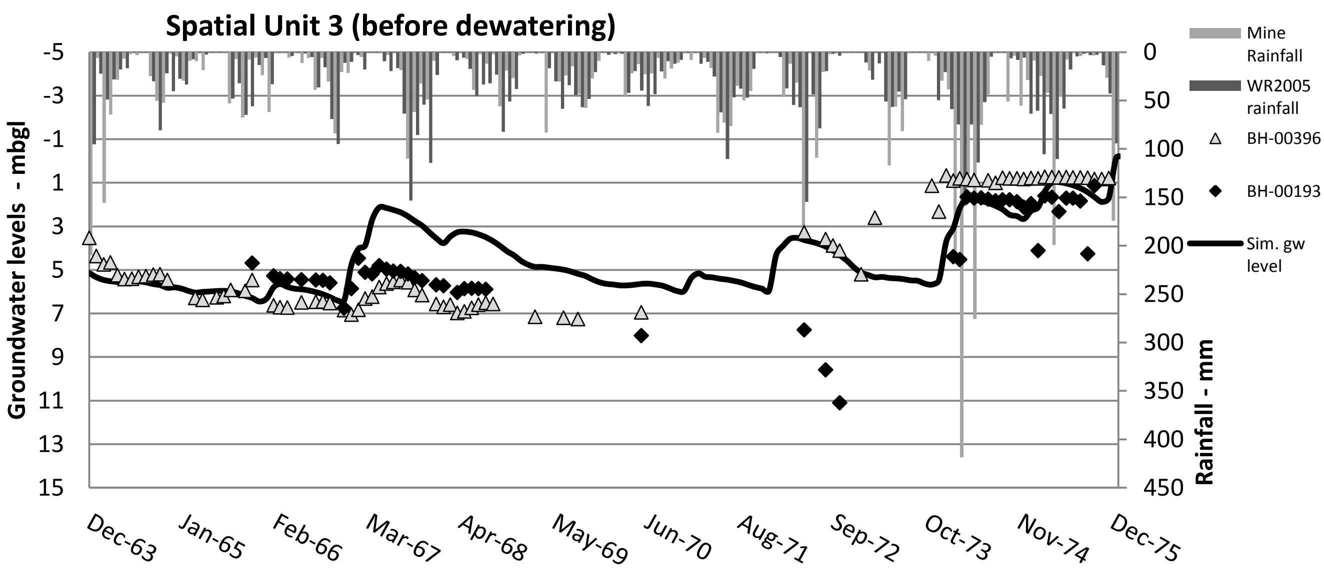


Figure 4-25 **Spatial unit 3 (dewatered zone) groundwater levels represented as metres below ground level (mbgl) before the start of dewatering, with both WR2005 and the mine rainfall included.**

Figure 4-25 indicates that the model captures the increase in water levels for the 1971 wet season and also simulates the large recharge events in 1973 and 1974 (Figure 4-24). High transmissivity and storativity parameter values were included in the model setup as it was assumed that the relatively shallow groundwater levels fell within the overlying alluvial aquifer. This is consistent with the fairly stable and non-variable nature of the observed groundwater levels.

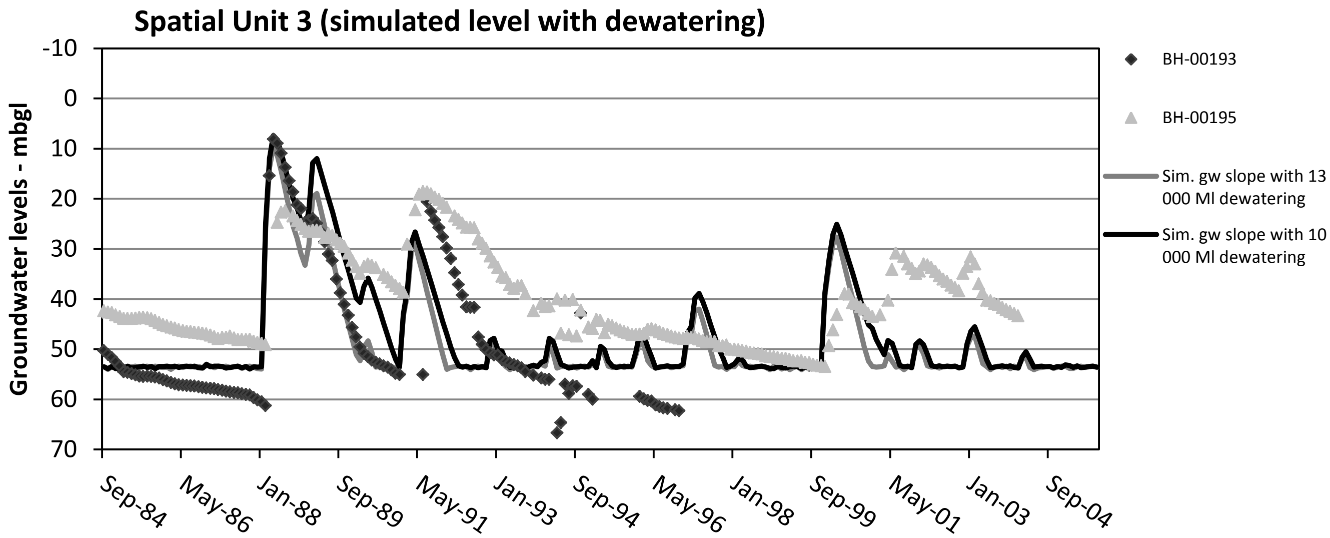
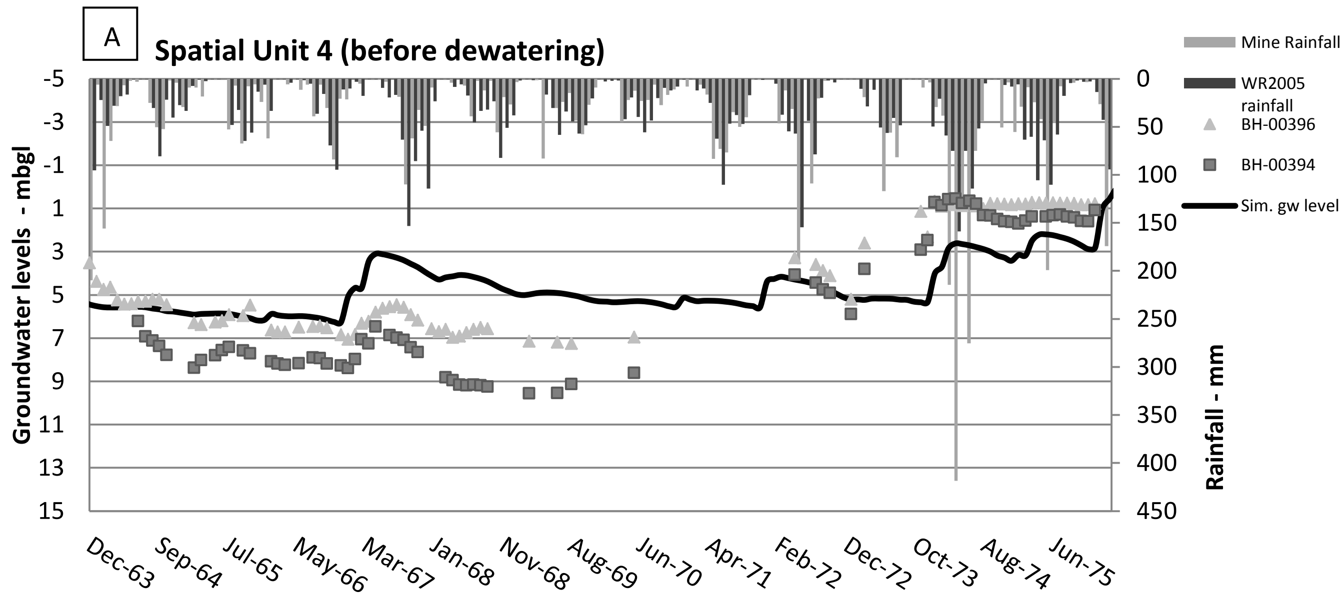


Figure 4-26 Spatial unit 3 (dewatered zone) groundwater levels represented in metres below ground level (mbgl) including a simulated groundwater level without groundwater abstraction (dewatering) and two simulated groundwater levels with different volumes of groundwater abstraction incorporated (10 000 and 13 000 Ml y^{-1}).

The storativity parameter value was decreased for this model run (Table 4-6) as the reduced groundwater level was assumed to fall within the underlying regional aquifer (the alluvium has been completely dewatered). This increased the variability of the simulated groundwater level, consisted with observations in the observed data. The aquifer was far deeper than the aquifer depth reported by the GRA II database (10 m) with measured borehole levels in the dewatered zone reaching a maximum of 68 m. The simulated groundwater levels are more comparable to borehole BH 193 located near to the channel which is to be expected as the model is simulating the near channel groundwater levels. The simulated variations in groundwater levels in Figure 4-26 are dominated by the transmission losses simulated by the model and therefore by the simulated upstream inflows. Both these and the transmission loss function are highly uncertain without any observed stream flow data. The results after dewatering are somewhat confused by the non-stationary effects of dewatering, while the model simulates stationary conditions with fixed parameter values. The model does not simulate a large enough range of groundwater level change (see 1992 to 1994). The extent to which this result is related to variations in the rate of dewatering or inadequate simulations of transmission losses cannot be resolved without additional information. However, since both observed boreholes (one near and one far from the channel) reflect large fluctuations in groundwater levels, it is likely that the fluctuations are due to the effects of dewatering. It is also important to note that the transmission losses are not only dependent on the structure of the losses function, but also on the amount of stream flow simulated by the model, which cannot be assessed in the absence of observed data.



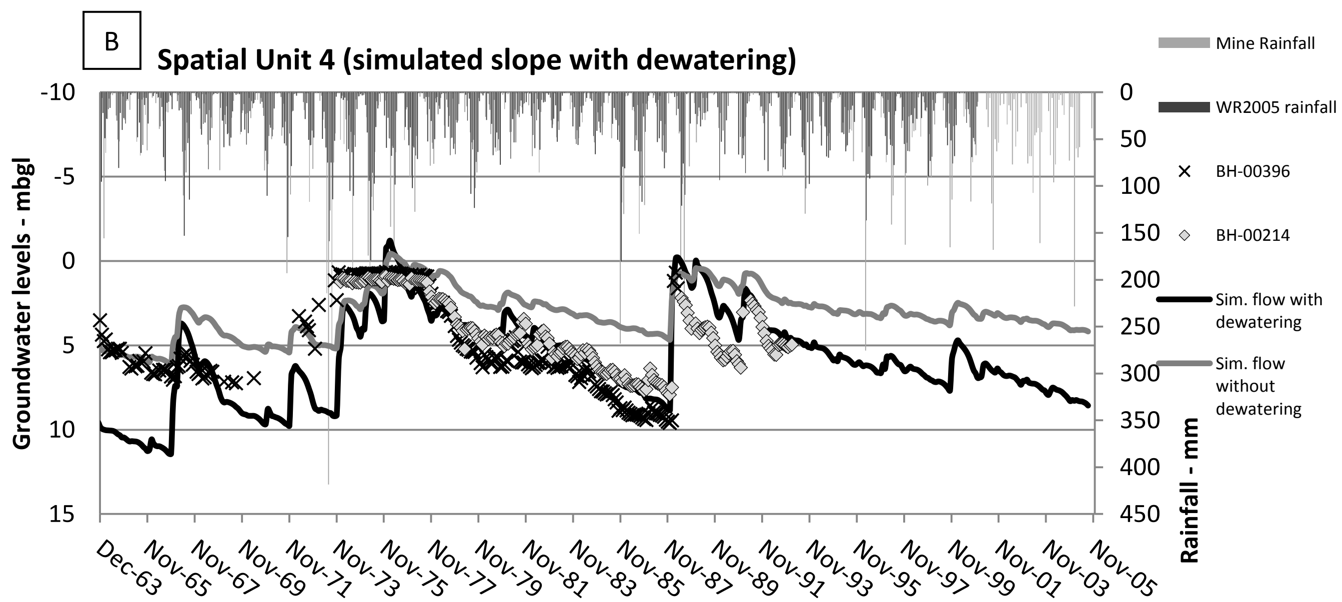


Figure 4-27 Node 4 (downstream zone) groundwater levels before the start of dewatering (A) and after the start of dewatering (B), with both WR2005 and the mine rainfall included.

Figure 4-27 illustrates that the comparison between the observed and simulated groundwater levels in the downstream catchment (below the dewatered zone) visually follow the same pattern. The comparisons before dewatering are very similar to those within the dewatered zone (Figure 4-27A) largely due to very similar parameterisation which included high storativity and transmissivity values consistent with alluvial aquifer characteristics. Figure 4-27B shows a small reduction in groundwater levels compared to those prior to dewatering. While the storativity value was decreased due to the assumption that the alluvium is fully dewatered at this point and the groundwater level falls within the underlying aquifer, it was not reduced as much as the storativity value in the dewatered area, as the dewatering had less of an effect in spatial unit 4 and the groundwater levels were far more shallow. As the downstream area is assumed to be hydraulically isolated from the dewatered zone, the differences in both the observed and simulated levels after dewatering are assumed to be associated with reduced stream flows and therefore reduced availability of water for transmission losses (recharge from the channel). The model calculates two components of channel loss, that from incremental runoff generated within the spatial unit (not affected by upstream dewatering) and that from flow in the main channel generated upstream of the spatial unit (affected by upstream dewatering). Table 4-8 provides the water balance components for the final simulations and all of the values are within expected ranges for this type of catchment.

Table 4-8 Gamagara water balance results for the simulations (10^6 m^3 $month^{-1}$).

	N1	N2	N3	N3	N4	N4
	Danielskuil	*Postmasburg*	*Dewatered zone* before dewatering	*Dewatered zone* after dewatering	*Downstream* before dewatering	*Downstream* after dewatering
Area (km^2)	1580	456	316	316	1522	1522
INPUTS						
Groundwater inflow	0.000 (0%*)	0.000 (0%)	0.039 (20.5%)	0.039 (16.1%)	0.032 (6.0%)	0.003 (0.6%)
Recharge	0.370 (100%)	0.128 (100%)	0.074 (39.0%)	0.074 (30.6%)	0.415 (77.4%)	0.415 (86.6%)
Channel loss	0.000 (0%)	0.000 (0%)	0.077 (40.5%)	0.129 (53.3%)	0.089 (16.6%)	0.061 (12.7%)
OUTPUTS						
Groundwater outflow	0.027 (7.3%)	0.012 (9.3%)	0.032 (16.8%)	0.003 (1.2%)	0.064 (11.9%)	0.058 (12.1%)
Riparian evapotranspiration	0.343 (92.7%)	0.116 (90.6%)	0.158 (83.2%)	0.239 (98.8%)	0.472 (88.1%)	0.421 (87.9%)
Simulated outflow	0.200	0.029	0.202	0.137	0.366	0.321

*% of the different water balance components of either the inputs or outputs.

4.3.5. Conclusion

While there is still a large amount of uncertainty associated with understanding the dynamics of this environment, it was fairly straightforward to set the model up using the available data with some knowledge of the physical environment. The large volume of channel losses in the dewatered area is due to the simulated gradient being maintained at high negative levels due to the dewatering (Figure 3-14C). The contributions to groundwater through transmission losses are subject to uncertainties in the simulated stream flows as well as the transmission loss function itself. The rate of groundwater level rise is dependent upon an accurate representation of rainfall and upstream inflows as well as an accurate portrayal of transmission losses in the model so it is unrealistic to expect the simulated groundwater levels to exactly represent the observed data. A realistic evaluation can only really assess the response of the simulated groundwater levels to rainfall events. In addition, the comparisons between the observed and simulated groundwater levels within the dewatered zone are somewhat confused by the lack of detailed information on the history of pumping rates. There is additional uncertainty in this particular environment related to the proportioning of the groundwater gradient into 40% (near channel) and 60% (far from channel). The far channel slope is intended to represent the regional aquifer, while the 40% gradient is designed to represent the more dynamic near channel environment. It is not known if these percentages are appropriate in this setting although the comparisons with observed data suggest that the proportioning is appropriate. This is probably due to the seemingly strong hydraulic connection between the alluvium and surrounding aquifer. In boreholes located some way from the channel, the expectation is that there will be less variability in groundwater levels as effects of alluvial recharge (transmission losses) will be less. However, groundwater levels in boreholes located both far from the channel (BH 195) and near to the channel (BH 193) seem to be fairly dynamic with significant fluctuations. This environment represents a dual aquifer (alluvium and surrounding regional aquifer) which are not straightforward to simulate at larger scales. There is a lack of adequate structure in the Pitman model to represent these types of environments, however, different sets of parameters were used for when the groundwater levels were assumed to be within the alluvium (before dewatering) and within the regional aquifer (after dewatering). Despite these limitations, the model has reproduced most of the observed variations in groundwater levels. It has therefore been concluded that the model is representing the dominant, catchment scale, processes (recharge, intermittent stream flow, groundwater evapotranspiration losses, transmission losses and dewatering effects) in a behavioural manner, even if the exact quantification of these processes remains uncertain.

4.4. Karst environment – Molopo dolomitic eye

4.4.1. Introduction

Reliable characterisations of surface and groundwater interactions in South African karst environments are not straightforward, and models have had great difficulty simulating the processes in these settings. While there are observed flow data in this sub-catchment and a number of investigations have been carried out (Bootsman, 1997; Bredenkamp, 1998; 2007), wide ranging recharge estimates together with uncertain catchment divide boundaries render it difficult to characterise this environment satisfactorily. These uncertainties have been examined in terms of their effects on the final model outputs.

4.4.2. Description of study area

The catchment (designated D41A) forms part of the Molopo Drainage region in the Lower Vaal Water Management Area (Figure 4-28). The region is semi-arid and experiences hot summer months and cold dry winters. The annual rainfall in the southern Kalahari region is highly variable; MAP is 500-600 mm y^{-1} and the mean annual potential evaporation is 1800 to 2000 mm y^{-1} (Midgley *et al*.,1994). Extensive dolomite rock formations are found around the Mafikeng/Lichtenburg area (Figure 4-29). Wilson *et al*., (2006) concluded that the dolomites have undergone at least four phases of karst development which has led to a series of weathering features in the dolomite consisting of depressions and pans. These are in-filled by alluvial material. Surface water is scarce throughout the area; the absence of water over such an extensive area of gravel-covered dolomite is ascribed to the advanced development of the karst leading to an aquifer with a high carrying capacity. Where the formation is heavily jointed, considerable quantities of water may be present (von Backstom *et al*, 1953). A dolomitic eye (Figures 4-30 and 4-31), approximately 30 km east of Mafikeng, has a flow gauge (D4H014) monitored by the department of Water Affairs (DWA) which were used to assess the model simulation results.

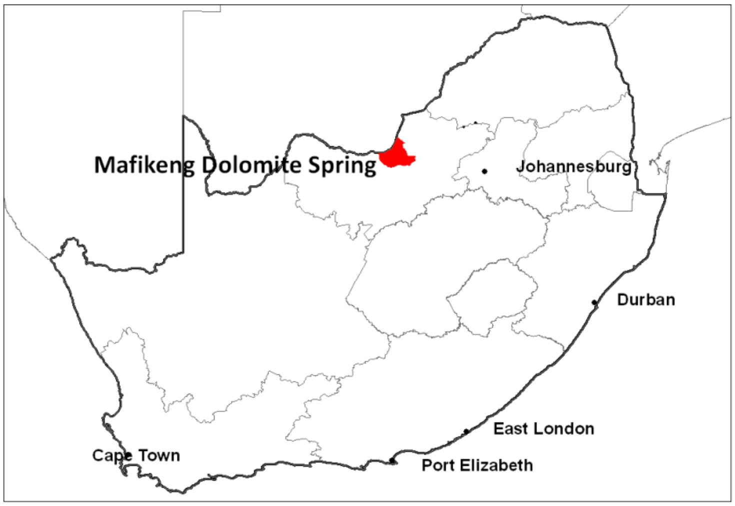


Figure 4-28 Location of the Molopo dolomitic eye in South Africa.

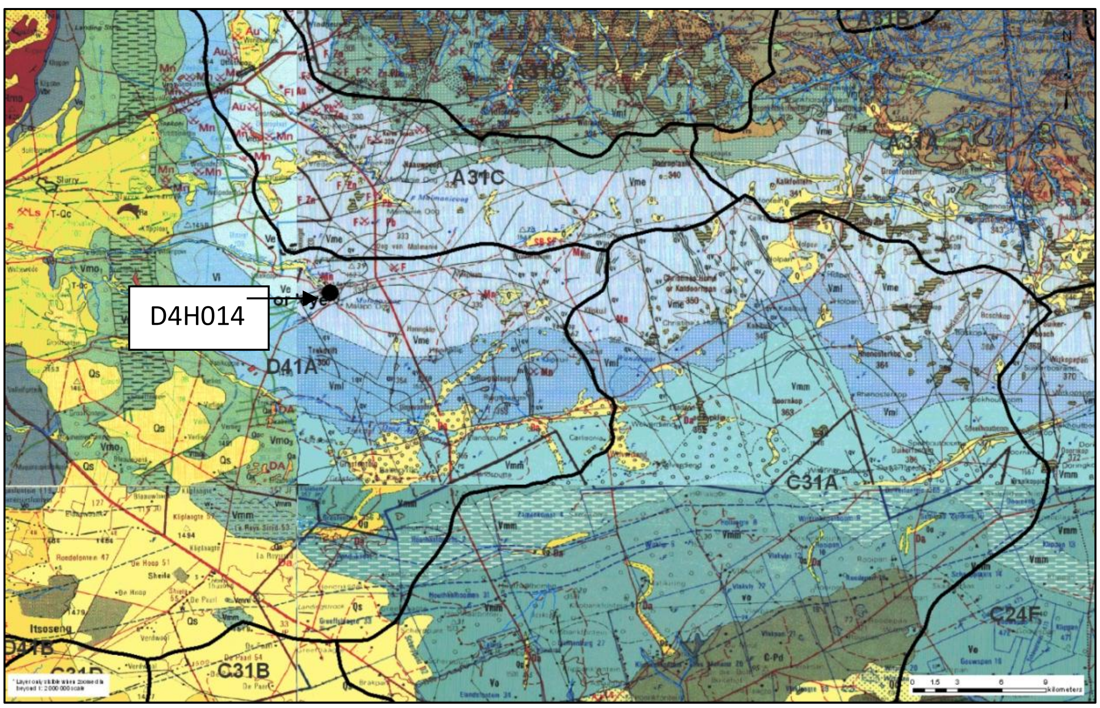


Figure 4-29 Sub-catchment D41A showing the geology of the region. Underlying the dolomite is granite shown in red (west), the band of dolomite is represented by the green and blue layers, overlying the dolomite are brown shales and slate (north and east) and the overlying Kalahari sands are shown in yellow (Geological survey, 1981; 1991).

Figure 4-30 Dolomitic eye with downstream flow gauge (Figure extracted from Google Earth– Eye elevation: 3.00 km). According to the DWA the flow gauges are 12 m apart.

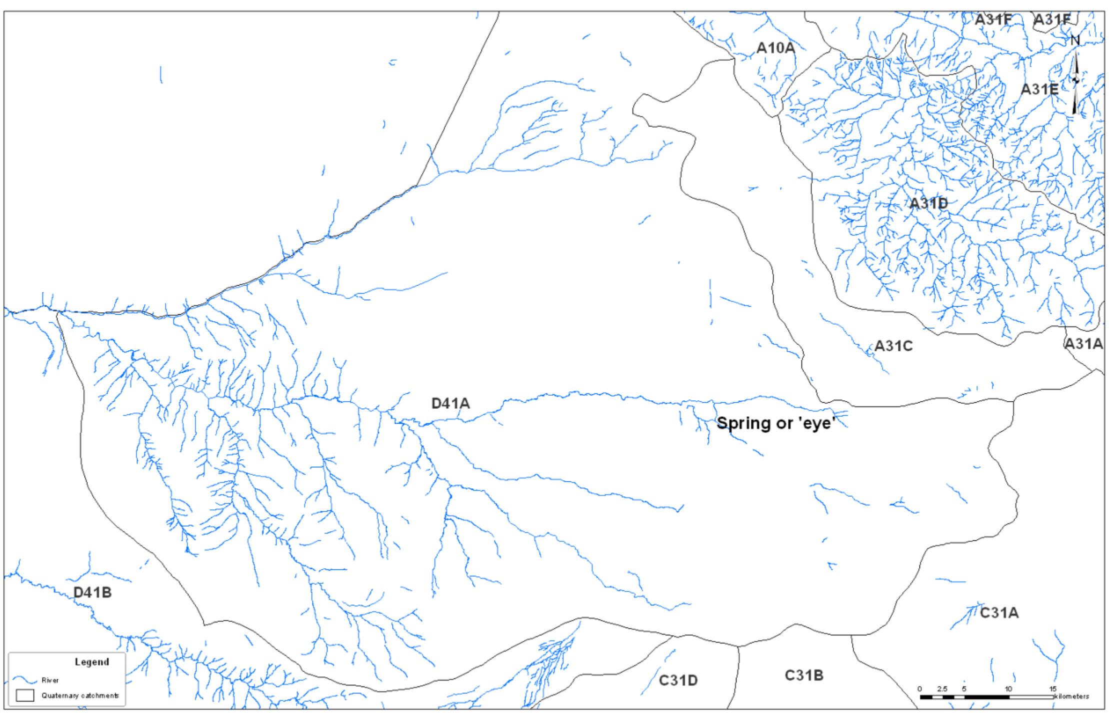


Figure 4-31 Sub-catchment D41A showing the river network and location of the spring or 'eye'.

Due to the presence of diamonds in the area, the geology has been well documented (Wilson *et al.*, 2006; Marshal and Norton, 2009). According to DWA (Midgley *et al.*, 1994) the upstream catchment area of the gauging station is 22 km^2.

4.4.3. Conceptualisation of the interaction processes

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 reliable estimates of the catchment area (Bailly-Comte *et al.*, 2009). The three recharge values given by GRA II (Conrad, 2005a) 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 4-9). 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 4-9.

Table 4-9 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}; Conrad, 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 4-9 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. Figure 4-32 illustrates the extent of the different catchment areas (using simple circular areas) used in Table 4-9.

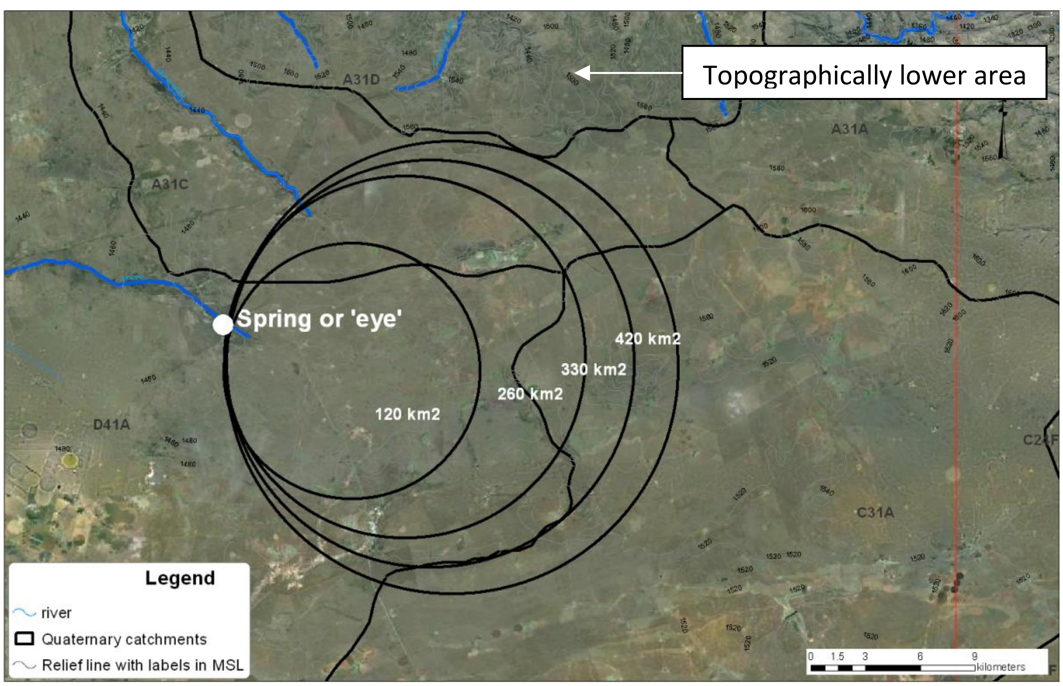


Figure 4-32 Possible groundwater catchment divides, based on available recharge estimates and spring outflow, overlying surface water catchment divides.

The area to the north of the spring is topographically lower so is unlikely to be part of the contributing area, while the rivers flowing east also provide a cut-off point of the possible contributing area. The model was set up for three different catchment sizes (Table 4-9) with the corresponding recharge volume (identified as within the range of the most probable estimates of recharge – 26.02 to 42.33 mm y^{-1}).

The outflow records from the spring (D4H014) are shown in Figure 4-33 along with the rainfall data (WR90, Midgley *et al.*, 1994). There seems to be a small time lag between the rainfall and the spring outflow.

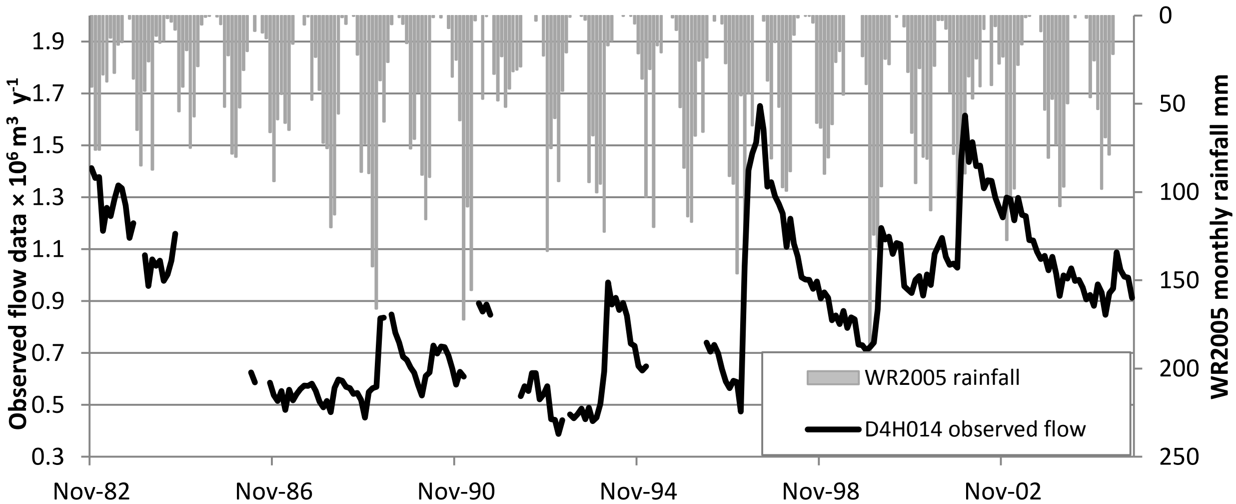


Figure 4-33 Observed monthly stream flow data compared to rainfall (WR2005 – Bailey, 2007) in the Mafikeng area.

4.4.4. Model setup and results

Table 4-10 provides the groundwater parameter values for each of the model simulations carried out. While the recharge parameters were adjusted accordingly, it was not necessary to adjust the remaining groundwater parameters as the resulting simulations were fairly similar (Figures 4-34 and 4-35). The surface runoff parameters were set to generate no surface runoff as records within the basin suggest there are none.

Table 4-10 Model parameters, simulated volumes and model performance statistics for each of the model simulations.

	Simulation 1	Simulation 2	Simulation 3
Catchment size (km^2)	260	330	420
Groundwater parameters			
ST	70	70	70
FT	0	0	0
POW	0	0	0
GW	28	17	14
GPOW	3	3	3
Drainage density	0.3	0.3	0.3
Transmissivity (m^2 d^{-1})	25	25	25
Storativity	0.008	0.008	0.008
R	1	1	1
Recharge (mm y^{-1})	42.33	33.00	26.02
Simulated outflow × 10^6 m^3 y^{-1}	11.35	9.52	10.2
Model performance			
Nash Coefficient (Untransformed)	0.25	0.21	0.34
Nash Coefficient (Ln transformed)	0.33	0.30	0.37
Nash Coefficient (Inverse transformed)	0.34	0.33	0.35
% Bias in simulated monthly flows	-2.84	0.90	2.90
% Bias in simulated monthly ln(flows)	58.69	-120.13	-110.11

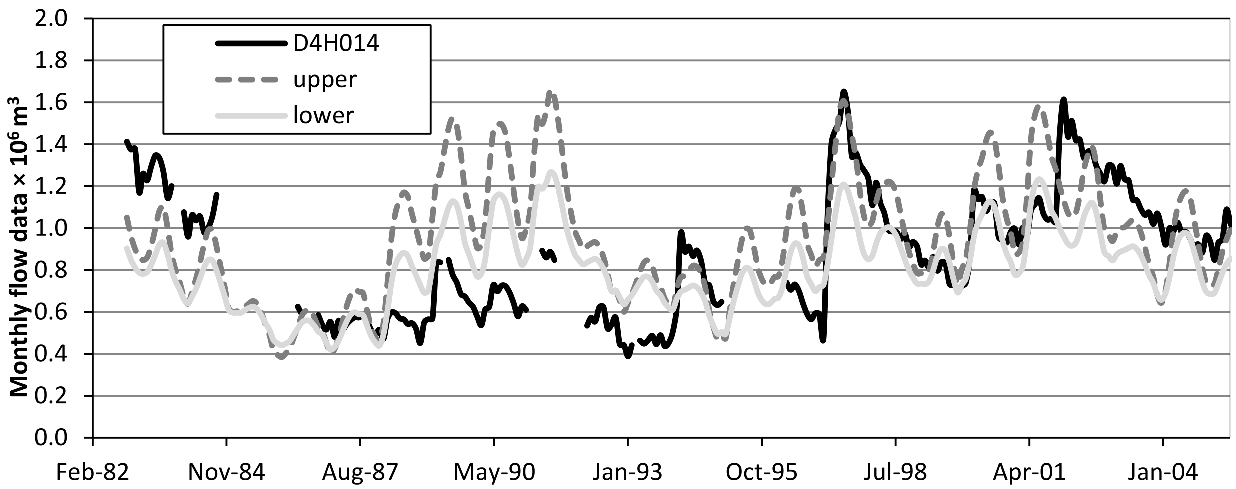


Figure 4-34 **Time series graph comparing observed spring flow from a dolomitic eye (D4H014) to the flow simulated by the Pitman Model. 'Upper' represents the highest volumetric simulation (Table 4-10 - Simulation 1) while 'lower' represents the lowest volumetric simulation (Table 4-10 - Simulation 2).**

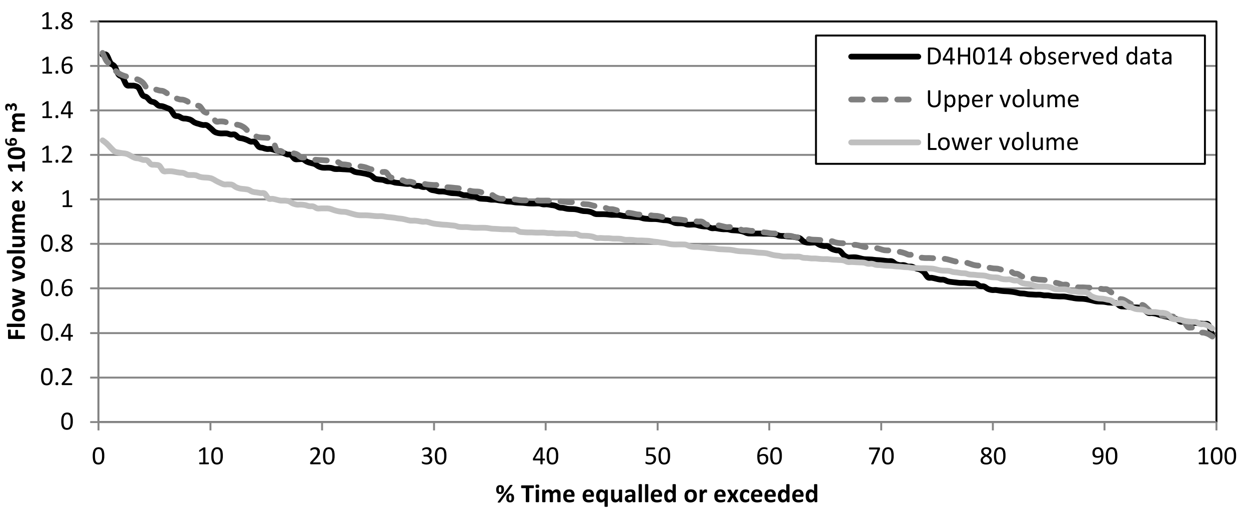


Figure 4-35 **Comparison of observed spring flow from a dolomitic eye (D4H014) to the flow simulated by the Pitman Model. 'Upper' represents the highest volumetric simulation (Table 4-10 - Simulation 1) while 'lower' represents the lowest volumetric simulation (Table 4-10 - Simulation 2).**

A comparison between the simulated outflows and the observed flow data showed that the model was able to capture the general pattern characteristics (Figure 4-34), but was unable to represent all the fluctuations present in the observed data. The statistics of model performance are given in Table 4-10.

The volumes of flow simulated using the different catchment boundaries varied from just below (9.52×10^6 m^3 y^{-1}) to just above (11.35×10^6 m^3 y^{-1}) the observed flow volume (10.60×10^6 m^3 y^{-1}). While the flow patterns were generally captured by the model, in one period (1988 to 1992) the simulated flow volume was substantially higher than the observed flow data. No amount of calibration was able to improve the simulation of this period without affecting the remaining well simulated periods. It is not straightforward to identify the reason for the volume disparity, especially as the later years are satisfactorily simulated. Potential rainfall data discrepancies which were examined revealed differences in the data sets available (WR90, WR2005 and a local dataset) but none substantial enough to explain the volume difference during 1988 to 1992. The volume disparity could be due to a time lag between the rainfall and outflow from the spring as this process has been noted at other karstic sites. Similarly should a threshold need to be reached before an increase in outflow occurs in some part of the aquifer, this would also explain the flow discrepancy. However, no time lag is evident within later years of the simulation. A period of large abstraction during this time could account for the lower observed volumes, however it is unlikely that there was only one short intensive period of abstraction. Lastly flow gauge errors could have resulted in a period of incorrectly measured flow.

There was a degree of equifinality present during the calibration which was irresolvable due to the high parameter uncertainty in karst environments. However, this was only present for a few key parameters and did not significantly alter the simulation results. In particular the drainage density and transmissivity parameters were in a sense interchangeable, with similar results being obtained using high drainage density and low transmissivity parameter values to that of low drainage density and high transmissivity parameter values. Table 4-11 provides a set of alternative groundwater parameter values all based on a GW of 17 and a GPOW of 3 and Figure 4-36 illustrates the results using the four different parameter sets.

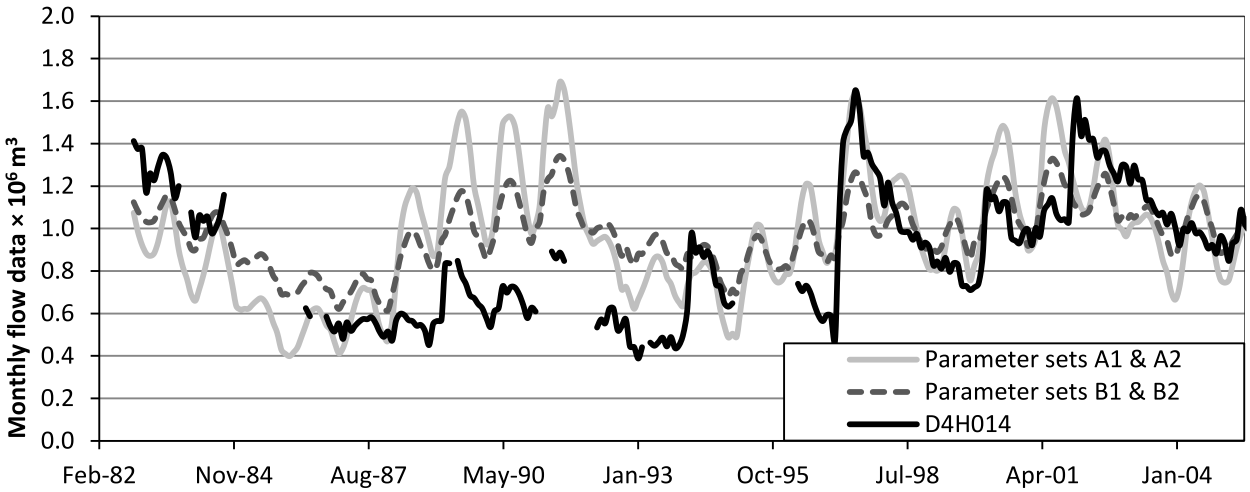


Figure 4-36 **Comparison between the observed spring flow from a dolomitic eye (D4H014) and the flow simulated by the Pitman Model (for two different parameter sets A and B).**

Table 4-11 **Four parameter sets demonstrating the equifinality present in this setting. Both A1 and A2 create very similar outflows (A in Figure 4-34) and similarly with parameters sets B1 and B2 (B in Figure 4-34).**

Parameter	A1	A2	B1	B2
Drainage density	0.1	0.3	0.3	0.1
Transmissivity (m2 d^{-1})	80	10	25	80
Storativity	0.008	0.01	0.008	0.003

Lower transmissivity estimates were used in the final simulations (Figures 4-34 and 4-35) as these values were similar to those transmissivity estimates reported by GRA II - 30 m^2 day^{-1} (DWAF, 2005a).

4.4.5. Conclusions

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 reliable 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 with the exception of one period (the reason for the discrepancy could not be determined). While there appeared to be a slight time lag between the rainfall and outflow from the spring, this did not seem to affect the model outputs which reflected no time lag.

4.5. Fractured rock environment: Upper Breede River

4.5.1. Introduction

This interaction environment forms part of the headwaters of the Breede River located in the intensively folded Cape Fold Belt (Figure 4-37) region of the Western Cape Province and includes important fractured rock aquifers which are complex due to intense tectonic activity in the past. The data available in the area are sparse and uncertain largely due to extreme spatial heterogeneity (steep mountain slopes and flat valley floors). While there is a flow gauge located downstream of the sub-catchments, large volumes of water use (both surface and groundwater) and return flows from an urban area suggest that the flow data are also not representative of natural hydrological responses. For this reason models that have been applied in the region (Conrad, 2005; DWAF, 2006b) and in other TMG aquifers (Roets *et al*., 2008; Parsons, 2009) have not been sufficiently validated and many conflicting results have been reported. Previous investigations carried out on the Breede River include DWAF (2003b) and Steynor *et al*. (2009). The most thorough regional investigation was the Berg Water Availability Assessment Study (Berg WAAS) commissioned by the Department of Water Affairs (DWAF, 2007a) which included parts of the Upper and Middle Breede River. The investigation included a detailed characterisation of all the aquifers in the region using all available data and numerical groundwater models. Although no additional field data were collected, this study represents one of the more detailed investigations to be undertaken in South Africa for the purposes of water allocations that includes both surface and groundwater. Using the results from the investigation, the Pitman model was set up and run to assess whether it was possible to simulate the major interaction processes using a simpler model. Given the lack of real data, it was found to be very difficult to compare and validate the results of the detailed numerical model of the Berg WAAS study and the Pitman model; a common situation in the predominantly data scarce areas of the country. The investigation in this region therefore included an assessment of the dominant sources of uncertainty, how they affect the overall understanding of the surface-groundwater interactions, and how they may be reduced.

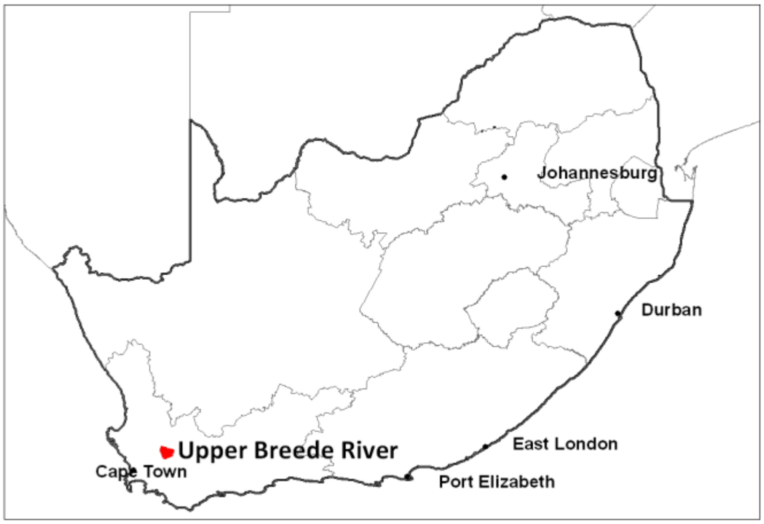


Figure 4-37 Location of Upper Breede River in South Africa.

4.5.2. Description of the study area

The headwaters of the Breede River are sub-divided into four sub-catchments designated H10A to D (Figure 4-38). The dominant land use activity is deciduous fruit orchards. The area experiences a typical Mediterranean climate with moderate temperatures and winter rainfall. The mean annual precipitation (MAP) varies significantly across the study area (500 to 1000 mm y^{-1}) due to the orographic influence of the topography, while mean annual potential evaporation is about 1600 – 1700 mm y^{-1} (Midgley *et al.*, 1994). The seasonal pattern of rainfall has an effect on recharge volumes since the aquifers are recharged in winter when the temperatures, and therefore evapotranspiration, are low (DWAF, 2008a). The flow regime is highly variable, partly as a consequence of the rainfall seasonality. The rivers are largely perennial and the baseflows are expected to consist of a large volume of unsaturated zone interflow due to numerous springs and seeps found on the steep slopes.

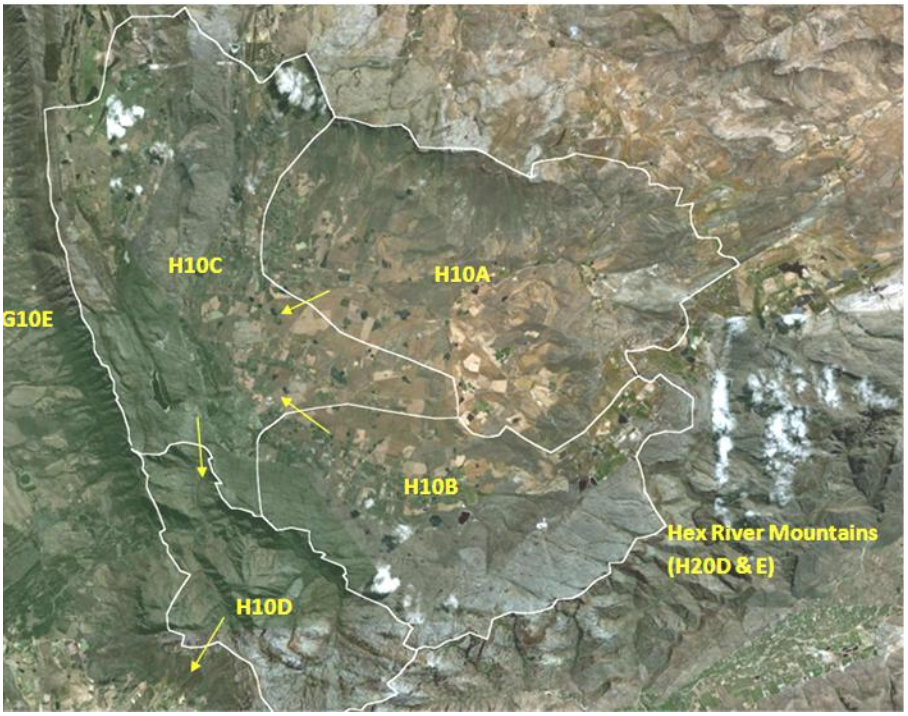


Figure 4-38 Sub-catchments H10A to H10D with arrows representing surface water flow.

The fold system is comprised of deep quartzite and the valley floors comprise poorly fractured shales and sandstones. The quartzite forms part of the Table Mountain Group (TMG) and is exposed on the top and flanks of the anticlinal folded mountains and is confined in the synclinal valleys by shale and sandstones termed the Bokkeveld Group (Figures 4-39 and 4-40). Elevation in the mountainous areas ranges from 1400 m to 2000 mamsl. The lower relief regions and valleys have elevations ranging from 200 m to 700 mamsl. The two major formations making up the TMG in the study area are the Peninsula (around 1400 m thick) and Nardouw (around 250 m thick) Formations; these are overlain by the Bokkeveld Group (approximately 1200 m thick) which in turn, is overlain by the Witteberg sandstone in the north of the catchment area.

4.5.3. Conceptualisation of the interaction processes

The fold system is the main structural element that determines the natural boundaries of groundwater and surface water flow. Surface water flow drains from the bounding mountain ranges into the valley and out through Mitchell's Pass in the south-western corner (H10D). The Berg WAAS formulated a perceptual model by incorporating available data into a GIS based model (before detailed groundwater modelling was carried out on selected portions). Individual aquifers were identified within each sub-catchment and characterised separately with recharge zones and discharge zones delineated. The perceptual model formulated by the Berg WAAS (DWAF, 2008a) is shown in Figures 4-39 and 4-40. They assume that the three main aquifers in the catchment follow different flow paths. The Bokkeveld shales and Nardouw TMG aquifers both have their recharge zones and discharge zones within the catchment (i.e. follow the topography). However, although the Peninsula TMG aquifer recharges in the catchment on the mountain tops, it is assumed to discharge in the Hex River catchment system to the east through deep groundwater flow (DWAF, 2008a).

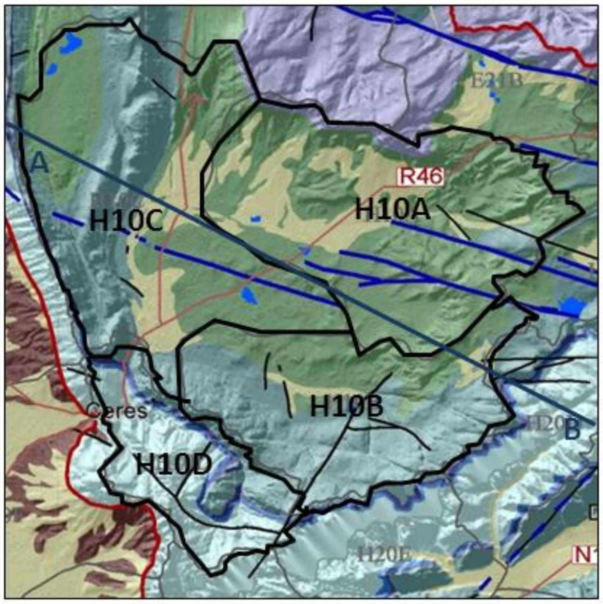


Figure 4-39 Simple geological map of sub-catchments H10A to H10D - interpretation of aquifer types in cross section below (adapted from DWAF, 2008a).

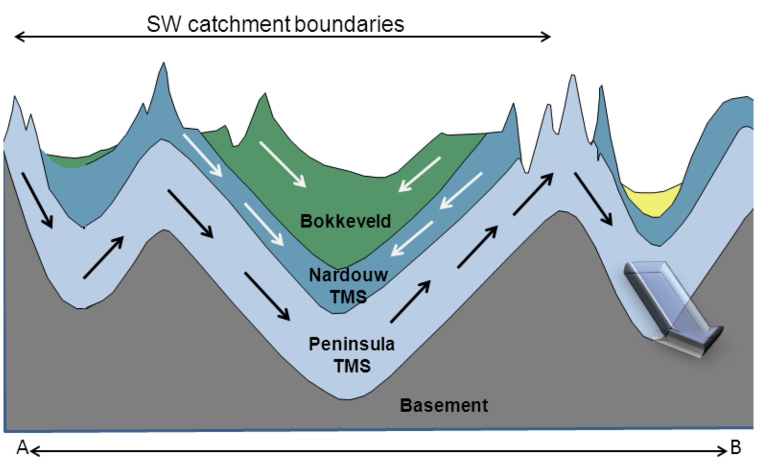


Figure 4-40 Simplified cross section A-B illustrating the assumed flow paths for the three major aquifers – colours correspond to geological map above (adapted from DWAF, 2008a).

The Berg WAAS compared a variety of recharge estimates from prior investigations which were found to vary significantly within the catchment. These methods included a variety of GIS based methods: GRA II (DWAF, 2006a); a rainfall-recharge relationship (BRBS Method; DWAF, 2002); an aquifer specific water balance model (ISP Method; DWAF, 2005b); a map-centric simulation method (DWAF, 2003a) and a water level fluctuation method (DWAF, 2007a). The results from each method were recalculated using the GIS model to obtain recharge per aquifer type. The GIS method recalculated the recharge using rainfall, runoff and evapotranspiration together with a delineation of recharge and discharge zones for each aquifer. The averages of the recalculated recharge from the different methods were used as the best estimate of recharge (DWAF, 2007a). The uncertainties in the recharge estimates were not therefore accounted for in the study.

While the conceptual understanding presented by the Berg WAAS seems to be sensible, there is a large amount of uncertainty in the available data, which inevitably translates into uncertainty within the perceptual model. Although regional groundwater flows are a possibility in a folded system such as the Ceres Basin, there are no data available to confirm the occurrence of the process and certainty no information available on the volume of water moving out of the catchment or the destination of the regional flow. The wide ranging recharge estimates also render the averaged recharge rather uncertain, as many of the estimates are clearly inappropriate. The difficulty in estimating reliable recharge values is partly related to the complexity of the rainfall in the region. The rainfall uncertainties are expected to be associated with the density and representativeness of the gauging network (many of the mountainous areas are not easily accessed) and the high degree of spatial variability of the real rainfall (Hughes and Mantel 2010a). A comparison of the rainfall estimates from the two most recent national water resource databases, WR90 and WR2005, reveals that there were clearly different assumptions made about the orographic effects in many areas. For example, in the Upper Breede River, the WR90 (Midgley *et al*., 1994) study assumed a mean annual rainfall across quaternary catchment H10C of 674 mm y^{-1} compared with 1064 mm y^{-1} for the WR2005 (Bailey and Pitman, 2005) analysis. The Berg WAAS used a rainfall dataset specifically developed for the project (DWAF, 2007b) and assumed a mean annual rainfall of 862 mm y^{-1} across quaternary catchment H10C.

Even though historical records of data are available in the catchment (flow gauge H1H003 located at the outlet of sub-catchment H10C), there is the problem of unaccounted for and poorly documented upstream human interferences and it is difficult to isolate the natural hydrology regime of the catchment. The observed data for flow gauge H1H003 records flows from 1923 to the present. These data were patched during the WR90 (Midgley *et al.* 1994) study to provide estimates at times when there were missing data. A similar patching exercise for H1H003 was carried out in the Institute for Water Research (Hughes and Mantel, 2010b) which found that there were a large number of high flow periods that recorded stages above the gauge rating curve limit. Therefore, no information on the volume of flow during those periods exists within the records. In addition, the low flows are partly impacted by wastewater return flows and storm runoff from the urban area of Ceres (with a population of about 40 000 and located close to the outlet of H10C), which would be reflected in the observed records.

Water use in the catchment is dominated by the operation of small farm dams (there are more than 350 farm dams in the study area), run of river abstractions and groundwater abstractions. The degree to which the farm dams are fed by surface water or groundwater is unknown. Examining aerial images of the catchment indicates that most of the farm dams are located near the main channels or tributary channels, which seems to indicate that the dams are mostly fed by surface water. Hughes and Mantel (2010b) carried out a study of uncertainties in simulations of natural and modified stream flow regimes within the total H10 catchment. Their results suggest that the impacts on downstream flows from farm dams will be substantial. There are databases (National Groundwater Database, NGWD; Groundwater Information Project, GRIP; WARMS; WR2005; GRA II, NGA, among others) which provide information on groundwater use in the region, however, unaccounted for groundwater use is widespread and many of the databases report conflicting volumes.

Spatial uncertainty in this type of environment will be high as the mountain ridges with steep topography and the flatter valley floors with deeper soils and colluvial deposits would be represented in one quaternary sub-catchment (in a lumped rainfall-runoff model). In reality, the main mountain ranges dominated by quartzitic sandstone are assumed to have a high recharge potential due to the higher rainfall and thinner soils while the valley floors, dominated by interbedded shales and sandstones as well as overlying colluvial deposits, are assumed to have a lower recharge. It is also expected that the gradient of the local water table will be relatively high in the mountain areas, suggesting a dominance of groundwater movement toward the flat valley bottoms (with low groundwater gradients) where it will emerge as baseflow.

4.5.4. Comparing the water balance components of the groundwater portion

This part of the study entailed a comparison of the water balance components of the Berg WAAS with that of the Pitman model and represents a preliminary analysis using default model parameters. There were two types of discharge calculated during the Berg WAAS project which included natural discharge via springs or contributions to stream flow (groundwater baseflow) and groundwater abstraction. Baseflow data were obtained from the Groundwater Resource Directed Measures database (GRDM) (DWAF, 2006b). These baseflow estimates were disaggregated into aquifer specific values, using assumptions and knowledge about the distribution of discharge sites. Groundwater use data were obtained from the WARMS database (DWA, 2012b), which includes registered borehole use, and from the National Groundwater Database (NGDB) (which has been replaced by the National Groundwater Archive; DWAF, 2012a). There are no estimates available for other forms of groundwater discharge, such as downstream groundwater movement or lateral recharge via hydraulic connections between aquifers, therefore no data on these processes were given.

The Pitman model provides estimates of the above components as well as estimates of the volume of interflow, downstream groundwater outflow and riparian evapotranspiration directly from groundwater. Estimates of the volume of downstream groundwater outflow were derived using topographic data and assumed transmissivity values. The topography in sub-catchments H10A to H10D suggests low downstream hydraulic gradients associated with the flat valley floors. Estimates of the riparian strip parameter (that determines evapotranspiration losses from groundwater) were based upon the average width of the riparian zone interpreted from aerial images and Google Earth. The Pitman model normally uses the GRA II (DWAF, 2005a) database to obtain values for the groundwater parameters in the model if no alternative data are available, while the Berg WAAS has produced an alternative set of input and output figures. Data from both the GRA II database and the Berg WAAS estimates have been incorporated into the Pitman model in a series of model runs designed to assess the Pitman models ability to reproduce the different water balance components as presented by the Berg WAAS (Table 4-12).

Table 4-12 Outline of model runs undertaken using different sets of data.

Model Run	Surface water parameters	Recharge calibrated to against:	Groundwater use	Groundwater baseflow calibrated against:
R1	WR90 recommended parameter values with calibration	GRA II database	None	WR90 naturalised flow data
R2	WR90 recommended parameter values with calibration	GRA II database	Berg WAAS estimate	Patched observed data – flow gauge H1H003
R3	WR90 recommended parameter values with calibration	Berg WAAS estimate	None	WR90 naturalised flow data
R4	WR90 recommended parameter values with calibration	Berg WAAS estimate	Berg WAAS estimate	Patched observed data – flow gauge H1H003
R5	WR90 recommended parameter values with calibration	Berg WAAS estimate	Berg WAAS estimate	Berg WAAS estimate

Model simulation R1 represents the type of model setup usually undertaken when there are no additional data available and uses the existing national datasets (WR90, Midgley *et al*., 1994 and GRA II, DWAF, 2005a). While the default parameters were initially incorporated in the setup, they were calibrated against the naturalised data to obtain an improved simulation. The simulated outputs are compared to the naturalised flow data for H10C in Figure 4-41. Table 4-13 shows the model parameters used in the setup while Table 4-14 gives the statistical measures of the model performance for the entire simulation period (1920 to 1990) for H10C sub-catchment. Clearly the model was able to reproduce conventional wisdom (WR90 naturalised flow; Midgley *et al*., 1994) using the national datasets with a limited calibration effort.

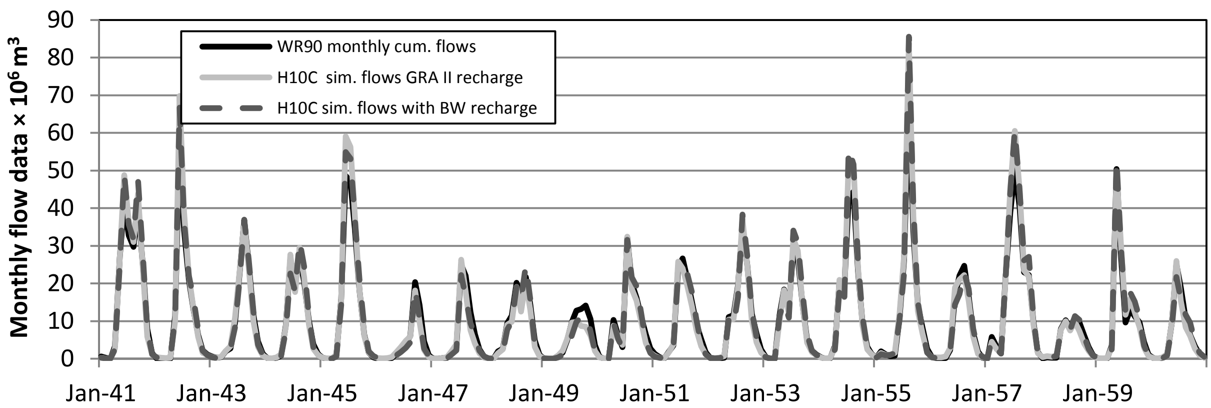


Figure 4-41 Comparison of the results of the Pitman simulation (R1) using GRA II recharge estimates (natural conditions) and the WR90 naturalised flows between 1940 and 1960 for sub-catchment H10C.

Table 4-13 Parameter values for each of the model runs (R2 and R4 parameter sets are the same as R1 and R3 parameters sets respectively but with groundwater use incorporated). The remaining surface water parameters (not included in the table) are the same for all model runs.

	R1			R3			R5		
Parameter	H10A	H10B	H10C	H10A	H10B	H10C	H10A	H10B	H10C
ST	130	130	190	180	200	180	190	190	190
POW	2	2	2	2	2	2	2	2	2
FT	50	40	50	30	30	30	75	75	75
GW	34	16	30	25	33	64	118	65	110
GPOW	3.5	2.9	3	2.9	3	3	2.5	2.5	3
Drainage density	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3
Transmissivity (m^2 d^{-1})	30	30	21	37	44	35	40	40	60
Storativity	0.004	0.004	0.004	0.004	0.004	0.004	0.02	0.02	0.014
Regional GW drainage slope	0.008	0.008	0.009	0.008	0.008	0.008	0.004	0.003	0.003
Rest water level (m below surface)	25	75	75	25	75	75	25	75	75
Riparian Strip Factor (% slope width)	1.2	1.2	1.8	2	2.5	2.5	3.8	4.7	6.4

Table 4-14 Statistical measures of the model performance for H10C sub-catchment for model runs R1 to R5.

	R1	R2	R3	R4	R5
Compared against	WR90 Naturalised flow	Patched observed data	WR90 Naturalised flow	Patched observed data	Patched observed data
Nash Coefficient (Untransformed)	0.94	0.94	0.95	0.83	0.44
Nash Coefficient (Ln transformed)	0.95	-0.84	0.94	0.91	0.60
% Bias in simulated monthly flows	1.99	-10.62	0.27	23.57	70.10
% Bias in simulated monthly ln(flows)	0.48	-107.71	-0.22	36.49	102.40

Model simulation R3 (R2 is discussed later) represents a model setup undertaken using the recharge estimates given in the Berg WAAS. This model run did not include groundwater use and was therefore also compared to naturalised stream flow data (WR90, Midgley *et al*., 1994). While the GRA II database (used in model run R1) provides three recharge values based on different recharge estimation methods, previous experience suggests a value between the lower and middle estimate is most often appropriate. In most of the sub-catchments the Berg WAAS recharge values were far higher and are comparable with the highest GRA II values. Once the recharge values had been increased, the model parameters were recalibrated against the naturalised flow data and achieved similarly good model statistics when compared to model run R1. The results of simulation R3 using the generally higher recharge values reported by the Berg WAAS are illustrated in Figure 4-41, while the parameters used in the setup, along with the statistics of model performance are given in Tables 4-13 and 4-14 respectively.

The results of model simulations R2 (GRA II recharge) and R4 (Berg WAAS recharge) include the same parameter sets used in simulations R1 and R3 but with groundwater use incorporated. These simulations are compared with the patched stream flow data from flow gauge H1H003 since they are no longer representing natural stream flow. The patched stream flow data are not strictly comparable to the simulated flows as no surface water use has been incorporated into the simulations. The stream flow in the catchment is expected to be heavily impacted by a large number of farm dams and return flows (Hughes and Mantel, 2010a). The surface water use (from farm dams) was not included in this part of the study as the focus of the comparative investigation is on the groundwater components of the water balance. While the simulation results have been compared to the observed patched data in H10C for a general comparison, the focus in this study remains on the simulated groundwater balance, hence it was deemed unnecessary to include the highly uncertain surface water use data (this has been included at a later stage as part of an uncertainty assessment). The outputs from simulations R2 and R4 are illustrated in Figure 4-42. The parameter values used in the setup along with the statistics of model performance are given in Tables 4-13 and 4-14 respectively. In addition, the simulated groundwater balance components are shown in Table 4-15 together with the results reported by the Berg WAAS.

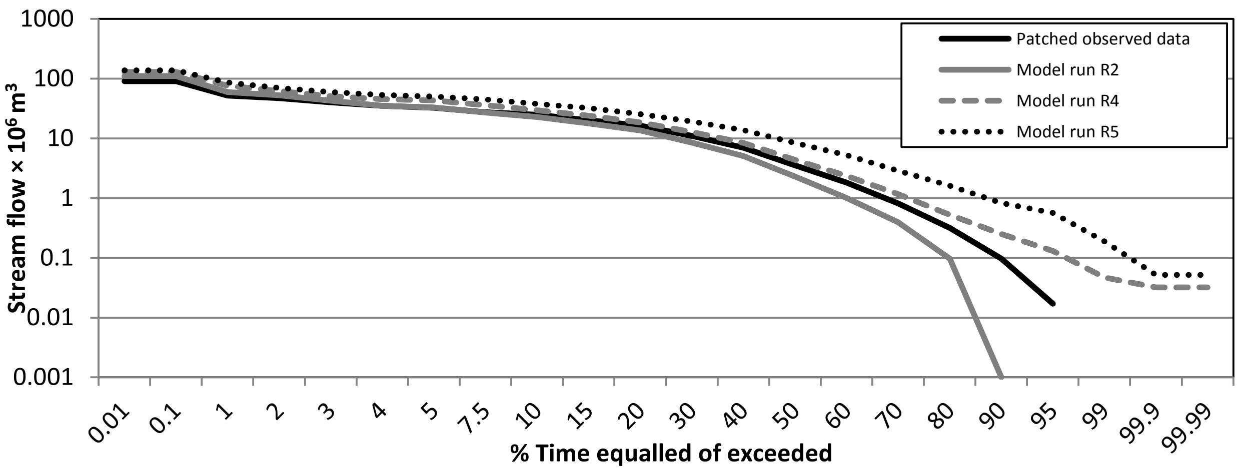


Figure 4-42 Comparison of the model runs with groundwater use included.

Table 4-15 Simulated groundwater balance components (10^6 m^3 y^{-1}).

	Berg WAAS	R2 Pitman	R4 Pitman	Berg WAAS	R2 Pitman	R4 Pitman
	H10A			H10B		
Inputs						
Recharge	12.34	4.90	12.34	13.24	8.19	13.24
GW inflow		0.00	0.00		0.00	0.00
Total		4.90	12.34		8.19	13.24
Outputs						
GW cont. to baseflow	0.75	0.25	7.12	3.53	<0.01	2.43
Evapotranspiration		2.35	2.71		0.69	2.90
GW outflow		0.69	0.90		0.31	0.72
Water use	1.61	1.61	1.61	7.19	7.19	7.19
Total		**4.90**	**12.34**		**8.19**	**13.24**
	H10C			H10D		
Inputs						
Recharge	23.83	10.80	23.84	11.68	9.28	11.68
GW inflow		1.00	1.62		0.14	0.99
Total		11.80	25.46		9.42	12.67
Outputs						
GW cont. to baseflow	5.33	<0.01	2.34	2.05	8.11	10.54
Evapotranspiration		<0.01	5.23		0.26	0.74
GW outflow		<0.01	0.99		0.75	1.09
Water use	16.90	16.90	16.90	0.30	0.30	0.30
Total		**-**	**25.46**		**9.42**	**12.67**

The groundwater use figures estimated by the Berg WAAS are fairly high in comparison with the groundwater use estimates reported in the GRA II database (especially for H10B and H10C) and this had a marked effect on the baseflow volumes. During model run R2, the simulated baseflow figures were lower than those reported by the Berg WAAS for all the sub-catchments except for H10D which has minimal water use. The water use in H10C is clearly not sustainable based on the GRA II recharge data. For model run R4, however, the outputs were substantially different due to the increased recharge. In model run R4, the model generated much higher baseflow volumes in H10A and H10D and lower baseflow results for H10B and H10C (relative to the Berg WAAS). In sub-catchments H10B and H10C, the high recharge is offset by the high groundwater use. However, the Berg WAAS reported baseflow values are highest in these two sub-catchments. Similarly in sub-catchments H10A and H10D where low water use figures did not offset the high recharge input, the Pitman model generates baseflow volumes that are substantially larger than those reported by the Berg WAAS.

The results of the Pitman simulation (R5) which includes the incorporation of both the Berg WAAS recharge estimates as well as the Berg WAAS baseflow values are illustrated in Figure 4-42. The parameter values used in the setup along with the statistics of model performance are given in Tables 4-13 and 4-14 respectively. In addition, the simulated groundwater balance components are shown in Table 4-16 together with the results reported by the Berg WAAS. To decrease the baseflow values in sub-catchments H10A and H10D as well as achieve a water balance, the excess water had to be removed via either downstream groundwater outflow or through evapotranspiration in the riparian margins. The model was set up with a high riparian strip factor as high evapotranspiration was deemed a more likely process than high downstream groundwater outflows, due to the catchment topography which does not support large downstream groundwater gradients. In order to remove (evaporate) enough water to sufficiently reduce the baseflow volumes, the riparian strip had to be increased to 6.7% (H10A) and 4.8% (H10D) of the total catchment area. These values are very high and although riparian evapotranspiration estimates are uncertain, such high values in this area are questionable. While calibration of the remaining components of the model was undertaken to try and achieve the best possible fit to the observed patched data, the comparison was extremely poor and unable to be improved via calibration.

Table 4-16 Simulated groundwater balance components for R5, using Berg WAAS inputs and outputs (10^6 m^3 y^{-1}).

	Berg WAAS	**R5** Pitman	Berg WAAS	**R5** Pitman	Berg WAAS	**R5** Pitman	Berg WAAS	**R5** Pitman
	H10A		H10B		H10C		H10D	
Inputs								
Recharge	12.34	12.33	13.24	13.21	23.83	23.64	11.68	11.72
GW inflow		0.00		0.00		1.38		0.90
Total		12.33		13.21		25.02		12.62
Outputs								
GW cont. to baseflow	0.75	0.77	3.53	3.53	5.33	5.37	2.05	2.02
Evapotranspiration		9.21		1.85		1.85		0.90
GW outflow		0.74		0.64		0.90		9.40
GW use	1.61	1.61	7.19	7.19	16.90	16.90	0.30	0.30
Total		**10.72**		**6.02**		**8.12**		**12.32**

Possible explanations for the discrepancy between the Berg WAAS and Pitman model water balances include regional groundwater flows which remove recharge water from the local surface water system completely. The Berg WAAS assumes that recharge into the Peninsula aquifer (exposed on the mountain tops of H10C and H10D) is discharged into the Hex River (an adjacent catchment to the east) through regional groundwater flow. A portion of the recharge could therefore be moving out of the surface water catchment system completely which would account for the incomplete water balance. The Pitman model formulation does not currently account for this process (Hughes, 2004), although a model component that allows for the movement of groundwater out of the system would be relatively simple to incorporate in the model. This feature has not been developed as yet, largely due to the lack of conceptual and quantitative information on this process in South Africa. One way to replicate the process in the Berg WAAS scenario is to use the downstream groundwater outflow function to remove a large portion of the groundwater from the catchment, thereby replicating the removal of recharge through regional groundwater flows. This method has been demonstrated in R5 for H10D to try to simulate a match with the WAAS water balance. However, at a larger regional scale, the destination of this water would be incorrect. While the possibility of miss-matched surface and sub-surface catchments may provide some explanation for the differences in the baseflow estimations, this is mainly applicable to H10D which comprises large portions of the exposed Peninsula Formation. There are no Peninsula Formation outcrops in catchments H10A or H10B (Figure 4-39). In addition, the Berg WAAS recharge estimates were reported per aquifer type which means that the volume of recharge 'lost' to the Peninsula aquifer was quantified in the study. This volume of recharge into the Peninsula aquifer was not sufficient to reduce the volume of baseflow satisfactorily given the recharge into the remaining aquifers in the catchment.

To summarise:

- The water balance components of both studies were very different and the Pitman model could not achieve a sensible water balance using the values estimated by the Berg WAAS.
- The main issue with the water balance components seems to be high recharge values coupled with low groundwater baseflow values which meant the remaining groundwater discharge pathways had to be set unrealistically high to close the water balance.
    - Two of the sub-catchment water balances (H10B and H10C) were satisfactory as Berg WAAS groundwater use estimates were high enough to sensibly balance the majority of the high recharge estimates.
    - H10A and H10D however, do not have the volume of groundwater use (as estimated by the Berg WAAS) necessary to close the water balance sensibly.

- The attempts to balance the groundwater components with high recharge input values and low baseflow outflow values resulted in conceptually unrealistic results.
- While regional groundwater flows could explain the discrepancy in H10D, there is no exposure (recharge area) of the deep aquifer in H10A. In addition, the volume of estimated recharge (the Berg WAAS reported recharge per aquifer type) into the deep aquifer in H10D is still not sufficient to sensibly close the water balance.

Both studies are subject to very large uncertainties in the data used to quantify the components of the water balance; particularly the volume of groundwater recharge, but it would seem that the Berg WAAS data, even when regional groundwater flows are considered, do not result in a sensible water balance. After all the conceptualisation and modelling undertaken, both numerical and conceptual, much uncertainty remains.

4.5.5. Quantification of the uncertainties

Following the comparison of the water balance components of the Pitman model with those of the Berg WAAS, the uncertainty version of the Pitman model was used to examine the sources of uncertainty in the catchment related to the surface and groundwater interaction processes. In many circumstances it is difficult to fully understand the sources of uncertainty and therefore they cannot be properly quantified. Consequently it can be argued that a substantial contribution to uncertainty reduction can be made through a better quantitative understanding of the different sources of uncertainty and their relative contribution to total uncertainty. To attempt to identify and isolate the dominant sources of uncertainty in the Upper Breede River, the Pitman model was run within an uncertainty framework (Kapangaziwiri and Hughes, 2009) using the parameter estimation routines (Kapangaziwiri and Hughes, 2008) developed specifically for the model. Both the WR2005 (Bailey and Pitman, 2005) and WR90 (Midgley *et al*., 1994) rainfall data were used in the simulations in an attempt to assess the uncertainty in the available rainfall datasets. The methodology followed in the uncertainty assessment is outlined below:

(1) The first step included the generation of an uncertain parameter set (the natural runoff parameters) using the parameter estimation tool (Kapangaziwiri and Hughes, 2008) and the physical basin property data (AGIS, 2007). The uncertainty version of the model was then run using the defined uncertainty in the parameter set and 10 000 ensembles of natural hydrology were generated.

(2) Secondly the GW parameter was increased to generate recharge values closer to the value estimated by the Berg WAAS (DWAF, 2007a) and the 10 000 ensembles rerun.

(3) Thirdly the groundwater slope parameter was increased from 0.01 to 0.05 to generate a larger volume of groundwater outflow which represented the sub-surface inter-basin transfer of water. Although this is not an explicit representation of regional groundwater flows, it represented a discharge pathway that was increased in an attempt to replicate the process of regional flow. While an estimate of the volume of recharge lost through this process was provided in the Berg WAAS (0.3 × m^3 10^6 – H10C) (DWAF, 2007a), this volume had to be increased (to 2.9 m^3 × 10^6) as the value reported by DWAF (2007a) was too small to have any substantial effect on the water balance. The 10 000 ensembles were then rerun.

(4) Data on the water use in the catchment were incorporated into the model and the 10 000 ensembles run for the fourth time. The water use data were calculated by estimating the irrigated areas in the catchment together with the crop types (Hughes and Mantel, 2010b). The water use was assumed to be exclusively surface water use from farm dams in the fourth simulation.

(5) Groundwater use data from both the GRA II database and Berg WAAS were averaged and incorporated into the simulation as a percentage of the total water use (estimated in step 4), before the 10 000 ensembles were run for the fifth time.

(6) Lastly, the model was run using rainfall data from the WR90 database (Midgley *et al*. 1994) (the earlier simulations used WR2005 data) to assess the effect of rainfall uncertainty on the simulated outputs.

The initial uncertain parameter set which represented the natural hydrology of the catchment, generated using the set of parameter estimation equations, is given in Table 4-17. Each stage of the assessment (stages 1 to 6) added further uncertainty on top of the prior stages uncertainty (the prior uncertainty was not removed). The output file produced by the model consists of a text file of all parameter sets (usually 10 000), the summary statistics data (mean monthly runoff, mean monthly recharge, slope of the flow duration curve, the 90th, 50th and the 10th percentiles) and five objective functions which include the coefficient of efficiency for the normal (CE N), natural logarithm transformed (CE L) and inverse values (CE 1/d) and the mean monthly runoff error for both normal (% Diff N) and natural logarithm transformed values (% Diff L). The output from each model run included three time series based on ranking (ascending order) the 10 000 simulated flows for each month of the time series. The “lower” time series represents the 5th percentile of the ranked values, while the "central" and "upper" time series represent the 50th and 95th percentiles, respectively. The lower and upper time series therefore represent the bands covering 90% of all simulated flows.

Table 4-17 Physically based parameter estimates for sub-catchment H10A-C generated for the initial model run (natural hydrology) with a normal distribution*.

Parameters	H10A		H10B		H10C	
	Mean	Standard Deviation	Mean	Standard Deviation	Mean	Standard Deviation
Rain Distribution Factor	1.28	0	1.28	0	1.28	0
Summer intercept cap.(Veg1) PI1s	0.75	0.08	0.65	0.07	0.6	0.1
Winter intercept cap.(Veg1) PI1w	0.75	0.08	0.65	0.07	0.6	0.1
Power of Veg recession curve	0	0	0	0	0	0
Annual Pan Evaporation (mm) PEVAP	1670	0	1650	0	1650	0
Summer min.abs.rate (mm/m^{-1}) ZMINs	64	11	123	11	94	12
Winter min.abs.rate (mm/m^{-1}) ZMINw	64	11	123	11	94	12
Mode abs.rate (mm/m^{-1}) ZAVE	600	0	850	0	774	0
Maximum abs.rate (mm/m^{-1}) ZMAX	1196	26	1158	26	1000	38
Maximum storage capacity ST	167	23	167	21	166	23
No recharge below storage SL	0	0	0	0	0	0
Power : storage-runoff curve POW	2	0.1	1.9	0.1	2	0.1
Runoff rate at ST (mm/m^{-1}) FT	14	4	45	24	36	17
Max. Recharge rate (mm/m^{-1}) GW	15	2	26	2	23	2
Evaporation-storage coefficient R**	0.5	0	0.5	0	0.5	0
Surface runoff time lag (mnths) TL	0.25	0	0.25	0	0.25	0
Channel Loss TLGMax(mm)	0	0	0	0	0	0
Power : Storage-Recharge curve GPOW	3	0	3	0	3	0
Drainage density	0.4	0	0.4	0	0.4	0
Transmissivity (m^2/day)	50	10	50	10	50	10
Storativity	0.001	0.001	0.001	0.001	0.001	0.001
Initial GW drainage slope	0.01	0	0.01	0	0.01	0
Rest water level (m below surface)	25	0	25	0	25	0
Riparian Strip Factor (% slope width)	0.2	0.02	0.2	0.02	0.2	0.02

*Distribution type 1, normal distribution; type 2, log-normal distribution and type 3, uniform distribution.

**Uniform distribution with minimum value of 0.7 and maximum value of 1

Figures 4-43 to 4-48 compare the range of the ensemble flow duration curves (FDCs) for all the runs of the model.

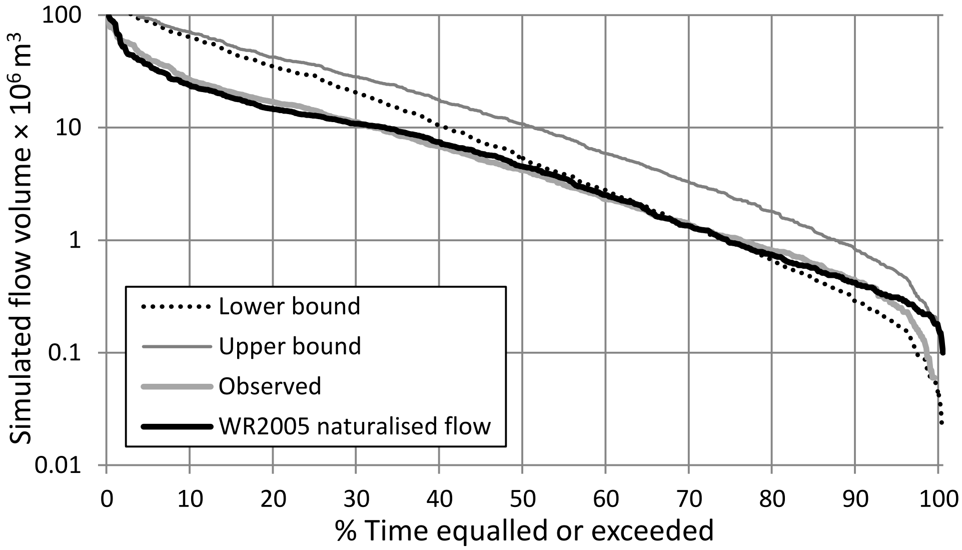


Figure 4-43 Results of the parameter estimation routines for H10C representing the natural hydrology in the catchment (step 1).

Based on the relative spread of the simulated ensembles around the central value, there is more uncertainty in the simulation of natural low flows (90% exceedance) than for moderate to high flows (50% and 10% exceedance). This is largely a consequence of the uncertainty in the drainage and groundwater recharge and discharge parameters of the model. The volume of moderate and high observed flows is far lower than that simulated by the model. This could possibly be due to uncertainty within the model simulation, problems with the rainfall interpolation data used or uncertainties within the measured flow gauge data.

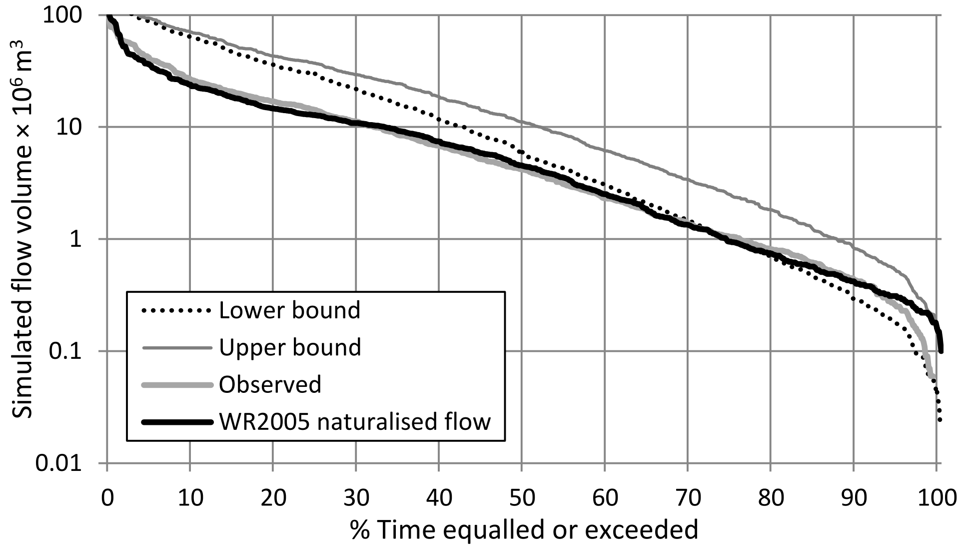


Figure 4-44 Uncertainty estimation for H10C with the incorporation of higher recharge (step 2).

The GW parameter was increased to generate recharge values closer to the value estimated by the Berg WAAS (DWAF, 2007a). The GW and GPOW parameter values used to generate recharge values similar to those from the GRA II database are H10A – 25 and 2.95, H10B – 16 and 2.95, H10C – 30 and 3, while the values used to generate recharge values similar to those from the Berg WAAS are H10A – 118 and 2.5, H10B – 65 and 2.5, H10C – 110 and 3. Increasing the recharge to match the values given in the Berg WAAS (DWAF, 2007a) did not have an appreciable effect on the magnitude of uncertainty. The implication is that the increased recharge reduces the unsaturated zone interflow and compensates.

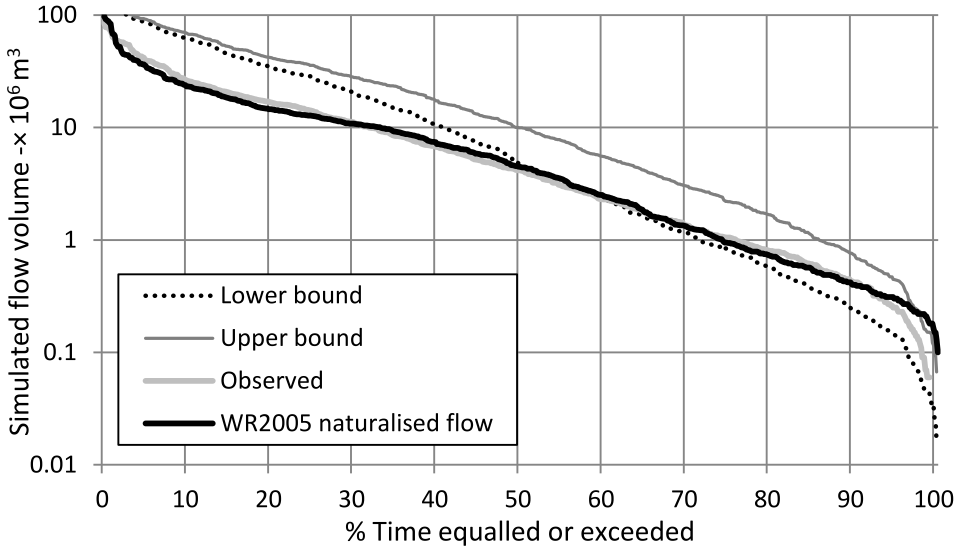


Figure 4-45 Uncertainty estimation with the incorporation of regional groundwater flow from H10C (Step 3).

The movement of excess water downstream (an attempt to replicate regional groundwater flow) did reduce the volume of low flows, however did not have an appreciable effect on the range of uncertainty simulated by the model.

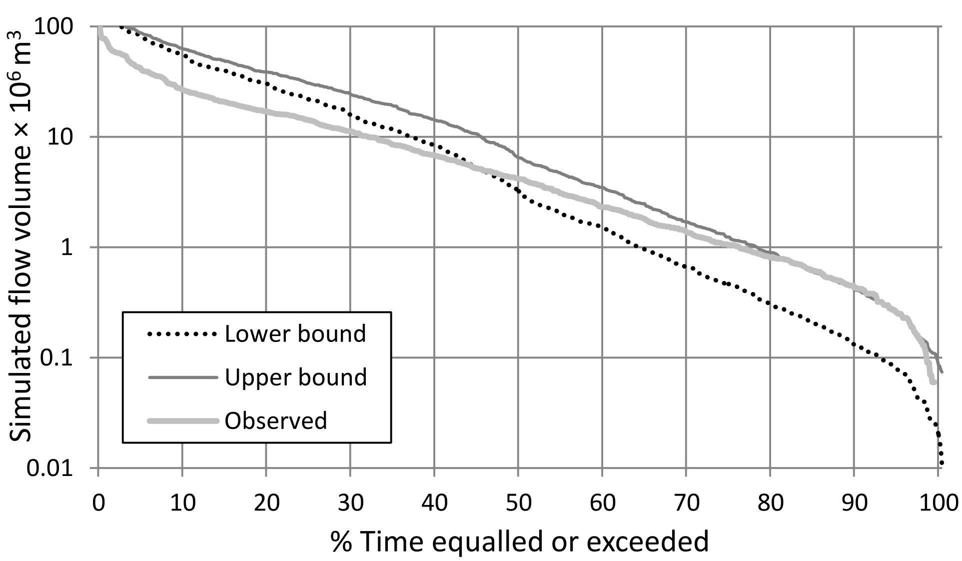


Figure 4-46 Uncertainty estimation for H10C with the incorporation of water use (simulated as surface water use) (Step 4).

The water use in this model run was assumed to be exclusively from the 376 farm dams located in the sub-catchments. Details of the estimation of the farm dam parameters (storage volume, upstream catchment area, water use etc.) in the model are given in Hughes and Mantel (2010b). The irrigated area in the sub-catchments was estimated based on google earth aerial images and estimated to have a mean value of 68km^2.

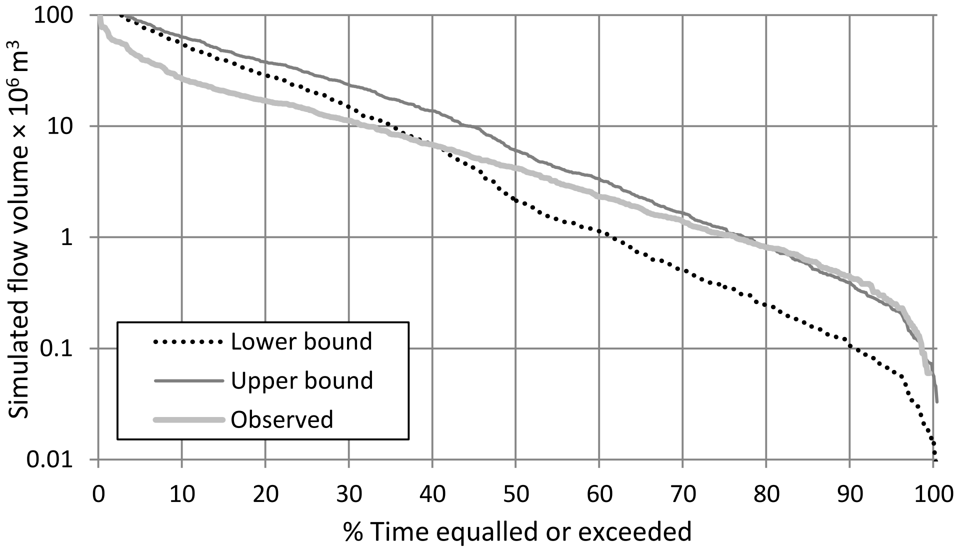


Figure 4-47 **Uncertainty estimation for H10C with the incorporation of water use (simulated as both surface water and groundwater use) (Step 5).**

Incorporating groundwater use into the model simulation clearly lowered the simulated low flows although did not have an appreciable effect on the uncertainty. The simulations were therefore unable to confirm whether the water use is predominantly from surface water or groundwater. The disparity between the simulated and observed data are likely to be partly due to return flows from the town of Ceres which have not been included as part of the model setup. These return flows could increase the volume of the low flows considerably as Ceres town is located close to the outlet of H10C with a population of about 40 000 people. A rough calculation carried out to determine the approximate volume of return flows included the approximate average water use per person – 25 m^3 month^{-1} × % of the water used likely to contribute to return flows (40%) × approximate population of Ceres (40 000) which gives a total of about 4.8 × 10^6 m^3 y^{-1}. In addition, while the simulated flows represent stationary development conditions associated with the fixed parameter values, the observed flow data are expected to be non-stationary, reflecting the history of water resources development (Hughes and Mantel, 2010b).

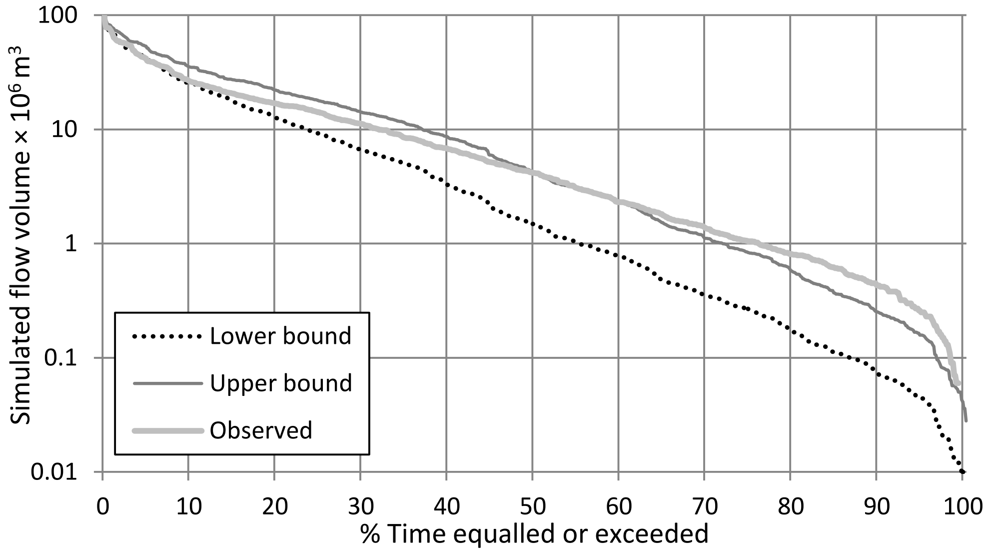


Figure 4-48 Uncertainty estimation with the incorporation of a different rainfall dataset – WR90.

The rainfall figures used in the previous uncertainty model runs were from the WR2005 (Bailey and Pitman 2005) data. The last simulation, however was run with WR90 (Midgley *et al*. 1994) rainfall data (Figure 4-48). This had a marked effect on both the volume of flows and the amount of uncertainty in the simulations. The WR90 data results in a markedly improved simulation of high flows compared to observed data.

Despite attempting to more clearly identify the dominant sources of uncertainty, the complexity of the processes in this catchment have resulted in an unresolved amount of uncertainty from a number of sources. The alternative recharge values, the incorporation of groundwater use and the inclusion of groundwater inter-basin transfers all produced a similar degree of uncertainty in the outputs of the model simulations but with somewhat different deviations from the observed flow records. The largest influence on the uncertainty in this catchment seems to be from rainfall inputs. The highly variable rainfall datasets available in such a topographically heterogeneous environment will clearly affect the outputs from any model. While in this circumstance it has proved difficult to isolate the dominant sources of uncertainty, it is still necessary to identify and evaluate these sources as far as possible. One of the main reasons that these issues are unresolved is that the data available to test various parts of the process are not generally available, despite the fact that the catchment is gauged at the outlet.

4.5.6. Sensitivity analysis

A regionalised sensitivity analysis (Wagener *et al*. 2002; Kapangaziwiri 2010) was also applied to evaluate the parameter sensitivity within the catchments. A sensitivity analysis can identify the parameters and/or parameter combinations that lead to non-behavioural ensembles. The results of the regionalised sensitivity analysis in this study are based on the measure of distribution of the model response that resulted from the 10 000 Monte Carlo input parameter groups sampled. The output ensembles are ranked on the basis of the assessment criteria sorted into five equal groups, then normalised cumulative distribution curves are plotted (Y-axis) for each parameter (X-axis). The sensitivity of the parameter is measured by the degree of divergence of the cumulative curves, i.e. the wide separation of the curves indicates that the parameter is very sensitive based on the assessment criteria considered. Two categories of assessment criteria used in this study are the flow metric (i.e. Mean Monthly Flow (MMF), Mean Monthly Recharge (MMR), slope of the Flow Duration Curve, the 10^{th}, 50^{th} and 90^{th} percentiles of the cumulative frequency distribution of flows) and the objective functions (Tshimanga *et al*. 2011).

The sensitivity tests carried out highlighted the influential and non-influential parameters for the basin. Given the large number of the sensitivity analysis results, only samples of plots are shown to illustrate the trend of parameter sensitivity in the basin (See Table 4-18 and Figure 4-49). There is little doubt that some of the non-influential parameters could be a result of an inappropriate combination of the parameters. The model simulations were very sensitive to the parameters ST and FT in particular based on various assessment criteria used in the sensitivity tests. These parameters represent the soil moisture store and the interflow processes within the model, which suggests that interflow plays an important role in the Ceres Basin. The dominance of the Table Mountain Group in the catchment would support this.

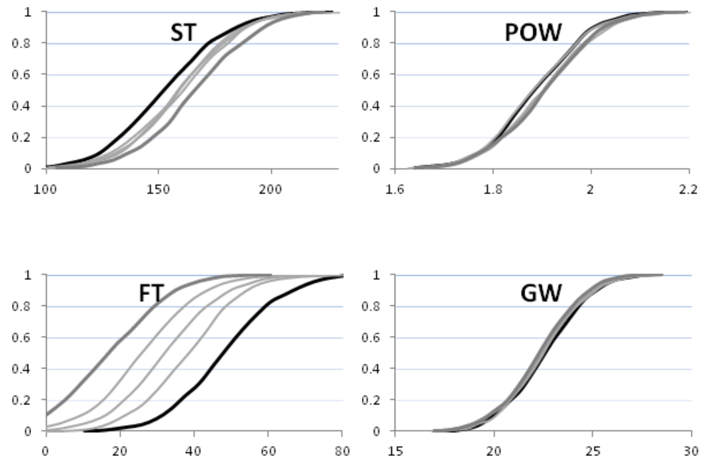


Figure 4-49 **Model parameter sensitivity for selected parameters. The black line, light grey lines and dark grey line represent the top 20%, middle 60% and lower 20% of the ensembles, respectively. All of these parameters have been assessed on the Mean Monthly Flow criteria and represent highly sensitive (FT), moderately sensitive (ST), weakly sensitive (POW) and insensitive (GW) parameters.**

Table 4-18 **Sensitivity analysis results for selected parameters. H represents a high sensitivity, M represents a moderate sensitivity and L represents a low sensitivity.**

	Flow metrics					Objective functions				
Parameter	Mean monthly flow	Slope of FDC	Q10	Q50	Q90	CE N	CE L	CE 1/data	% Diff N	% Diff L
H10A										
ST	H	H		M						
POW										
FT	H	M	H	H						
GW	M	L								
T	H	H	M	M						
H10B										
ST	L	M		L						
POW	L		L	M						
FT	H	H	H	H						
GW										
T										
H10C										
ST	M	M		M	M	L	H	M		
POW	L		L	M		M	H	L	M	
FT	H	H	H	H	H	H	H	H	H	
GW										
T										

4.5.7. Reducing the scale of the simulations

Simulating the sub-catchments as single units represents a further source of uncertainty and the third part of the investigation focused on an alternative modelling approach that has been developed (Hughes *et al*., 2012) where each sub-catchment is split into different zones representing recharge and discharge areas. The work detailed below forms part of the initial investigation published in Hughes *et al*. (2012). The natural hydrology was examined and the results compared to the naturalised stream flow data (Midgley *et al*., 1994). In reality recharge variability is high over a single sub-catchment and recharge volumes on hill tops are assumed to be higher than recharge over the lower slopes and valley bottoms. The alternative modelling approach enabled an assessment of whether the model can simulate the effects of the spatial separation of recharge and discharge zones more clearly. The Upper Breede River is an ideal catchment to examine spatial scale uncertainties due to the steep mountain topography and flat valley floors within a single catchment. Parameter sets for each of the recharge and discharge areas in the sub-divisions of each sub-catchment were generated using the parameter estimation tool (Kapangaziwiri and Hughes, 2008). The natural hydrology was simulated and the assessments of the performance of the parameter estimation routines were therefore based upon estimates of the naturalised observed flows from the WR90 database (Midgley *et al*., 1994). The WR90 rainfall data were utilised as they seem to generate more realistic outputs in the model based on the results from the second part of the study (Figure 4-48).

Table 4-19 summarises the catchment areas in the original sub-catchments and the sub-divisions, which have been based on the variability that exists within the sub-catchment. The sub-division of the rainfall was based on the assumption that the steep mountain slopes will receive higher rainfall than the flat valley bottoms. While the variability of all the physical characteristics were examined in Hughes *et al*. (2012), only the relevant characteristics applicable to surface and groundwater interaction processes are detailed here. These include variations in topography, soil depth and geology as the main mountain ranges dominated by quartzitic sandstone are assumed to have a high recharge potential and the valley floors dominated by interbedded shales and sandstones are assumed to have a lower recharge. H10A-1 is defined as the ridge on the northern boundary of H10A, H10A-2 represents the valley floor, while H10A-3 represents the hills and mountain tops on the eastern boundary of H10A. H10B and H10C have been similarly sub-divided into areas representing the mountain ridges with steep topography and shallow soils and areas representing the flatter valley floors with deeper soils and colluvial deposits.

Table 4-19 Catchment areas and mean annual rainfall (MAP) for the original sub-catchments and the smaller sub-divisions.

	Sub-catchment or sub-division						
	H10A			H10B		H10C	
Area	233.7			162.5		259.6	
MAP (mm)	510.2			704.7		670.7	
	A-1	A-2	A-3	B-1	B-2	C-1	C-2
Area (km^2)	28.0	119.2	86.5	92.6	69.9	116.8	142.8
MAP (mm)	887.7	421.5	510.2	875.2	478.5	849.8	524.2

Modifying the parameter values for the subdivisions relies upon a conceptual interpretation of the differences that occur at smaller scales. For example, higher volumes of recharge will be expected in the mountain areas that have higher rainfall and thinner soils. It is also expected that the gradient of the local water table will be relatively high in the mountain areas, suggesting a dominance of groundwater transfers to the next downstream area rather than re-emergence as river flow within the sub-division. While it is not unrealistic to expect that sound conceptual assumptions can be developed in most situations on the basis of hydrological process understanding, translating these assumptions into appropriate parameter value changes will remain a challenge (Hughes *et al*., 2012). The groundwater parameters are based on several assumptions about the differences between the sub-divisions. The drainage density of channels receiving groundwater drainage is expected to be higher in the valley floor, as is the riparian strip factor. The GW Slope parameter was set relatively high in the topographically steep areas to represent steeper hydraulic gradients assumed to transfer the groundwater to the valley floor zone. The GW Slope parameter in the valley floor zone was set relatively low to reflect the assumption that drainage out of the sub-catchment as a whole will be limited by low gradients.

Figure 4-50 compares the range of the ensemble flow duration curves (FDCs) for the original sub-catchment distribution system with the naturalised flow data (WR90 – Midgley *et al*., 1994) and the observed flow data (flow gauge H1H003). The simulations do not include water use and represent the natural hydrology of the basin, therefore the naturalised flow data are more comparable to the simulated flow ensembles. The expectation is that the observed record would reflect the impacts of water use and return flows in the moderate to low flows relative to the natural flow regime. The uncertain parameter estimation procedures for the lumped scale simulations have clearly generated results that are consistent with the naturalised flow data except in the region of extreme low flows (90% exceedence). However, the high flows of the observed data are not expected to be impacted by water use and return flows, which means the naturalised flow data (together with the lumped scale simulations) could be over-simulating the high flows. Figure 4-51 illustrates the results of the simulations based on the spatial sub-divisions. The reduced scale simulations result in an extended amount of uncertainty in the lower part (low flows) of the flow duration curve but a much reduced amount of uncertainty in the upper part (moderate to high flows). The reduced scale simulations also seem to correlate better with the observed data than the naturalised data. The range of the sub-basin uncertainty, expressed as the difference in the extremes of mean monthly runoff as a percentage of the median, is 31.1%, while this reduces to 18.3% for the sub-division uncertainty. However, there has also been a general decrease in the simulated runoff of approximately 17%, based on the mean monthly runoff volumes of the median ensembles. Comparing Figures 4-50 and 4-51 indicates that this results from lower simulations of moderate flows, while the low flows simulated by the sub-division approach have increased. While there remains some uncertainty about the real patterns of natural flows, there is evidence to suggest that the reduction in spatial scale has improved the low flow simulations, even if the range of uncertainty in the simulations has increased.

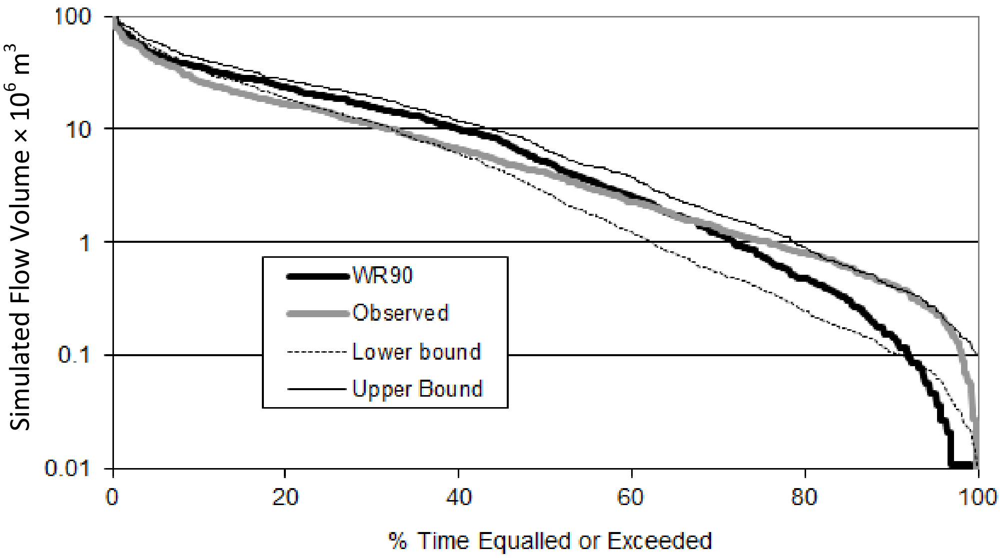


Figure 4-50 **Flow duration curves of the previous simulation of natural flow (WR90), the bounds of the sub-catchment uncertainty ensemble and the naturalised observed flow data at the outlet of H10C.**

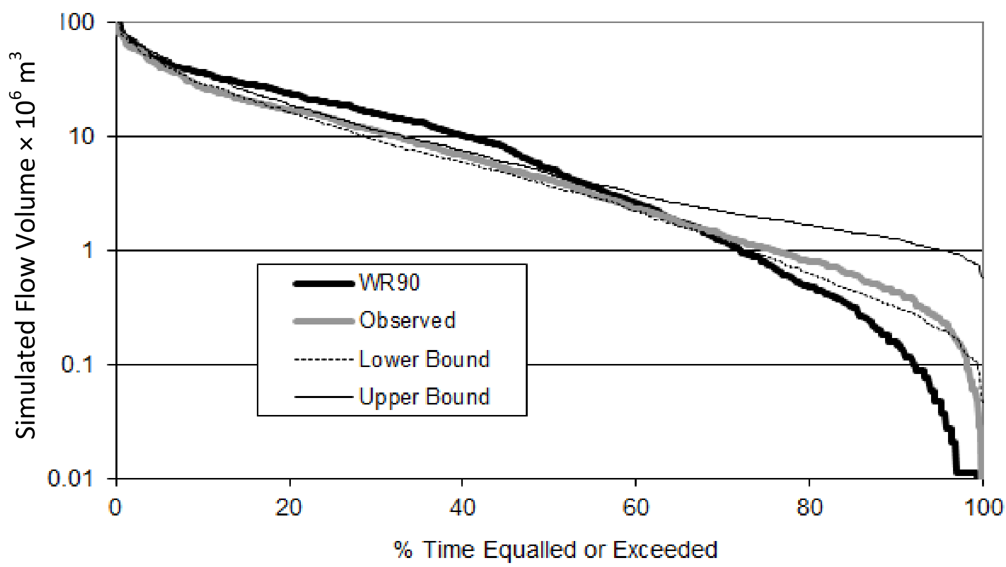


Figure 4-51 **Flow duration curves of the previous simulation of natural flow (WR90), the bounds of the sub-division uncertainty ensembles and the naturalised observed flow data at the outlet of H10C-2.**

There are clear advantages in the reduction of spatial scale. The first is the ability to more explicitly represent spatial climate variations in areas where there are steep topographically controlled gradients in rainfall and evapotranspiration demand (although the latter has not been addressed in this study). The second is that unrealistic combinations of physical property variables (deep soils and steep slopes) are less likely to occur within the parameter estimation equations, leading to more realistic parameter uncertainty ranges. One disadvantage is related to an increase in the level of subjectivity required to establish the groundwater parameters for the reduced scale simulations, given that the existing database (GRA II, DWAF, 2005a) used for the groundwater parameter estimates is based on the surface water catchment scale.

4.5.8. Conclusion

It is not straightforward to assess the results of the Pitman model in this environment. One of the reasons that the uncertainties remain unresolved is that the data available to test the process understanding and quantification of the water balance are not generally available, despite the fact that this area has been the subject of detailed investigations in the past (DWAF, 2005a; DWAF, 2008a). While a thorough conceptualisation of the processes occurring in the sub-catchments assisted in setting the model up, a quantitative assessment of the simulation results is not possible, even though the catchment is gauged at the outlet. While Midgley *et al*. (1994) and other sources of information in the country include 'naturalised' time series of gauged records, there clearly remains un-quantified uncertainty in these results. The detailed conceptualisation did, however, enable the model simulations to explore different 'possible' process descriptions, i.e. different recharge and baseflow estimates. While both the Berg WAAS and Pitman model study are subject to very large uncertainties in the data used to quantify the components of the water balance; particularly the volume of groundwater recharge, it would seem that the Berg WAAS data, even when regional groundwater flows are considered, do not result in a sensible water balance. It is accepted that regional flow could be a dominant process in two of the sub-catchments, and the model's current structure is unable to represent that sufficiently but this process alone could not explain the discrepancy between the two investigations. After all the conceptualisation and modelling undertaken, both numerical and conceptual (modelling), uncertainties remain. The attempt to reduce the spatial uncertainty associated with applying the model in such a heterogeneous environment certainly made sense conceptually, and significantly reduced the uncertainty in the simulations of both the high and moderate flows. The uncertainty in the low flows, however, was increased, although it could be argued they were more behavioural than the lumped simulation. The more explicit representation of recharge and discharge processes led to an improved conceptual representation of the surface and groundwater interaction processes, despite the fact that these are difficult to validate given the lack of data. While the conceptualisation of the processes occurring in the catchment seemed to be successful based on the 'knowledge' available, it was still not possible to obtain a reliable quantitative estimate. Part of the problem lies in the inherent complexity in the catchment coupled with uncertain flow data impacted by numerous farm dams and return flows. This makes it difficult to properly validate any of the model outputs, hence two very different conclusions reached by two different models. It is clear that any modelling undertaken in this type of environment for the purposes of water allocation requires some sort of uncertainty assessment to reduce the risks associated with water management decisions. While the model is able to simulate the different possible conditions, it is still not possible to reject or confirm any of the possible hypotheses. Therefore the model as a means to test hypothesis works in some ways, (the identification of poor water balances) but not in others (differences between low recharge and no outflow or high recharge and some outflow).

4.6. Fractured rock environment: Buffelsjag River

4.6.1. Introduction

The interaction environment in this investigation forms part of the Buffelsjag River which, like the Upper Breede River study area, is located within the Cape Fold Belt region (Figure 4-52). Similarly to the Upper Breede River, the data available in the area are sparse and uncertain largely due to extreme spatial heterogeneity (steep mountain slopes and flat valley floors). This work forms part of a study published in Kapangaziwiri *et al*. (2011) which explored different approaches to resolving the uncertainties in the simulated source of low flows using limited water quantity and quality data.

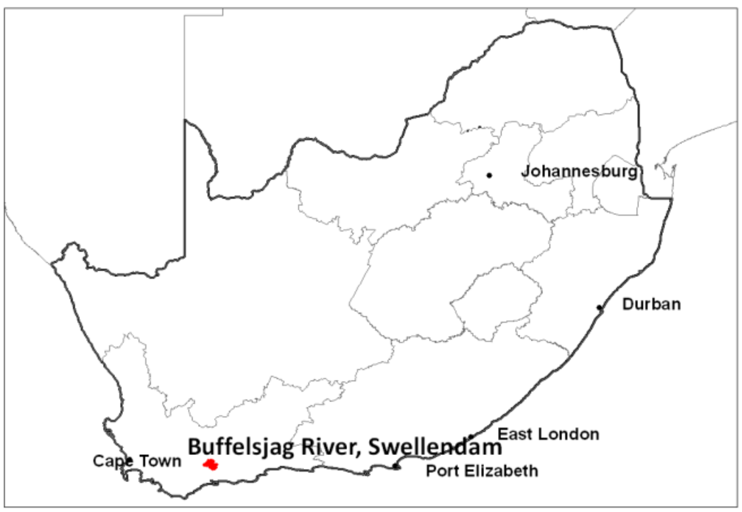


Figure 4-52 Location of Buffelsjag River in South Africa.

4.6.2. Description of the study area

The river drains quaternary sub-catchments H70C and H70D (total area of 457.8 km^2) located within the steep topography of the Cape Fold Belt (Figure 4-53). The study area experiences a typical Mediterranean climate with moderate temperatures and winter rainfall. Mean annual rainfall varies from over 900 mm y^{-1} on the ridge to less than 400 mm y^{-1} to the north due to the orographic influence of the topography, while mean annual potential evaporation is around 1400 to 1600 mm y^{-1} (Midgley *et al*., 1994). The rivers are largely perennial and the baseflows are expected to consist of a large volume of unsaturated zone interflow due to numerous springs and seeps on the steep slopes. The fold system is comprised of deep quartzite and the valley floors comprise poorly fractured shales and sandstones. A ridge of the quartzite termed the Table Mountain Group (TMG) runs east – west separating H70C from H70D, while the remaining areas both the north and south of the ridge are underlain by interbedded shales and sandstones of the marine derived Bokkeveld series (Figure 4-54). There are two gauging stations, one at the basin outlet (H7H003) and the other on a 28km^2 tributary (H7H004).

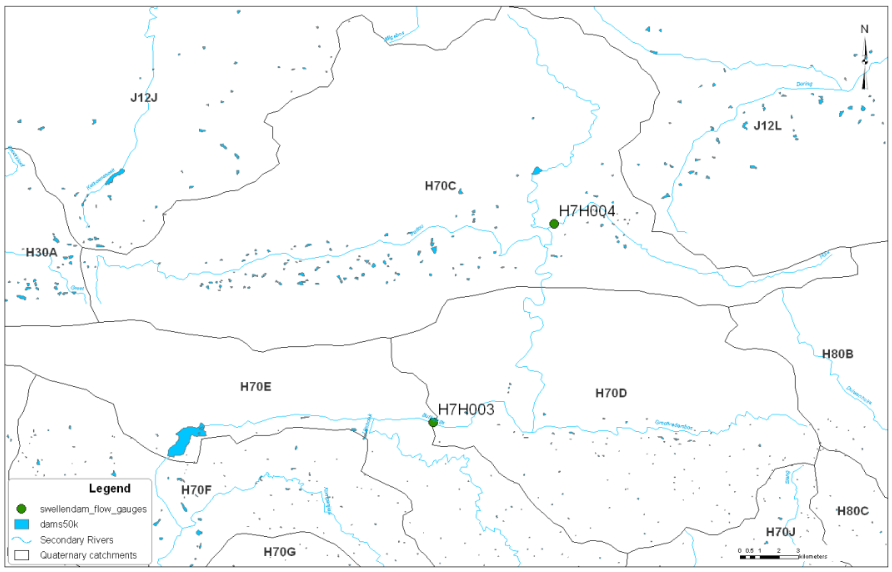


Figure 4-53 **Drainage network of sub-catchments H70C and H70D with location of farm dams.**

Figure 4-54 **Google earth image (43.47 km eye altitude) showing the Buffelsjag River and associated flow gauges (used for water quality and flow data).**

4.6.3. Conceptualisation of the surface and groundwater interaction process

The fold system is the main structural element forming natural boundaries of groundwater and surface water flow. The TMG ridge is assumed to have a higher recharge potential than the valley floors due to the higher rainfall and thinner soils. Low flows are expected to be derived from a combination of groundwater discharges in the main valley bottoms together with springs on the steep rocky hillsides derived from temporally saturated fracture zones.There are numerous farm dams in the valley bottoms located along the base of the ridge (Figure 4-53) which seem to be fed by the river channel or by the interflow emerging from the springs in the quartzite ridge. There are few boreholes in the region largely due to the abundance of surface water. A groundwater level was obtained from a single borehole located near the study site in the valley bottom area (Bokkeveld) and measured 30 m below ground level. The long-term mean annual recharge of this area has been estimated to be between 1.3 to 8% of mean annual rainfall (DWAF, 2005a).

Figure 4-55 shows monthly rainfall and mean daily flow time series for the two observed flow gauges. Borehole water quality data for the region were obtained from the DWA (DWA, 2011). Based on this data, the groundwater in the TMG generally has Total Dissolved Solids (TDS) values of less than 200 mg L^{-1}, while the Bokkeveld shales are substantially more saline (>400 mg L^{-1} and often over 1000 mg L^{-1}).

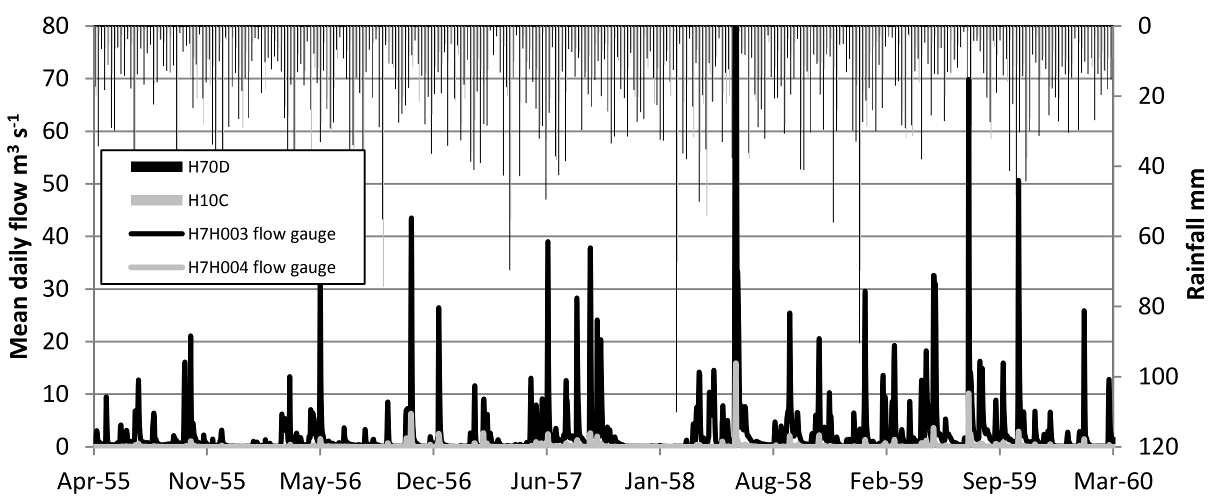


Figure 4-55 Rainfall and mean daily stream flow (gauging stations H7H003 and H7H004) for the Buffelsjag River. H7H004 represents a tributary channel within H70C.

4.6.4. Model setup and results

H70C was split into two areas, one to represent the tributary at H7H004 (lying just to the north of the crest of the ridge) and one to represent the remainder of H70C (in the rain shadow to the north of the ridge). The third sub-basin is H70D which includes the southern ridge slopes and the lower lying area to the south. The modelling was therefore based on three spatial units consisting of one complete sub-catchment (H70D) and one divided sub-catchment (H70C). The Pitman model was established with the regional rainfall and evaporation demand data available in WR90 (Midgley *et al*., 1994), and was calibrated against the observed flow data for 1950 to 1965 to avoid impacts of recent increases in irrigation. Subsequently, the possible combinations of baseflow generating mechanisms were explored to assess whether the model simulations together with the observed flow data could resolve some of the uncertainties in the simulated source of low flows. Figure 4-56 shows the results of three simulations compared with the observed data. The relevant parameters associated with each of the model runs together with the model performance statistics are given in Table 4-20.

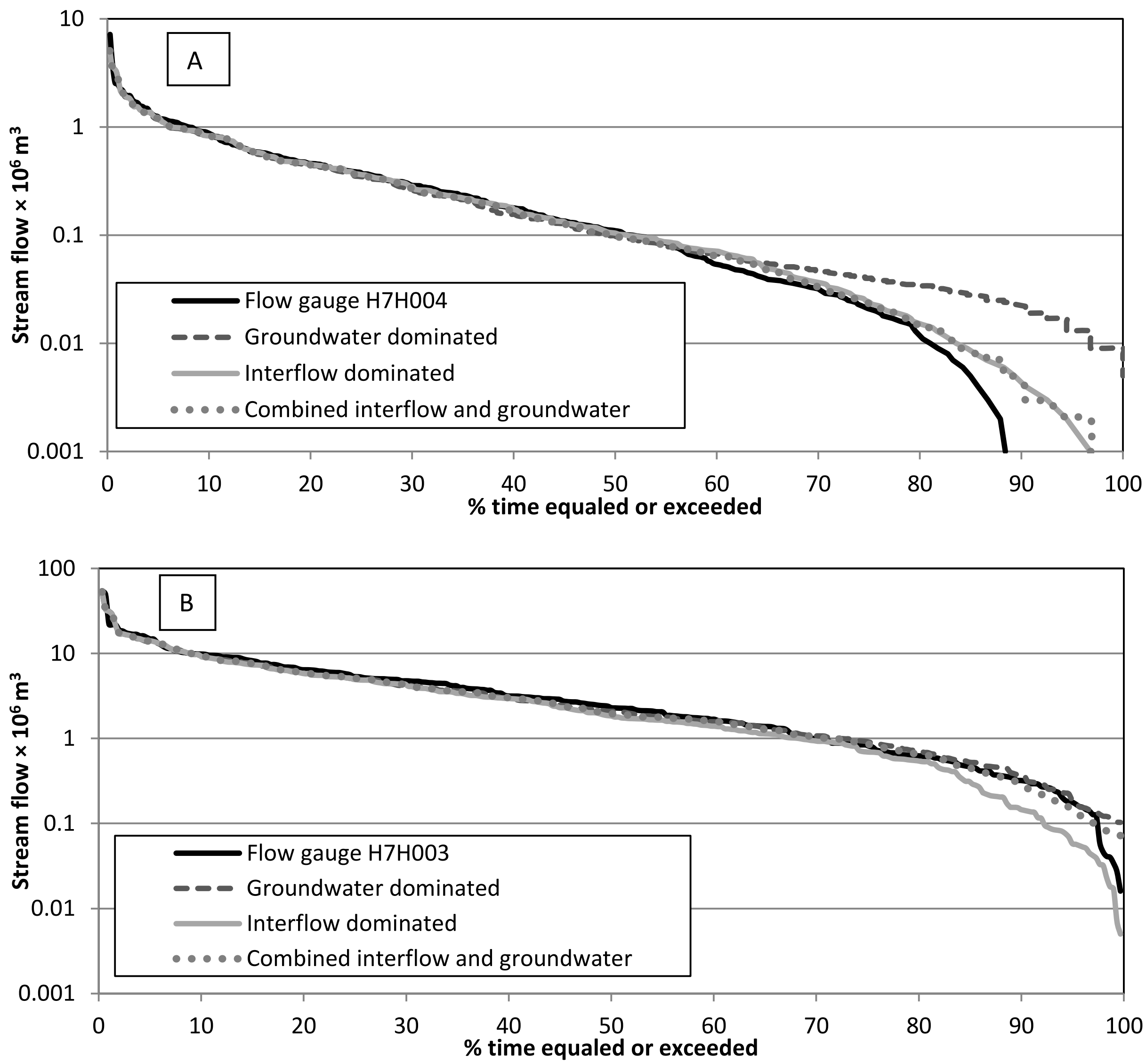
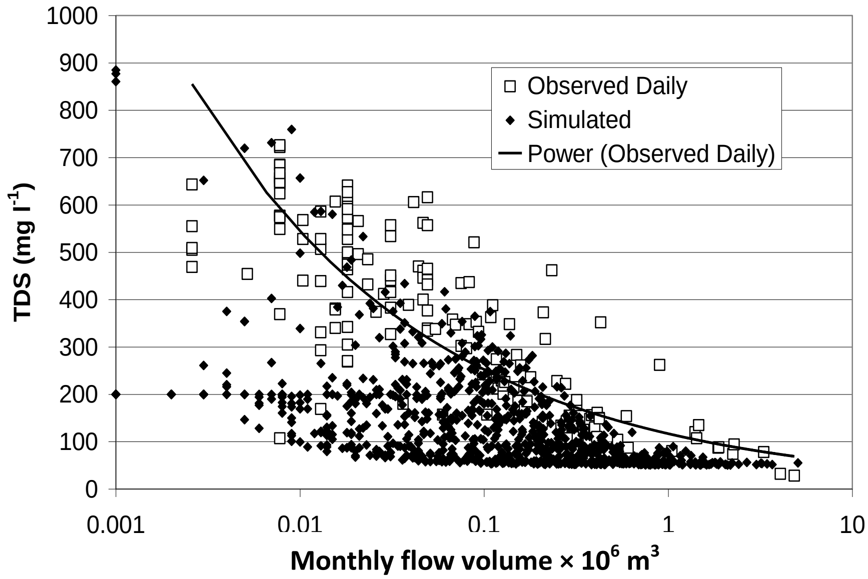


Figure 4-56 Comparisons of combinations of runoff generating mechanisms for H7H004 (Node 1) (A) and H7H003 (Node 3) (B).

Table 4-20 Parameter values and associated model performance statistics.

	Node 1 (H7H004)			Node 3 (H7H003)		
Parameter values	Ground water dominated	Interflow dominated	Combination	Ground water dominated	Interflow dominated	Combination
GW	35	0	12	35	0	12
FT	2	15	12	2	18	12
Model performance						
Nash Coefficient (Untransformed)	0.225	0.241	0.236	0.548	0.533	0.543
Nash Coefficient (Ln transformed)	0.212	0.326	0.207	0.368	0.146	0.337
% Bias in simulated monthly flows	-8.4	-8.0	-9.4	-6.0	-9.0	-6.2
% Bias in simulated monthly ln(flows)	-6.8	6.1	7.7	0.5	-34.3	-6.8

While the comparisons varied slightly both visually and in the performance statistics, there is no obvious combination that clearly points toward a likely dominant baseflow generating mechanism. However, both the performance statistics and visual analysis seem to weakly indicate that Node 1 is interflow dominated while Node 3 is groundwater dominated. The flow vs. Total Dissolved Solids (TDS) relationships for both gauging sites together with the simulated results based on applying TDS signatures to the three modelled runoff components (Table 4-21) were thus examined and the results are illustrated in Figure 4-57. Given that the observed relationships are based on daily flows it is expected that the simulated TDS values would be more scattered and generally lower than the daily values for the equivalent flow volume as the monthly simulations contain a combination of runoff events and low flows. The data for H7H004 shows a strong power relationship indicating a dominant poor quality source diluted during higher flows with a better quality source. The H7H003 data suggest a greater mixture of (and lower TDS) water quality signals, consistent with its more downstream position and the dominance of water derived from the TMG rather than Bokkeveld shale formations.



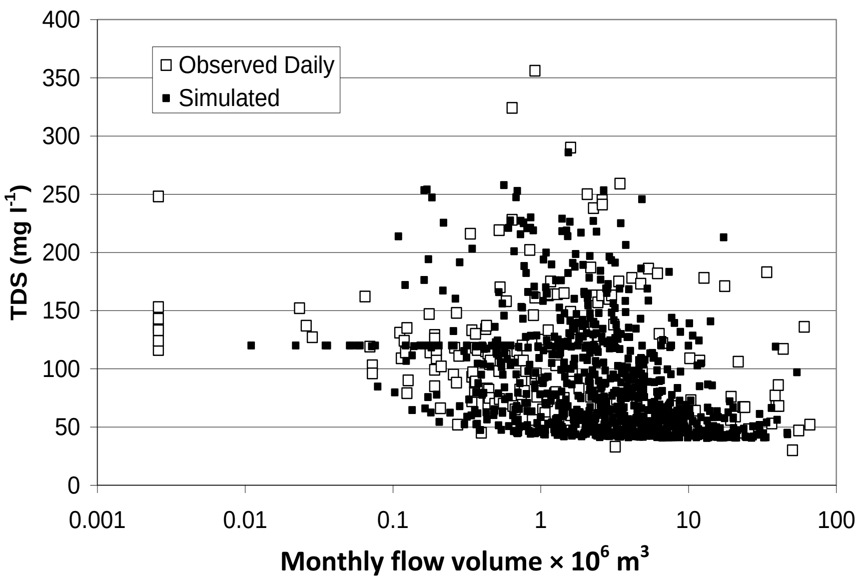


Figure 4-57 Comparison of observed (daily; converted to equivalent monthly volumes) with simulated (monthly) flow volume v TDS relationships for H7H004 (top) and H7H003 (bottom).

The simulated runoff components given in Table 4-21 have been based on manual calibration to achieve the best possible fits to the observed flow duration curves and Figure 4-57 (together with the very simplified approach to estimating simulated TDS) suggests that these model outputs can approximately account for TDS variations. However, it was also possible to achieve almost equally good simulations without the groundwater runoff component. Replacing the interflow with groundwater as the main source of low flows produced poorer flow simulations and could not match the observed TDS variations regardless of the TDS signature used. While the evidence is far from conclusive and there are low flow quantity and quality processes that have been neglected (such as pool evaporation and some anthropogenic impacts), the balance of evidence suggests that the low flows are mainly derived from interflow in saturated fractures above the general water table level but with some contributions from groundwater.

Table 4-21 Simulated runoff components and water quality signatures.

Sub-basin	Runoff component	Surface	Interflow	Groundwater
H7H004	TDS signature (mg l^{-1})	50	200	900
	Simulated % of total runoff	89.2	9.6	1.2
H7H003	TDS signature (mg l^{-1})	40	120	400
	Simulated % of total runoff	82.5	14.4	3.1

Subsequent to the modelling exercise, a one off sampling event in the Swellendam area was undertaken in April 2011. Isotope samples were collected from springs, rivers and one borehole (there were very few boreholes in the area) (Figure 4-58) and the results of the isotope analysis are illustrated in Figure 4-59.

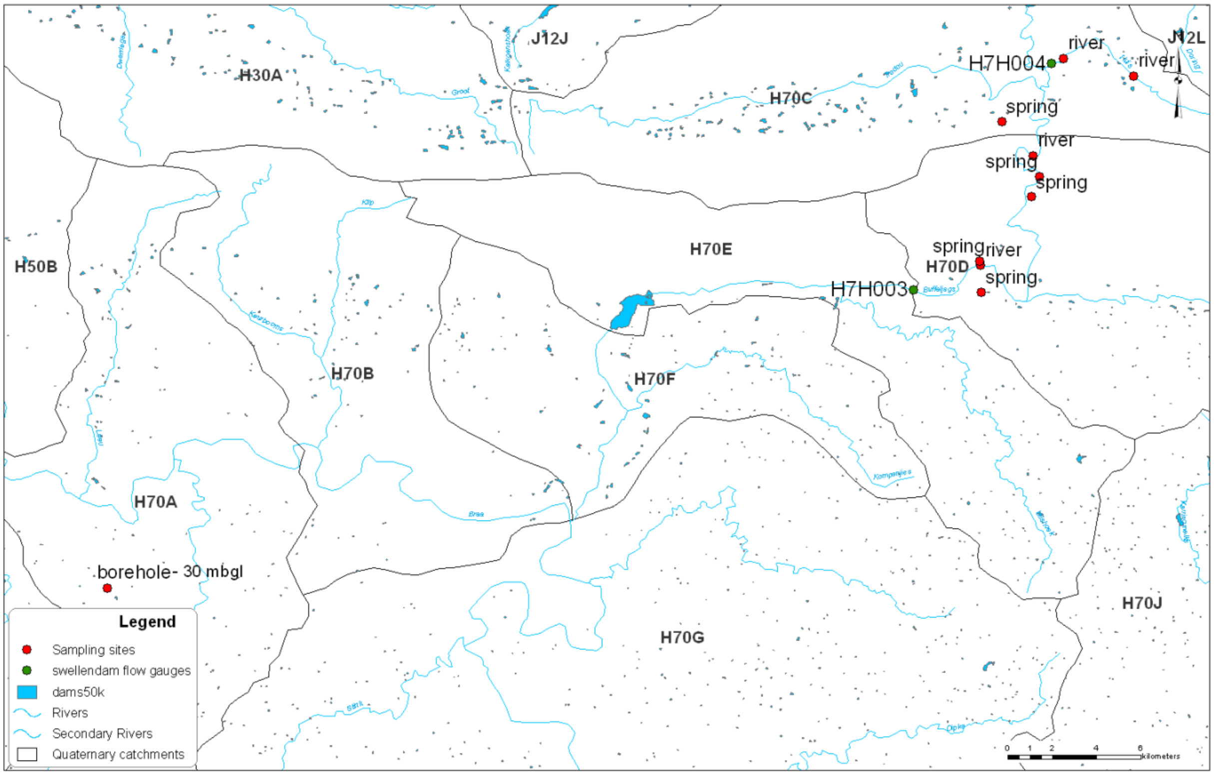


Figure 4-58 **Location of sampling points along the Buffelsjag River.**

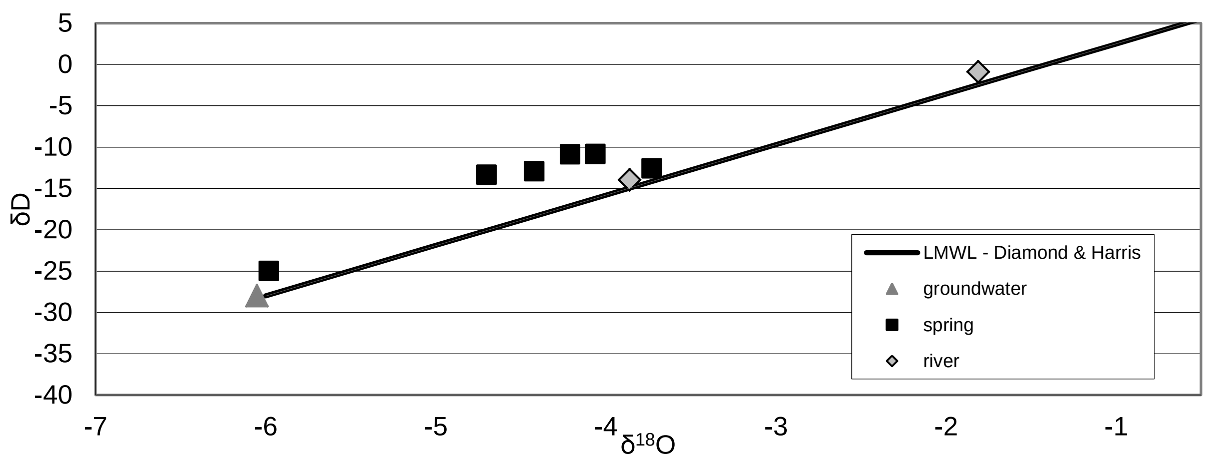


Figure 4-59 **Plot of δD versus $\delta^{18}O$ for isotope samples taken from boreholes, springs and rivers in the Swellendam area. The local meteoric water line was obtained from a study in the Western Cape (Diamond and Harris, 1997).**

From a field survey carried out during the sampling event it was apparent that the majority of the springs in the area are derived from shallow sub-surface flow (originating from marsh like areas on the steep slopes of the catchment). While no rainfall samples were obtained, a local meteoric water line from a study carried out in the Western Cape (Diamond and Harris, 1997) was used to give a general indication of the rainfall values. The linear slopes of both the river samples and the spring samples are similar to the LMWL indicating that rainfall is a dominant long term source of both the river and the springs. Without actual local rainfall samples, however, this cannot be confirmed. The relatively low value of the groundwater sample suggests the groundwater is depleted in ^{18}O due to deeper infiltrating water with a long transit time which allows for greater levels of mixing of water from different events.

4.6.5. Conclusion

While the evidence from the analyses (hydrological, water quality and isotope) was far from certain, it supports the conclusion that the low flows are mainly derived from interflow in saturated fractures above the general water table (consistently fed by small wetland type areas) but with some contributions from groundwater, although the volume of groundwater contribution is very uncertain. While results from the one-off sampling event supported the conclusions reached during the modelling study, further field information such as additional isotope samples and chemical analysis could contribute to an improved interpretation of the processes occurring in the catchment.

There are many catchments in the regions of steep topography associated with the Cape Fold Belt and a better understanding of the water quantity and quality dynamics of sub-surface water (both 'real' groundwater as well as flow in fractures above the regional groundwater table) in the different strata (mainly TMG and Bokkeveld) could potentially contribute to the management of water resources in these areas.

4.7. Fractured rock environment: Elands River

4.7.1. Introduction

The Elands River is located in the North-West Province of South Africa (Figure 4-60). The study area is complex largely due to the extensive development of the region which includes large urban areas (Rustenburg), extensive platinum and chrome mining, agriculture, as well as large dams. There are a number of flow gauges located in the study area (Figure 4-61), however complex and extensive water use together with return flows suggest that the flow data are not representative of natural hydrological responses. A groundwater study was carried out used a simple groundwater yield model (Vivier *et al*., 2007) developed to assess groundwater availability and resource management on a macro catchment scale. The groundwater yield model was used to determine the groundwater balances and volumes available for abstraction in the region. The Pitman model was set up and run for comparative purposes. The study area is large (6215 km^2) which prevents a detailed conceptualisation of the surface and groundwater interaction environments, however, there are relatively detailed data on the water use in the form of a water reconciliation strategy study, (DWAF, 2008c and DWAF, 2008d) available. Given the uncertain flow gauge data, it was found to be very difficult to compare and validate the results of both modelling exercises, a common situation in the predominantly data scarce areas of the country.

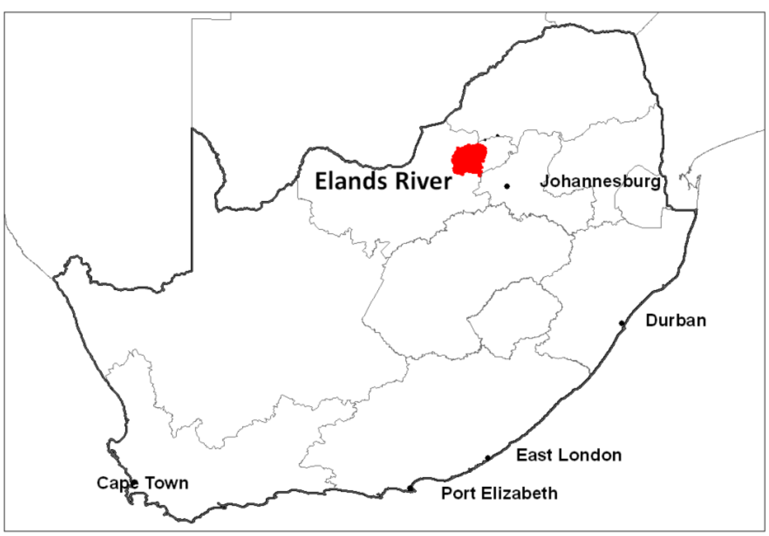


Figure 4-60 Location of the Elands river in South Africa.

4.7.2. Description of the study area

The Elands River is sub-divided into nine sub-catchments designated A22A to A22J (Figure 4-61). The MAP ranges between 500 and 700 mm y^{-1} over the entire catchment area. The Mean Annual Evaporation varies from 1700 mm to 1800 mm over the majority of the catchment although increases to between 1800 and 2000 mm over the west of the catchment. The region is underlain by fractured/weathered hard rock aquifers consisting of rocks from the Transvaal Supergroup (quartzite, shale and dolomite) and the Bushveld Igneous Complex (gabbro, norite and granite) with major aquifers formed by alluvium and regional fault zone structures.

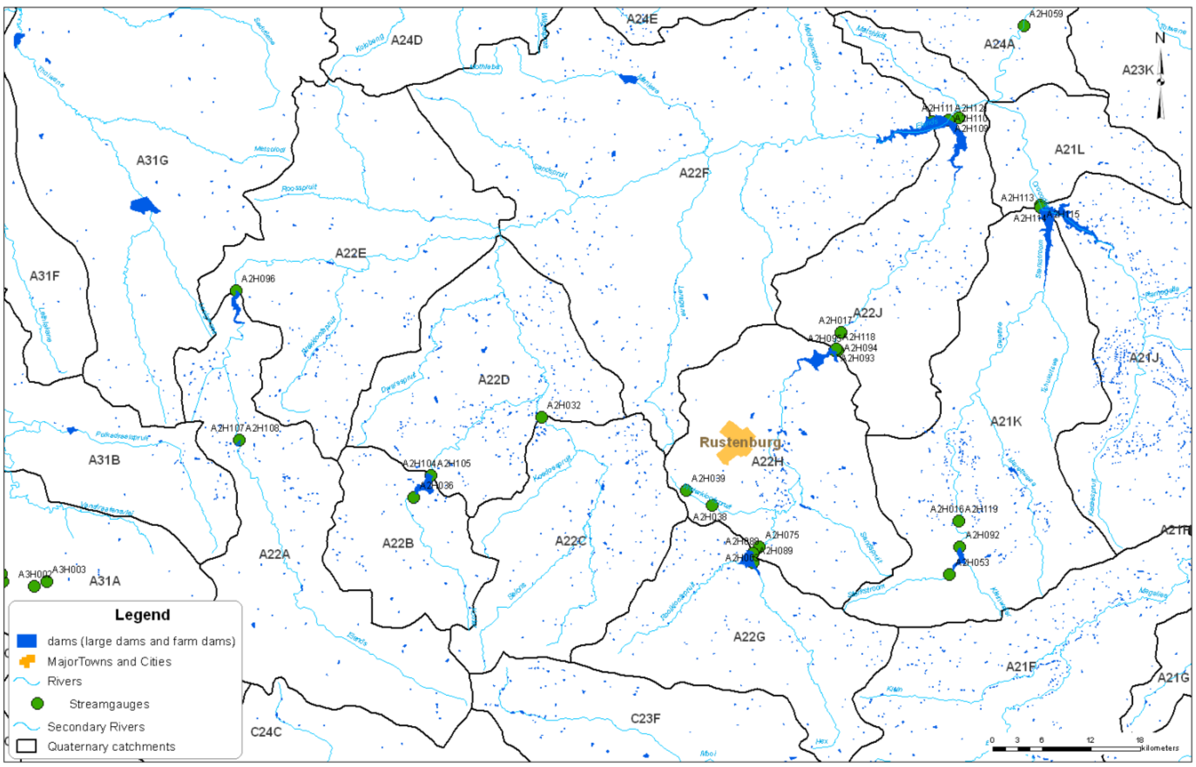


Figure 4-61 Quaternary catchments A22A to A22J (Elands River) with large and small (farm) dams shown as well as the location of the city of Rustenburg.

4.7.3. Conceptualisation of the surface and groundwater interaction processes

In an investigation carried out at this scale (6215 km^2) it is very difficult if not impossible to conceptualise the interaction processes in any detail. Recharge for the region has been estimated at between 1 and 11 % of the MAP (average recharge over the entire area) by the GRA II database (DWAF, 2005a). Vivier *et al.* (2007) determined groundwater recharge from regional hydro-geological maps as well as data from site specific aquifer investigations and estimated recharge to be between 2% and 3% for the hard rock fractured aquifers. The groundwater study (Vivier *et al*., 2007) found that significant losses occur in the evapotranspiration zone along surface water drainage channels. The groundwater modelling study indicated that evapotranspiration losses were the biggest outflow component and that most of the quaternary catchments had negative groundwater balances (transmission losses).

4.7.4. Model setup and results

There were several databases with information on the study area that were compared during the model setup. Vivier *et al*. (2007) determined groundwater flow balances based on borehole information obtained from the National Groundwater Database (DWAF, 2012a) and the Groundwater Information Project (GRIP, 2012). The data utilised in the Pitman Model simulations were obtained from two reports from a water reconciliation project entitled, 'Assessment of water availability in the Crocodile (West) River catchment' undertaken by BKS (Pty) Ltd and Arcus Gibb (Pty) Ltd. (DWAF, 2008c and DWAF, 2008d). The project used the WARMS database (DWA, 2012b) and aerial imagery to determine irrigation areas and the sources of the irrigation water. While the water availability reports provide a relatively detailed account of water use in the catchment, and there are flow gauges, the volume and complexity of unaccounted for and poorly documented upstream human interferences, mean it is not straightforward to characterise the water use and its effect on the surface and groundwater interactions. Examples of the conflicting groundwater use estimates within the available databases are given in Table 4-22.

Table 4-22 **Groundwater use data for A22 from three different sources (10^6 m^3 y^{-1}).**

Quaternary catchment	DWAF (2008d) data	Vivier *et al*. (2007) data	WR2005 (Bailey, 2007) data
A22A	0.494	0.000	0.088
A22B	0.234	0.000	0.017
A22C	0.354	0.000	0.000
A22D	0.223	0.000	0.077
A22E	0.469	1.000	0.077
A22F	1.738	5.000	0.003
A22G	0.263	0.000	0.036
A22H	0.748	4.000	0.153
A22J	0.140	1.000	0.000

The Pitman Model was established with standard inputs (rainfall and evaporation demand) based on the regional information available in WR2005 (Bailey, 2007). Once the model was set up (without water use), the outflow was calibrated against the WR90 naturalised flow data (Midgley *et al*., 1994). Water use information was then incorporated which included irrigation from large dams, farm dams, groundwater, as well as irrigation water extracted directly from the river. In addition domestic water use and return flows were incorporated (including a transfer of water into Rustenburg from the Vaal River). The groundwater use data utilised in the model was obtained from DWAF (2008c) and DWAF (2008d) as it appeared to contain the most comprehensive and up to date data of all the available datasets. The parameter sets for sub-catchments A22A to A22J which includes all water use are given in Table 4-23 below.

Table 4-23 Parameter values for A22A to A22J.

Parameters	A22A	A22B	A22C	A22D	A22E	A22F	A22G	A22H	A22J
ZMIN (mm m^{-1})	70	70	90	80	60	70	75	85	85
ZMAX (mm m^{-1})	750	1000	900	1000	900	1000	1000	1000	1000
ST	150	135	250	250	250	250	230	250	250
POW	2	3	3	3	3	3	3	3	3
FT	2	7	5	0	1	0	2	0	0
GW	6	11	12	22	27	18	19	18	13
TLGMax(mm)	0.0	0.0	0.1	0.2	0.2	0.0	0.0	0.0	0.0
GPOW	2	3	3	3	4	4	4	4	4
Drainage density	0.35	0.2	0.3	0.4	0.4	0.4	0.25	0.4	0.4
Transmissivity (m^2 d^{-1})	15	12	13	20	11	5	8	20	12
Storativity	0.003	0.004	0.005	0.004	0.001	0.001	0.004	0.001	0.001
Regional GW drainage slope	0.011	0.011	0.010	0.012	0.011	0.010	0.010	0.012	0.011
Rest water level (m)	40	25	25	25	25	10	10	10	40
Riparian Strip Factor (% slope width)	1.05	0.87	0.95	1.1	1.2	1.2	0.8	1.2	1.2
Water use parameters									
Irrig.area (km^2) AIRR	0.0	0.0	0.0	1.3	9.5	0.0	0.0	3.6	0.0
Irrig. return flow fraction IWR	0.123	0.136	0.126	0.119	0.120	0.112	0.128	0.117	0.125
Effective Rainfall fraction	0.64	0.64	0.64	0.64	0.66	0.64	0.64	0.63	0.64
Non-Irrig. Direct Demand (Ml y^{-1})	0	0	0	0	0	0	0	8420	0
Maximum dam storage (Ml)	2470	840	1090	2090	760	3050	470	1310	2010
% Catchment area above dams	80	70	60	70	80	80	80	50	70
A in area volume relationship	52	0.2	39	52	4	73	31	60	38
B in area volume relationship	0.7	1.1	0.7	0.7	0.9	0.7	0.7	0.7	0.7
Irrig. Area from Dams (km^2)	0.25	1.10	0.60	0.72	0.20	0.39	0.40	1.02	0.00
GW Abstraction (Upper slopes-Ml y^{-1})	296	140	212	133	281	1043	158	449	84
GW Abstraction (Lower slopes-Ml y^{-1})	197	93	141	89	188	695	105	299	56
Reservoir sub-model parameters									
Reservoir Capacity ($m^3 \times 10^6$)	14.8	13.2	0.0	0.0	0.0	0.0	13.6	18.2	56
Dead Storage (% Capacity)	10	10	0	0	0	0	10	10	10
Initial Storage (% Capacity)	70	70	0	0	0	0	70	70	70
A in Area(m^2) = A × Vol. $(m^3)^B$	20	27	0	0	0	0	26	33	39
B in Area(m^2) = A × Vol. $(m^3)^B$	0.7	0.7	0.0	0.0	0.0	0.0	0.7	0.7	0.7
Annual Abstraction ($m^3 \times 10^6$)	0.5	0.7	0.0	0.0	0.0	0.0	0.0	0.2	24.7
Annual Compensation Flow ($m^3 \times 10^6$)	0.056	0.007	0.000	0.000	0.000	0.000	0.020	0.000	0.000
Transfer in (10^6 m^3 m^{-1})	0.0	0	0.0	0.0	0.0	0.0	0.0	0.89	0.0

Figure 4-62 illustrates the comparisons between the simulated and naturalised or observed flow data for sub-catchment A22C, for both the natural hydrology and with the inclusion of water use. Sub-catchment A22C represents the only catchment with observed data and no large dam and illustrates the Pitman models ability to satisfactorily simulate both the natural and artificial hydrology. The Nash Coefficient based on untransformed (CE) and natural logarithmic transformed data (CE{ln}) for the natural hydrology were 0.917 and 0.746 respectively (with -7% and 7% difference between the naturalised and simulated data) and the objective functions associated with the simulations which incorporated water use are given in Table 4-24. While the objective functions associated with the latter simulations are poor, there remains unaccounted for water use and return flows in the sub-catchment and the hydrograph in Figure 4-62 b illustrates an acceptable visual correlation between the observed flow data and the simulated flow.

Table 4-24 Model performance statistics for sub-catchments A22A, A22C and A22H (simulations including water use).

Model performance	A22A	A22C	A22H
Nash Coefficient (Untransformed)	0.415	0.062	0.658
Nash Coefficient (Ln transformed)	0.301	0.162	0.147
% Bias in simulated monthly flows	-2.812	15.562	-16.916
% Bias in simulated monthly ln(flows)	-2.602	3.216	-80.413

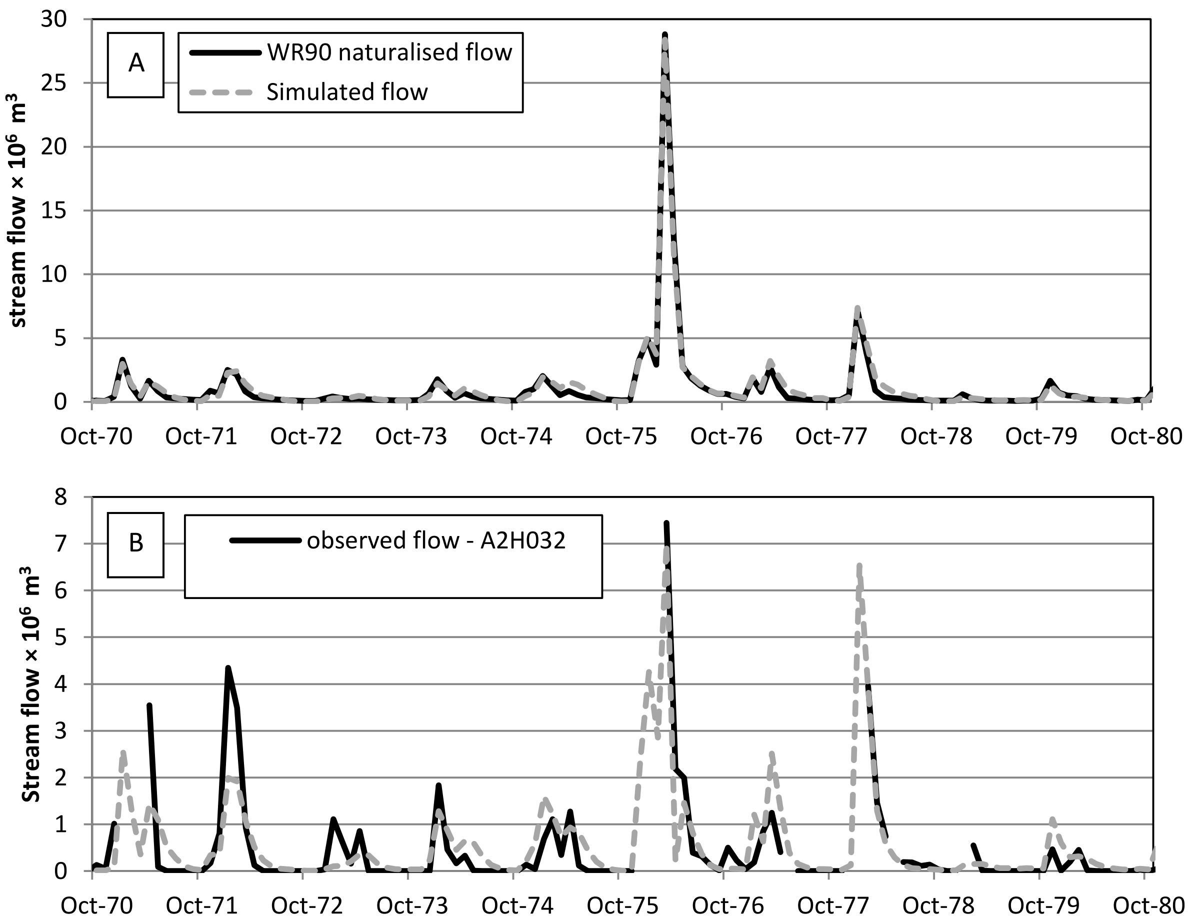


Figure 4-62 Comparisons of the naturalised flow data (Midgley *et al*., 1994) with the simulated flow data for the natural hydrology (A), and of the observed flow data and the simulated flow data with the incorporation of water use (B) for sub-catchment A22C.

Table 4-25 includes a summary of the simulated water balance components of both the groundwater yield model and the Pitman model. Comparisons of the groundwater baseflow values from both studies were quite different. Although some of the input values differed between the two models, these differences were not large enough to account for the disparity between the outputs of the models. Vivier *et al*. (2007) utilised a slightly higher recharge value (approximately 2% of MAP) then the Pitman Model (approximately 1.4 % of MAP) although the recharge estimates both fell within the recommended GRA II range. In addition, the groundwater use data incorporated into the models were from different sources (Table 4-22).

Table 4-25 Summary of the simulated water balance components (10^6 m^3 y^{-1}).

	Area	Recharge		GW use		GW baseflow		Channel losses		Evaporative losses	
	Km^2	Vivier	Pitman	Vivier	Pitman	Vivier	Pitman	Vivier	Pitman	Vivier	Pitman
A22A	705	9.0	8.6	0.0	0.5	0.0	0.1	10	0.0	18	8.0
A22B	284	4.0	3.6	0.0	0.2	0.0	0.1	3	0.0	7	2.8
A22C	515	7.0	6,5	0.0	0.4	0.0	0.4	5	0.0	11	5.1
A22D	541	11.0	9.5	0.0	0.2	0.0	1.5	2	0.1	13	6.7
A22E	812	16.0	9.9	1.0	0.5	0.0	2.4	4	0.1	19	7.4
A22F	1688	32.0	18.0	5.0	1.7	0.0	3.6	15	0.0	43	15.3
A22G	499	10.0	7.3	0.0	0.3	0.0	1.5	3	0.0	12	3.5
A22H	579	15.0	8.0	4.0	0.7	0.0	1.1	0	0.0	10	4.0
A22J	591	8.0	5.0	1.0	0.1	0.0	0.5	21	0.0	28	5.3

The disparity in the model outputs seems to have been caused by different assumptions regarding the drainage density within the catchment. Although both models assumed a similar riparian strip width (15 m), Vivier's model simulated far higher evapotranspiration volumes than the Pitman Model. The Pitman Model assumes that groundwater only contributes to stream flow within the main channel, whereas Vivier *et al*. (2007) assumes a much higher drainage density (i.e. length of riparian strip). Vivier *et al*. (2007) state that, "In areas like the Crocodile River West catchment, where the MAE can be more than double the MAP, the losses may have the effect that there is no visible baseflow, especially in the smaller streams, despite the fact that there is groundwater flow towards the channel". Clearly the yield model includes riparian evapotranspiration from smaller tributary streams. An accurate estimate of drainage density in the basin is unknown, however, the evapotranspiration simulated by Vivier *et al.* (2007), was so high (approximately 90% of the water balance) that it resulted in unrealistically large volumes of negative baseflow (transmission losses) in most of the sub-catchments. Therefore, not only is the high evapotranspiration resulting in a loss of groundwater baseflow but also in losses from the upstream inflows. The observed stream flow data does not support this conclusion and indicates that most of the quaternary catchments receive baseflow at some point. This would seem to point toward the conclusion, that although an accurate estimate of the drainage density is unknown, it is likely that the Vivier *et al.* (2007) value is too high. Sub-catchment A22A was examined more closely in an attempt to better compare the outputs from the two models. Sub-catchment A22A represents an upstream headwater catchment with a large volume of irrigation (Figure 4-63). The data from flow gauge A2H107, located some distance from the sub-catchment outlet, were used, since the gauge at the outlet is highly impacted due to the presence of a large dam. The observed flow data compared with the simulated flow data are shown in Figure 4-64 (model statistics are given in Table 4-24) and illustrate the Pitman model's acceptable visual correlation with the observed data. The groundwater yield model maintains that there is no baseflow in the sub-catchment and 10 × 10^6 m^3 losses into the aquifer from the channel. The observed data certainly does not support this conclusion and while there are periods of no flow in the channel, there are periods where baseflow is evident, even with the extensive irrigation in the sub-catchment.

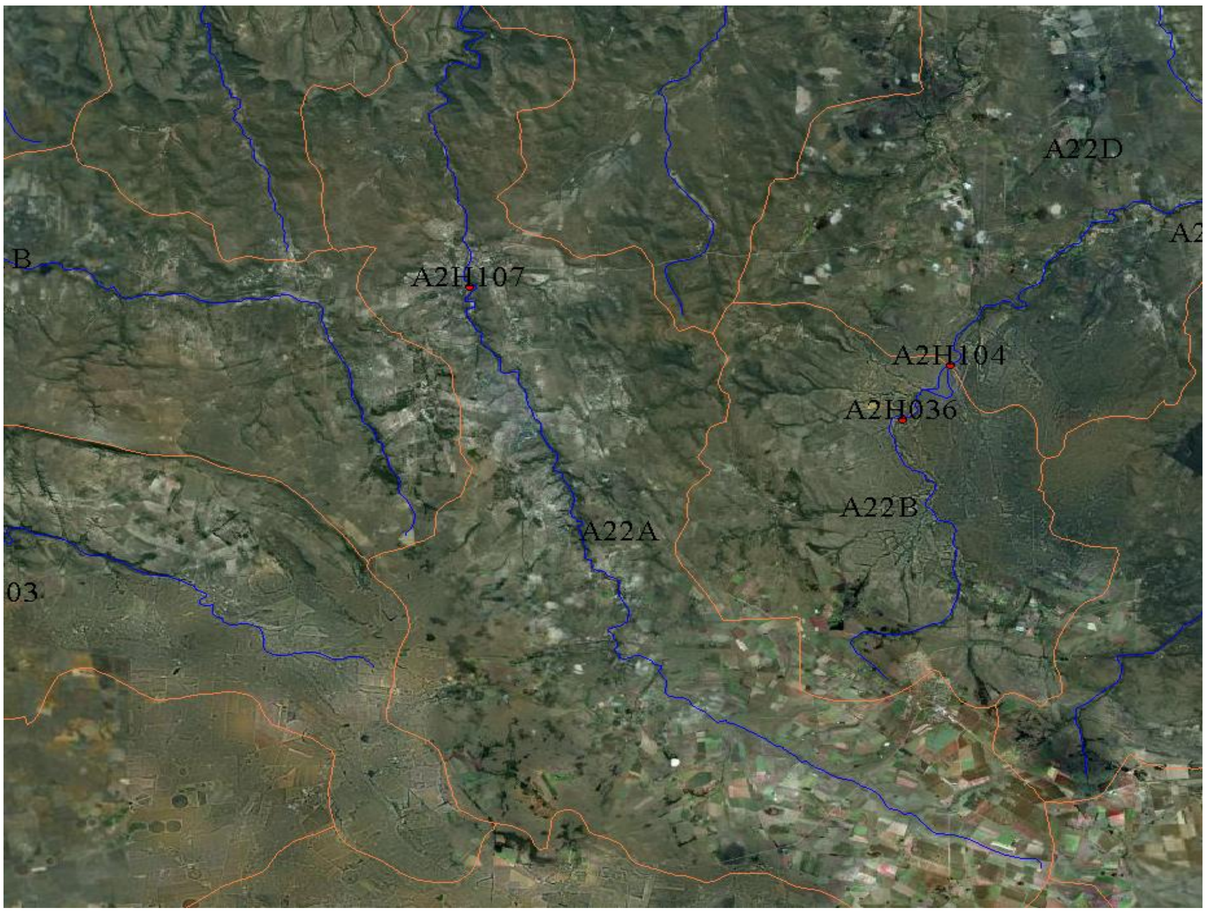


Figure 4-63 Sub-catchment A22A illustrating the volume of irrigation in the upstream parts of the sub-catchment.

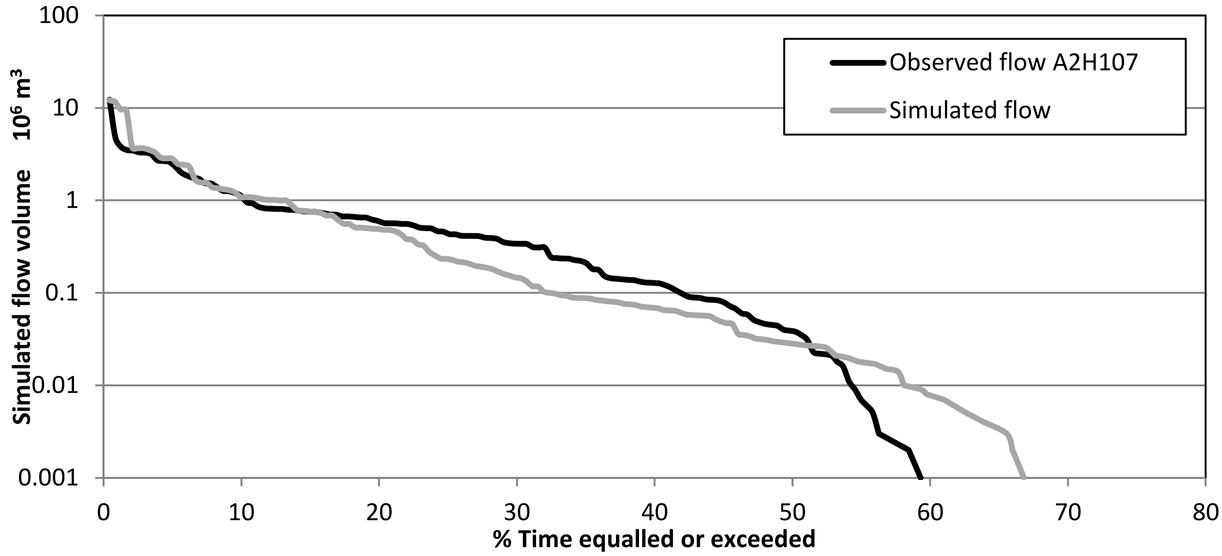


Figure 4-64 Comparison of simulated and observed flow data for A22A (flow gauge A2H107).

While the comparison of the two studies has highlighted the effects of different assumptions regarding the drainage density of a catchment, no study was able to conclusively confirm the likely drainage density. Although the results of the current study indicated that the large losses in evapotranspiration simulated by the groundwater yield model were unrealistic, the comparison of the simulated flow data (Pitman model) against the observed data were poor. Even though the current water use was incorporated into the simulations in some detail, the volume of unaccounted for and poorly documented upstream human interferences meant it was not possible to adequately reproduce the observed flow data with reasonable model performance statistics. Figure 4-65 shows the impacted nature of the observed flow data for sub-catchment A22H which has two large dams upstream of the flow gauge. The difficulty in reproducing the observed patterns of flow could be related to irregular releases from dams or rapid increases in water use in recent years.

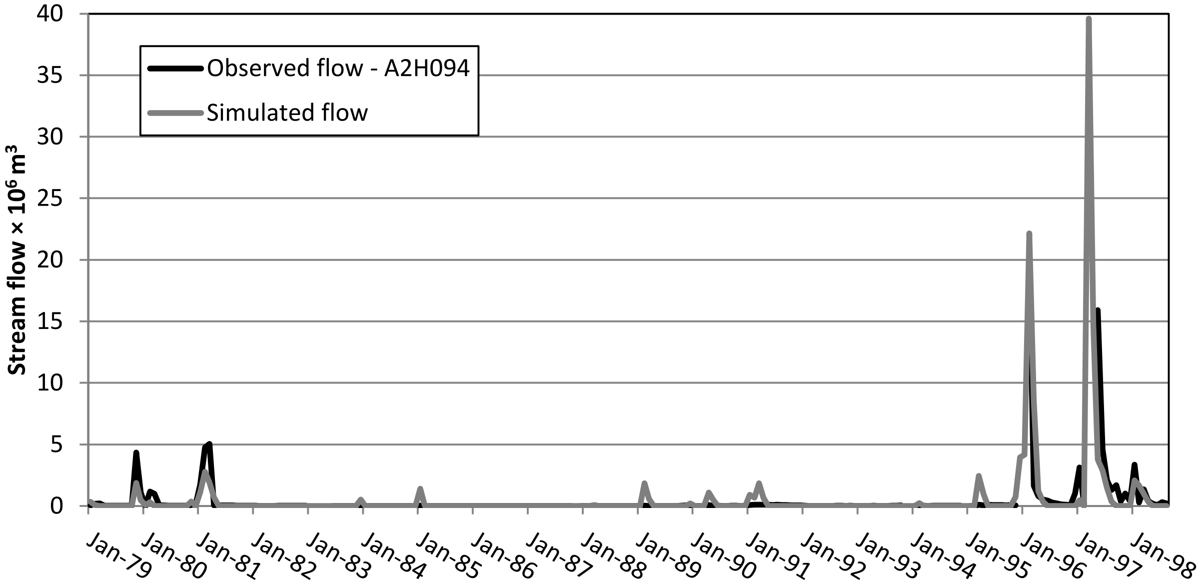


Figure 4-65 Flow data for A22H illustrating the difficulty of reproducing highly impacted observed flow data.

4.7.5. Conclusion

While the uncertainty in the outputs of the simulations is very high, the study has highlighted the significance of certain assumptions regarding the drainage density and subsequent effects on riparian evapotranspiration volumes. Both studies assumed similar riparian strip widths, however, the different assumptions regarding the drainage density resulted in very different volumes of evapotranspiration losses. The groundwater yield model assumed that there is interaction between surface and groundwater in both the main channel and smaller tributary channels whereas this study assumes that interactions between surface and groundwater only occur in the main channel. These different assumptions had a substantial effect on the remaining components of the water balance, with differences in riparian evapotranspiration volumes being the most significant. While drainage density values are uncertain, the high evapotranspiration values simulated by the groundwater yield model, meant the model had to incorporate unrealistically high transmission losses in order to close the water balance. For the present study, the groundwater yield model was unable to confirm the surface water system's ability to sustain this volume. Recent modifications to Vivier's groundwater yield model (subsequent to the Elands River modelling study) mean the model can now be constrained using surface water baseflow values.

5. DISCUSSION

5.1. Introduction

Developing an understanding of the interactions that occur between groundwater and river systems is critical for the effective management and allocation of water resources. However, the lack of observational data represents a serious constraint to understanding that is difficult to overcome, especially when water resources decisions need to be made within a short timeframe. Consequently scientists and practitioners in South Africa are attempting to formulate the most effective method of managing water resources under these circumstances. Much of the discussion is focused on the most appropriate model structure to use with attention given to other forms of uncertainty (input data, parameter uncertainty etc.) often lacking. Beven (2002) argues that the true level of uncertainty in model predictions is not widely appreciated and in South Africa models are frequently expected to produce predictions for water resource management based on insufficient and flawed data. The recent International Association of Hydrological Science's (IAHS) Prediction in Ungauged Basins (PUB) initiative (Sivapalan *et al*., 2003; Hrachowitz *et al*., 2013) focused research towards fundamentally changing the field of hydrological science from calibration based modelling to new approaches focused on fundamental understanding of hydrological systems. This initiative has highlighted the inherent unreliability of traditional approaches in modelling still used today. The question is then what methods can be used to generate the most reliable information given the data scarcity in South Africa? More traditional approaches include relying on expert opinion and testing a model against available data for a range of different circumstances. While these methods can be useful, they do have weaknesses which are difficult to resolve. Beven (2012) suggests a "limit of acceptability" approach to model evaluation as a way of testing models, which would involve thought to define critical experiments that will allow models and their setups to be adequately differentiated.

Without the considerable funding that would be needed for detailed field studies, the re-interpretation of existing data in the context of the possible significance of the process may contribute to the necessary improved understanding. Ward (1984) referred to the use of models to test conceptual hypothesis of catchment response and this type of approach is also used in this study. However, the limitations and uncertainties in both the model and the conceptual hypothesis need to be recognised to prevent the potential problem of a false belief in model results. If the uncertainty can be reduced by an improved interpretation of existing information coupled with limited field investigations, then it may be concluded that a model has the potential to contribute to integrated water resource planning and management. Perhaps one of the critical areas of future research is to determine what type of information can and should be collected to support the type of modelling approach discussed in this thesis.

Much of the data utilised in this study consists of regional values in the form of rainfall and evaporation datasets, naturalised flow data, regional parameter values (both surface and groundwater values) etc. While these types of data can assist in model setup and parameter quantification, there are questions regarding the appropriateness of regional datasets. Of particular concern is that they are often based on small scale data which has been up-scaled to cover large spatially heterogeneous environments. For example, most estimates of aquifer characteristics such as transmissivity and storativity are based on scattered borehole data representing extremely localised information, with very little data on key features like water level variability or aquifer depth. Similarly there are databases (National Groundwater Database, NGWD; Groundwater Information Project, GRIP; WARMS; WR2005; GRA II, NGA, among others) which provide information on regional groundwater use based on the scattered borehole information, with many of the databases reporting conflicting volumes. Data acquired using the methods available are also frequently different for the same location. This is particularly true for estimates of recharge, for which there are numerous different methodologies available which often give conflicting results for the same location. Historical records of observed stream flow are available in many catchments but there is the problem of unaccounted for and poorly documented upstream human interferences and it is often difficult to isolate the natural hydrology regime of a catchment. Conceptualising the processes occurring in a catchment and quantitatively setting a model up to represent these processes can be difficult to achieve due to the lack of robust observational data. The question is how to proceed in the face of these difficulties and this study has promoted the use of qualitative data sources and encouraged thoughtful examination of all the 'soft' and 'hard' data when conceptualising the dominant processes within an environment. A clear example of this approach is demonstrated in the Sabie River catchment where a careful examination of the observed data enabled three runoff generating processes to be clearly identified. The reason the observed data in this environment were able to be so effectively analysed is due to the extended droughts experienced in that particular region together with a large baseflow component in the stream flow. In a drought situation, there is a clear, rapid failure of a large portion of the baseflow component which is not expected in a setting where the baseflow consists predominantly of groundwater contributions (a groundwater contribution would respond more slowly). This analysis enabled the groundwater and interflow contribution to baseflow to be clearly defined which assisted in setting up the model to reflect the conceptualisation with high confidence. Other examples highlighted the value of different sources of information in improving confidence in a models setup. In the Gamagara River catchment there was a large amount of borehole information available but no observed stream flow data. This resulted in a fairly confident quantification of the effects of groundwater dewatering and abstractions, however, a reliable quantification of channel transmission losses was not possible. These types of data concerns have had an impact on all of the objectives outlined in this thesis.

A simple summary of the results of the case studies is given below and shown in Table 5.1. The outcomes of each of the case studies varied considerably and did not seem to exclusively rely on the available data or conceptual understanding. This is probably because critical thinking using all available hard and soft data, often provided unique insights into a system not immediately apparent when examining the typically available observed data.

- *Sabie River:* The observed data together with some knowledge of the geological and topographical setting indicated that there were contributions to the low flows from both interflow (a faster response to droughts) and groundwater (a more stable contribution). The model was able to represent the conceptual understanding sensibly while generating outputs that compared well with the observed data.
- *Grahamstown site:* Short-term simple periodic stream flow observations in a small grassland catchment were used to assess the validity of the low flow algorithms of the Pitman Model (applied in a daily time step version). The model was unable to simulate both a wet and a dry winter satisfactorily possibly due to issues with the evapotranspiration component of the model and not due to the unsaturated zone drainage components. Although much of the data that are available to quantify the expected ranges of some of the water balance components are somewhat uncertain, the simulations appeared to be behavioural, despite the fact that there remain some unresolved uncertainties about the operation of the model under dry winter conditions.
- *Gamagara River:* This environment consists of a large alluvial aquifer with a strong connection to the underlying aquifer which has been extensively dewatered as part of an open cast mining

operation. The only available data are a series of groundwater levels which have been used to assess the model by comparing them to the simulated groundwater slope. Despite data limitations, the model reproduced most of the observed variations in groundwater levels. It has therefore been concluded that the model is representing the dominant, catchment scale, processes in a behavioural manner, even if the exact quantification of these processes remains uncertain.

- *Molopo Dolomitic Eye:* This setting represents a fairly typical example of karst interaction environments in South Africa in terms of both physical environment and data uncertainty. While there are observed flow data in this sub-catchment, wide ranging recharge estimates together with uncertain catchment divide boundaries render it difficult to characterise this environment satisfactorily. Considering the lack of data available, the Pitman model was able to capture the general pattern of flow with the exception of one period (the reason for the discrepancy could not be determined).
- *Breede River:* A complex environment situated in the Cape Fold Belt region of the Western Cape where regional groundwater flows could be relevant. While the conceptualisation of the processes occurring in the catchment seemed to be successful based on the 'knowledge' available, it was not possible to obtain a reliable quantitative estimate. Part of the problem lies in the inherent complexity in the catchment coupled with uncertain flow data impacted by numerous farm dams and return flows. This makes it difficult to properly validate any of the model outputs, demonstrated by a poor comparison between the outputs of the Pitman model and a groundwater model. An uncertainty assessment carried out revealed a large volume of residual uncertainty but the complexity of the catchment prevented the isolation of the dominant sources of this uncertainty. Lastly the spatial scale in the Pitman model was reduced to better represent the heterogeneity inherent in the catchment and while it was not possible to validate the outputs, it improved the conceptual representation of the processes.
- *Buffelsjag River:* Another heterogeneous environment set in the Cape Fold Belt where the low flow contributions to stream flow are from both interflow and groundwater. While the evidence from the analyses (hydrological, water quality and isotope) was far from certain, it supported the conclusion that the low flows are mainly derived from interflow in saturated fractures above the general water table (consistently fed by small wetland type areas) but with some contributions from groundwater, although the volume of groundwater contribution is very uncertain.

- *Elands River:* A large and complex catchment due to impacts from industry, mining, agriculture and a number of large dams. The outputs from the Pitman model compared unfavourably to a groundwater yield model as a consequence of different assumptions regarding the drainage density in the region. While the uncertainty in the outputs of the simulations is very high, the study has highlighted the significance of certain assumptions regarding the drainage density and subsequent effects on riparian evapotranspiration volumes.

Table 5-1 Summary of results from the seven case studies.

Case study	Aquifer type *	Data availability	Conceptual understanding	Uncertainties	Comparison with observed data	Comparison with other models	Ability of model to reject or confirm hypothesis.
Sabie River	(1), (3)	Moderate	Good	The relative contribution of interflow and GW to baseflow	Good	NA	Good
Grahamstown	(1)	Moderate	Good	Evapotranspiration	Moderate	NA	Moderate
Gamagara River	(1), (2), (3)	Moderate	Moderate	Transmission losses, and the effects of GW abstraction on stream flow	Good	NA	Moderate
Molopo Dolomitic eye	(3)	Poor	Poor	GW catchment size	Moderate	NA	Moderate
Breede River	(1)	Poor	Moderate	Recharge, regional GW flows and GW baseflow	Moderate	Poor	Poor
Buffelsjag River	(1)	Moderate	Good	Recharge, and interflow and GW contribution to stream flow	Good	NA	Moderate
Elands River	(1)	Poor	Moderate	Drainage density and GW contribution to baseflow	Moderate	Poor	Good

*(1) Hard rock, (2) Alluvial, (3) Karst, (4) Primary

5.2. Has the thesis achieved its objectives?

5.2.1. Develop conceptual hypothesis about the surface and groundwater interaction processes in different environments.

A perceptual model is based upon the understanding of the real world catchment system, such as flow paths and runoff production mechanisms. If this understanding is poor, the perceptual model and subsequent quantitative modelling of the system will be highly uncertain. In an attempt to improve the understanding of the dominant interaction processes which occur under typical South African conditions existing data were examined and processes deemed important for large scale water resource investigations were identified and detailed. The interaction environments found over much of South Africa were categorised into four main types of interaction environment, including fractured rock aquifers (forming the majority of South Africa's aquifers), karst aquifers (forming the most productive aquifers), primary aquifers and alluvial aquifers. The dominant processes identified within each environment were outlined in Chapter 3 and the main conclusions included:

- The presence of large unsaturated zones, sometimes tens of metres thick, over large parts of South Africa mean that unsaturated zone processes are more important than traditional perceptual models often imply.
- Many springs are the result of fracture flow and can consist of unsaturated zone water (interflow) which emerges at the surface before reaching the regional groundwater table.
- While many aquifer environments are highly heterogeneous (particularly fractured rock and karst environments), at the larger scale characteristics such as transmissivity and storativity can be averaged. This is often termed a regional background value (Sanchez-Vila *et al.*, 1996; Manga, 1997) and is generally applicable over the entire aquifer (at large scales).
- Numerous dykes and sills over South Africa can impact aquifers by increasing the fracturing within the surrounding aquifer (at the contact zone), or by compartmentalising an aquifer and hydraulically disconnecting it from the surrounding aquifer system.
- These dykes and sills, as well as large fault zone systems, can cause preferential flow which could contribute to regional groundwater flows and can have strongly differing transmissivities to the regional background values.

While our understanding of the conceptual processes in South Africa is still somewhat uncertain, a perceptual model is still valuable in providing basic insights into a specific environment and identifying critical processes that may be present in a certain type of setting. The uncertainty stems from processes which are complex and not yet fully understood but, which have an important control on how the system works and which may be essentially unknowable (Beven, 2009). Many of these gaps in understanding are due to the difficulty in observing sub-surface processes at suitable scales. For example, processes such as regional groundwater flows are almost impossible to identify and measure, and hence have largely been ignored in integrated models.

Where possible a catchment specific perceptual model should be produced using any available qualitative and quantitative ‘knowledge’. This will assist in setting up a model to reflect the actual processes occurring within the catchment. There are not always enough data to produce a comprehensive perceptual model, and in some cases, such as the Upper Breede River, the data are available but too uncertain for a reliable quantitative understanding. Even though a relatively detailed perceptual model had previously been constructed based on the available ‘knowledge’, the extreme spatial heterogeneity combined with uncertain observed data resulted in a complex mix of uncertainty within the input data, the parameter values and the model structure (possible regional groundwater flows). The case studies examined in this thesis, however, nearly always demonstrated the benefit of a thoughtful examination of existing information in building up a general understanding about the dominant processes within a catchment. This can reduce the uncertainty in a model’s outputs by decreasing both a model’s structural uncertainty (does it include the critical processes?) and equifinality in the model setup (realistic quantification of the parameters). A study carried out by Hughes (2009) in the Seekoi River further highlights the advantage of this approach and demonstrated how a consensus between both surface and groundwater hydrologists resulted in an improved conceptual understanding of the prevailing processes based on an interpretation of the limited data. However, this required that some ‘conventional wisdom’ or traditional perceptions were put aside and that participating surface and groundwater hydrologists were able to be flexible and open minded in their interpretations.

Without a careful examination of the available information in a particularly uncertain setting, water resource assessments run the risk of misinterpreting critical functions leading to incorrect water resource decisions. For example, it is important to distinguish between the various runoff generating processes in a catchment. If low flows cannot be differentiated into interflow (from soil moisture runoff, unsaturated zone flow and perched aquifers) and groundwater contributions, it is possible that incorrect conclusions will be reached about the impacts of different catchment developments on the flow in rivers. This was demonstrated in the Sabie River where the 'rapidly responding' runoff (within the low flows) identified in the observed flow data, seem to be superimposed upon a more stable groundwater contribution to low flow. Since this catchment is underlain by quartzites and dolomite and includes an area of steep topography, it was hypothesised that the rapidly responding portion of the low flows was derived from a lateral flow vector in the fractures of the unsaturated zone and the large macro-pores of the dolomite. Similarly, water quality analysis combined with isotope sampling in the Buffelsjag River (Kapangaziwiri *et al*., 2011) seemed to confirm that a large proportion of the low flows were derived from unsaturated zone flow in the fractured quartzite ridge. The conclusions in both cases would have a significant impact on water resource decisions as the development of groundwater resources would not affect the portion of low flows which consist of unsaturated zone flow. Careful consideration within a perceptual model can also prevent the misinterpretation of available information. Clear thought regarding sensible process representations and outcomes can certainly prevent issues like those encountered in the Elands River catchment from occurring. In this case, incorrect assumptions (in a separate groundwater study) regarding the drainage density in an environment led to unrealistically high simulated riparian evapotranspiration volumes. While the extent to which groundwater contributes to surface water in smaller tributary channels is unknown, the lack of clear thought regarding the effects of such high values led to improbable values in the remainder of the simulated water balance.

5.2.2. Assess the conceptual 'correctness' of the Pitman model together with its ability to quantify the processes identified.

Integrated coarse scale models are still viewed with suspicion in South Africa, largely because their outputs cannot be sufficiently validated with available data. Given the current state of data availability in South Africa, is it possible to reliably assess if a model is sensibly representing the processes while generating realistic results, thus increasing the confidence in a models outputs? A further issue includes the validation of a model's outputs in ungauged catchments and whether it is possible to ascertain whether the model is producing realistic results even if it seems to be simulating the correct processes. Internationally, authors such as Beven (2009) and Oreskes *et al*. (1994) suggest that the verification and validation of models of open natural systems is impossible, despite their widespread use in the modelling literature. They point out that few, if any models are entirely confirmed by the available data, but equally few are refuted and argue only a conditional confirmation is possible. It is conditional as it may depend on the calibration of parameters or other auxiliary conditions and may also depend on the period of data used in evaluation. In addition, the processes represented by a model are always reliant on smaller scale processes which cannot be adequately represented without simplification. This means that although a model might be satisfactorily replicating the observed data, we cannot be certain that it is because the model is representing the processes sensibly. This does not mean that attempts to verify models are useless, only that we should be careful about the limitations of their domain of validity.

While the difficulty of validating a model based on its representation of catchment scale processes which are based on smaller scale processes has been mentioned, there are large scale processes which are clearly going to have an impact on the water balance in a catchment. These are important to include in any model attempting to simulate conditions in that catchment even though the mathematical representation is a simplification of the physical process. For example, the Upper Breede River represents a catchment where the evidence points toward regional groundwater flows being significant, namely geological evidence together with an unrealistic water balance. There are clearly catchment scale processes which when incorporated into a model have the necessary impact. While arbitrarily incorporating additional processes into a model's structure to try and improve the simulations is not the objective, adding new processes which make conceptual sense in terms of trying to represent the critical processes realistically, is. In many cases, however, it is not possible to determine if the representation is correct based on our current understanding and data availability.

Nevertheless assessing the conceptual 'correctness' or structural uncertainty within a model can be very important if one of the objectives is getting the right answer for the right reason (Kirchner, 2006). While it is difficult to differentiate between input, parameter and structural uncertainty, in this thesis perceptual models have been used to partly identify the uncertainty through a careful examination of the evidence for specific processes compared with the conceptual structure of the Pitman model. Arguably, there are two levels of structural uncertainty that should be considered. The first is whether or not certain processes, known to exist in the real world, are represented in a model. While the Pitman model does not encompass all of the critical processes found in South Africa, it does encompass the majority of the important processes identified (Table 3-3). Perhaps one of the most critical processes, for the purposes of IWRM, is the inclusion of a runoff generating mechanism which represents that portion of baseflow which does not originate from groundwater. The Pitman model has three possible runoff generating mechanisms which are surface runoff, groundwater runoff and unsaturated zone runoff. The unsaturated zone runoff component uses the same function to represent runoff from both the soil moisture zone and that portion of the unsaturated zone which generates interflow (Hughes, 2010b). The complexity and potential equifinality in the model is increased with the additional functionality, however, this was deemed necessary in a model developed for South African conditions and one which aimed to get the right results for the right reasons.

The second type of structural uncertainty is whether the algorithms used in the model can adequately represent the non-linearities or thresholds that occur within the relationships between storages and fluxes, or one flux and another. This type of structural uncertainty is less-straightforward to distinguish from parameter and data input uncertainty, largely because it varies between locations and requires comparison with observational data to resolve. The Pitman model uses a variety of output functions to assess this type of information in some detail, including various time series of simulated water balance components, such as recharge, groundwater baseflow, unsaturated zone baseflow, etc.

The confidence in the model outputs will inevitably vary with the amount and quality of the information that is available. In situations where the uncertainty in the results proves to be excessive, additional data collection would be needed to reduce the source of the uncertainty. While the model is clearly representing the dominant processes (recharge, intermittent stream flow, groundwater evapotranspiration losses, transmission losses and dewatering effects) in a behavioural manner, the exact quantification of these processes remains uncertain in many of the case studies. Even when the model outputs compared favourably with the available observed data, the uncertainty within the data together with the equifinality present as a result of uncertain input data and parameter values (process understanding can only reduce this uncertainty, not eliminate it), could not always be fully resolved with the available data . If models cannot be verified or validated but only conditionally confirmed, can we, particularly in a data scarce country, have any belief in their predictions? Designating a model as invalid or false is a difficult issue partly due to the fact that all models can be considered false if examined in sufficient detail, and partly due to the fact that, in any practical problem that demands some predictions, at least one model must be retained to carry out the simulations (Beven, 2009). Water resources need to be managed, and as long as a set of parameters can be found that give some acceptable fit to the available observations, then predictions can be made. While the arguments regarding the difficulty of model validation are compelling and the data on which most of the case studies are based scarce, the environmental modeller will have to deal with degrees of acceptability of one or more models for a particular purpose (Beven, 2009) and the argument here is that ensuring a model is realistically representing the processes (as opposed to calibration based modelling) can validate a models outputs to a better degree.

5.2.3. Investigate uncertainties associated with the structure and application of the groundwater interaction components of the model.

Le Moine *et al*. (2007) argue that models developed for use in hydrological analysis and engineering applications must be as robust and generic as possible due to limited time and input data resources. They argue for generality in models as a new model for every application would eliminate the learning that comes with repeated applications of the same model. Since all models are imperfect representations of reality, the knowledge of the imperfections is one of the prerequisites of conscious model use. There were uncertainties within the model structure identified in Chapter 3, many of which were examined in the case studies. Structural uncertainties, related to the spatial and temporal scales often utilised in conceptual type models, can be substantial since these models are typically designed to represent catchment scale processes (Wagener *et al*., 2004). These uncertainties include the extent to which a coarse scale model is able to adequately represent reality and there remain questions about spatial scale effects (Booij, 2003) on the estimation of parameter values. The complexity of sub-surface processes has frequently been used as an argument against the coarse scale quantification of surface and groundwater interactions (WRC Workshop, Pretoria, 25 October 2011, Witthuser, K.), however there are rarely enough available data to manage water resources at a finer scale. While coarse scale quantification has limitations, surface water resources have successfully been managed through a gross simplification of reality and although sub-surface processes have the added complexity of being difficult to measure at large scales, there have been examples of coarse scale models which have successfully quantified surface and groundwater interactions (Le Moine *et al*., 2008; Ivkovic *et al*., 2005). Most of these examples, however, are found in data rich scenarios where the validation of these models is relatively straightforward. In some situations, however, the representation of a sub-catchment as one spatial unit is not ideal, particularly within spatially heterogeneous sub-catchments. The Cape Fold Belt region is an example and both the Upper Breede River and the Buffelsjag River case studies demonstrated the difficulty of representing steep mountain slopes together with flat valley bottoms in the parameters of a single spatial unit. While the unreliable observed data in these sub-catchments meant the quantitative assessments were unable to be satisfactorily verified, breaking the sub-catchments down into smaller spatial units representing the zones of recharge and discharge certainly made conceptual sense in terms of both spatial climate variations and more representative parameter values. The results from the simulations where the total sub-catchment was divided into sub-areas, illustrate that the model is robust enough to handle reduced spatial scale simulations and that this approach could be essential in particularly heterogeneous regions.

Temporal uncertainties in integrated models are related to the differences in the lag and attenuation times of different processes, represented in a single model. The Pitman model does not have a groundwater or recharge lag routine, as the groundwater function acts as a routing reservoir. The transmissivity and drainage density parameters largely control the lag and attenuation of groundwater discharge for a given pattern of recharge and storativity. Studies show groundwater discharge hydrographs respond very quickly to rainfall events (Kelbe and Germishuye, 2010; Sun, 2004), and therefore the assumption is that the displacement effect negates the need for a specific lag routine. In some environments, however, a more substantial groundwater lag can be present and in these settings the model's routing reservoir might not be sufficient. This is especially true in karst environments which can have non-linear responses to inputs and are often affected by thresholds. There were no clear threshold effects evident in the observed flow data of the Molopo dolomitic eye case study and the model was able to reproduce the general patterns in the flow data satisfactorily, apart from one period (the reason for this discrepancy was unable to be determined). However, karst aquifers are very diverse and threshold effects have been reported in the literature (Bailly-Comte *et al*., 2009) therefore further application of the model in these settings is recommended to test the model across a wider range of karst conditions. Examining the performance of the model in terms of representing the lag and attenuation times in the other 'typical' interaction environments (fractured rock etc.) led to the conclusion that, while not always straightforward to determine, the model's current setup (at a monthly time step) seems to be acceptable. The daily version of the model, however, includes temporal uncertainties that would need to be resolved through further testing of the daily version of the model. The daily version of the model was applied to the Grahamstown study site where the daily variations in recharge were not relevant. In a catchment where recharge is significant, the daily model might need a routing factor to avoid recharge water reaching the groundwater table too quickly. Additionally groundwater outflow would probably need a routing factor as the 'displacement effect' relied upon at the monthly time scale would lead to an unrealistically fast response to recharge. Some of the issues introduced when implementing the model at the daily rather than monthly time scale can be resolved through parameter value changes, but issues such as the delay and routing of recharge and groundwater outflow would not easily be resolved in this manner.

The application of the model in the Upper Breede River highlighted the model's inability to represent regional groundwater flows. The Pitman model formulation does not currently account for situations where surface water and groundwater catchment boundaries do not coincide (Hughes, 2004), although a model component that allows for the movement of groundwater out of the system would be relatively simple to incorporate. Before the process can be satisfactorily included in the model structure, more conceptual and quantitative information on this process is needed. Also, the model cannot currently account for settings where the direction of groundwater movement is different from the direction of surface water movement. However, from available information (Vivier, 2009; Wright and Xu, 2000), indications are that sub-surface flow follows the topography and the direction of the surface water flow, for the majority of catchments in South Africa.

Direct recharge (unrelated to soil moisture storage) into bare rock areas is not accounted for in the model and can be important in fractured rock settings particularly in high relief areas. The assumption is that direct recharge would occur in areas with thin soils. High rainfall in an area with thin soils would lead to rapid saturation of the soil moisture store inducing high groundwater recharge, increased groundwater gradients and subsequent increased discharge to the channel. Making use of a high value for the parameter GPOW would simulate these conditions of substantial non-linearity in the relationship between recharge and soil moisture storage. This implicit process representation is assumed to sufficiently replicate direct recharge in bare rock areas and the model successfully simulated environments which almost certainly include direct recharge (Sabie, Buffelsjag and Upper Breede Rivers). Further information on this process, however, would be useful to fully evaluate whether the model setup is satisfactory. However, these type of data would not be straightforward to obtain, certainly at a scale appropriate for water resource management. A catchment with highly variable soil depths is assumed to be better represented by splitting a sub-catchment into recharge and discharge zones which would further improve the simulation of this process from a conceptually realistic perspective.

While the assumption is that the majority of groundwater discharge to surface water in South Africa is from unconfined aquifers, it is accepted that in some settings confined aquifers play an important role. While the Pitman model does not explicitly simulate confined conditions, it can represent them implicitly. Large scale data on confined aquifers in South Africa is not common, largely due to the difficulty in obtaining reliable information on confined aquifers located mostly at substantial depths, and further research into groundwater discharge to baseflow from these types of aquifers is needed.

The effect of large scale groundwater abstraction on groundwater discharge to stream flow is a process that is not yet well understood. The ability of the model to acceptably simulate this process is therefore questionable as there is no way to clearly validate the simulations. A number of pragmatic decisions were necessary while designing the abstraction functions in the model. Some of these included the implicit rather than explicit representation of direct recharge, the representation of the groundwater levels in a catchment using two groundwater gradients, the one directional connection between the far channel gradient and near channel gradient, etc. The two groundwater gradients are represented as 40% (near to the channel) and 60% (far from the channel) of the catchment area and it is uncertain how appropriate this assumption is in many environments. The model assumes water moves in only one direction (from the remote groundwater store toward the near channel groundwater store) even if the remote groundwater gradient is low. This could be significant in an environment such as the Gamagara River with a strong connection between the alluvium and surrounding regional aquifer, particularly after a large recharge event from transmission losses. While it is assumed that the model can produce estimates of sustainable abstraction volumes (if the recharge and riparian evapotranspiration estimates are reliable) (Hughes *et al*., 2010), it cannot account for any limitations related to the availability of suitable borehole sites, nor localised aquifer transmissivities and storativities that will affect sustainable pumping rates. However, the Pitman model was never intended for this purpose and always recognised that more detailed groundwater models would be needed for site specific borehole or well field design. In semi-arid areas groundwater abstractions will impact mainly on static channel pool storage and riparian vegetation. While channel pool storage has not formed part of the model configuration in the case studies in this thesis, a previous example in a ephemeral river (the Seekoi River; Hughes, 2009) illustrated that the model is capable of simulating the dynamics of pool storage affected by both surface and groundwater inflows. While Hughes *et al*. (2010) concluded that the model is able to simulate the effects of abstraction in a realistic manner, the values of the simulated components of the groundwater balance as well as estimates of sustainable abstraction volumes are subject to a large degree of uncertainty.

Perhaps one of the most important sources of uncertainty lies in estimating the values of some of the groundwater parameters. The model utilises groundwater data from the GRA II database (DWAF, 2005a) but these values have a large degree of uncertainty associated with them. The Elands River catchment study highlighted the significance of assumptions made regarding some of the groundwater parameters. In this case two different studies (a groundwater study and the present study) assumed different drainage density parameters in an environment, which resulted in very different simulated water balances. The likely drainage density remains unknown in these types of environments, and information on the degree to which groundwater contributes to baseflow in tributary channels would be very valuable. Parameter uncertainty is difficult to resolve without further detailed information.

5.2.4. Develop recommendations for environments where the model can and cannot be reliably applied.

The search for appropriate catchments (and associated information sources) to use for this study reinforced the perception that many of the hydrological records are of limited value to hydrological understanding. The assessments at most of the study sites are largely based on typically available data and with the exception of the Grahamstown site and a single sampling event at the Buffulsjag River site, no other site specific field studies have been undertaken. The available evidence for conceptualising the interaction processes is very variable and suggests that confidence in the validation of the results varies considerably. The results from the Sabie River, Gamagara River and Grahamstown site were fairly conclusive and they enabled the hydrological model parameters to be established with high confidence. The remaining study sites which include the Upper Breede River, the Buffelsjag River, the Elands River and the Molopo dolomitic eye are less conclusive and further data would have been very useful.

Since the majority of aquifers in South Africa consist of fractured rock aquifers, it is imperative that any model developed for South African conditions is able to handle the type of processes associated with this environment. While the Pitman model does not simulate individual fractures, it relies on the averaging of aquifer characteristics such as storativity and transmissivity (Sanchez-Vila *et al*., 1996; Manga, 1997) and settings with large fault zones or dyke systems would need to be accounted for implicitly in the model (for example, the catchment could be broken down into smaller spatial units). There were four case studies evaluated which were located in fractured rock environments namely the Elands, Buffelsjag, Sabie and Upper Breede Rivers. The importance of unsaturated zone fracture flow in this type of setting was demonstrated clearly in the case studies and the model seemed to simulate both the unsaturated zone and groundwater runoff responses in a behavioural manner. The Sabie River demonstrated a clear unsaturated zone runoff response which the model represented with high confidence. The results of the Eland River investigation were less conclusive largely due to highly impacted observed flows. The Buffelsjag and Upper Breede Rivers were represented in a conceptually sensible manner in the model, but the quantitative results were unable to be sufficiently validated. These investigations also highlighted the potential for regional groundwater flows in some parts of South Africa, which the model in its current form cannot satisfactorily represent.

Karst environments are especially complex largely due to highly heterogeneous systems often with lags and thresholds, together with high volumes of water use which impact on the reliability of the available observed data. These environments are not straightforward to structurally represent or to parameterise and frequently have surface and groundwater catchment boundaries that do not coincide. Despite these uncertainties, both the Sabie River and the Molopo Dolomite Eye study sites seemed to generate behavioural results based on the information available. While the Sabie study site represents a dual environment (both karst and fractured rock aquifers), the Molopo Dolomite eye represents a pure karst aquifer. In this environment the runoff was represented almost exclusively by groundwater as there is no surface runoff in the area. This enabled the groundwater catchment size to be approximately determined based on the outflow volume (of the eye) and regional recharge estimates. The groundwater catchment was clearly larger than the surface water catchment (a common characteristic of karst environments), and although in this case the conceptualisation was successful, the model would be unable to represent an environment with differing catchment boundaries if surface runoff was significant. The model was able to capture the general pattern characteristics of the observed flow data, but was unable to represent all the fluctuations. This could be due to the models inability to explicitly represent the structural heterogeneity of the karst aquifer (which is well developed) sufficiently. However, the model’s response was considered adequate for the purposes of large scale water resource management. In an aquifer where thresholds are clearly present, it would be fairly straightforward to incorporate this feature into the model through the use of storage reservoirs, however, information on threshold characteristics (storage volume, seepage loss etc.) in an aquifer are almost impossible to obtain. While knowledge of the existence of a process is valuable, without the data required to characterise the process it is extremely difficult to represent satisfactorily in a model.

Alluvial aquifers can play an important role in the availability of groundwater resources, especially in semi-arid areas, through infiltration from river channels during periods of stream flow. While the model does include a component to account for this process, it is extremely difficult to quantify and can vary considerably between locations. Setting up the model to represent an alluvial aquifer on a catchment scale is not straightforward since the characteristics, and hence parameterisation, of alluvium is very different from the surrounding aquifer. In the semi-arid parts of South Africa where the majority of large alluvial aquifers are found, regional aquifers often consist of fractured rock and the alluvium is assumed to recharge the surrounding aquifer. Alluvial aquifers in other parts of the country, such as the Cape Fold Belt region, are assumed to either be recharged by the surrounding aquifer or have an intermittent recharge relationship with the surrounding aquifer. The Gamagara River Study site represents an alluvial aquifer situated in a semi-arid area and while there is still uncertainty associated with understanding the dynamics of this environment, it was fairly straightforward to set the model up using the available data with some knowledge of the physical environment. Due to a lack of adequate structure in the Pitman model to represent a 'duel' environment, different sets of parameters were used to represent the groundwater levels within the alluvium (before dewatering) and within the regional aquifer (after dewatering). Despite these limitations, the model reproduced most of the observed variations in groundwater levels. It has therefore been concluded that the model is representing the dominant, catchment scale, processes (transmission losses and large scale abstractions) in a behavioural manner, even if the exact quantification of these processes remains uncertain.

No primary aquifers were included as part of this study, largely due to the difficulty of finding a catchment of sufficient scale with observed data that was not highly impacted. The majority of primary aquifers in South Africa are found in coastal regions with extensive impacts on the natural hydrology from land use change together with high volumes of water use and return flows. As conditions in this type of environment are relatively homogeneous, there are no reasons to suggest that the model would be unable to represent the catchment scale processes. However, further studies would be required to confirm this assumption.

Even though inherent uncertainty remains in all of the environments, the model was able to demonstrate largely behavioural results in the case studies presented. Further work in karst and primary aquifer areas is recommended to further explore the models capabilities in these environments. There are situations where the model can represent a particular process in an environment but not for the right reasons. For example the downstream groundwater outflow function was used to represent the removal of water from the Breede River catchment which in reality was possibly leaving the catchment via regional groundwater flows. While the destination of this water was ultimately incorrect, it enabled the process to be replicated in the relevant catchments. Ignorance regarding the models capabilities and limitations can introduce significant error and uncertainty into the modelling process and the application of the model requires training and experience, particularly for more complex environments which include processes that need to be represented implicitly or that require additional sub-model components (reservoir or wetland sub-model). Particular settings which the model cannot represent include:

- Environments with regional groundwater flows where large scale volumes of groundwater are moving out of or into a catchment.
- Confined aquifer systems are not represented explicitly in the model and while they can be represented implicitly, the data required to validate the model's behaviour in these environments is rarely available and therefore the use of the model in confined aquifers is highly uncertain.
- Karst aquifer systems with significant groundwater lags and thresholds are unable to be represented explicitly and the data required to represent these processes implicitly are rarely available.

6. CONCLUSION

Developing an understanding of the interactions that occur between surface and groundwater systems is critical for the effective management and allocation of water resources. This thesis has considered whether it is possible to reliably quantify the interactions on a scale appropriate for water resource assessments in a data scarce region such as South Africa. The first part of the thesis focused on characterising the 'typical' interaction environments found in South Africa and the second addressed the application and evaluation of the modified Pitman model (Hughes, 2004), which allows for surface and groundwater interaction behaviour at the catchment scale to be simulated.

In a data scarce environment such as South Africa, traditional forms of model testing have 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. In South Africa, there is still much controversy over the most 'appropriate' type of model to be used and part of the disagreement stems from the fact that we cannot validate models adequately. The issue is whether, given the different sources of uncertainty in the modelling process, we can differentiate one conceptual flow path from another in trying to refine the understanding and consequently have more faith in model predictions. While new philosophies and theories on modelling and results validation have been published (Beven, 2002; Gupta *et al*., 2008), in many cases, models are still being validated and compared using sparse and uncertain datasets. While this thesis does not delve into these new theories, it promotes a simple common sense approach to water resource modelling which makes use of all the available 'hard' and 'soft' data together with thoughtful conceptual examination of the processes occurring in an environment. This philosophy focuses on the fundamental understanding of hydrological systems rather than pure calibration based modelling and ensures as far as possible that a model is generating sensible results by simulating realistic processes. While it is necessary to calibrate many of the parameters in the Pitman model, the point is that pure calibration based modelling can give the right answers for the wrong reasons and this has water resource and land use management implications. This is especially true in a highly parameterised model, where the degree of equifinality is possibly significant. Focusing on realistically representing the conceptual understanding ensures that the parameter space is limited and the dominant processes are being represented as far as possible.

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 and models are frequently expected to produce predictions based on insufficient and flawed data. 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). Uncertainties within the input data used to run a model cannot be resolved without allocating substantial resources toward expanding existing monitoring networks. In the past there has always been a strong reliance on calibration in gauged basins. While calibration guidelines can help to reduce subjectivity, there is little doubt that different users will frequently end up with different parameter sets for the same basin, using the same data and the same model. Beven (2012) argues that any model that does fit a calibration data set exceedingly well should be considered suspect, since it may be fitting to the errors in the data, i.e. over fitting of a heavily parameterised model. There is clearly a need for additional information that is targeted at filling in some of the gaps in our understanding of hydrological processes at the scale of sub-catchments. There is also the question of data richness which relates not only to the amount of data available, but whether or not the data are directly appropriate for the type of water resources management and planning decisions that have to be made. This is the difference between data richness from a purely hydrological perspective and from a practical water resources management perspective. Given the limitations of human and financial resources that exist in South Africa, it is important to clearly identify what those gaps are and what is the most focused and cost effective methods of filling them. Tracer studies have the potential to provide useful information at relatively small cost and they have a wide applicability as they can be utilised in a various ways, from small to large scale studies and in many different hydrological environments. Tracer based hydrograph separation data yields groundwater discharge rates from reach to catchment scale and have even been used to monitor the pathways of inter-basin groundwater transfer in Costa Rica (Genereux *et al*., 2002). Further data that would be helpful includes:

- Information on large scale recharge.
- More extensive rainfall gauge networks.
- Information on the typical densities of rivers that receive groundwater, i.e. tributary channels.
- Average background values for transmissivity and storativity.

- Improved data on both surface water and groundwater use and return flows.
- Data on large scale transmission losses (water balance data) on some of the large river systems in South Africa (this data would ideally include more detailed rainfall data).
- Data on unsaturated zone/interflow contribution to river flow (probably tracer data).

While improved data would substantially reduce the dominant sources of uncertainty (input data and appropriate parameter values), it seems fairly certain that in the foreseeable future integrated hydrological models will have to make important predictions for water resource management based on the scarce data that are currently available. The question remains whether reliable integrated assessments can be carried out given the existing data. Certainly, with all the possible sources of uncertainty in a data scarce country such as South Africa, pure calibration based modelling is unlikely to produce reliable information for water resource managers.

The integration of surface and groundwater interaction processes at a scale suitable for water resource assessments will never be straightforward and integrated coarse scale conceptual models can never provide a high level of spatial detail such as 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 such as the Pitman model have shown to be as equally reliable as more detailed models. While many of the case studies demonstrated that the model could sensibly represent the dominant processes at sub-catchment scales, given the available data, it is recognised that hydrological (both surface and groundwater) processes can be highly spatially heterogeneous. In certain regions it may be necessary to employ the alternative modelling approach developed by Hughes *et al*. (2012) which divides the total catchment into sub-areas representing the zones of recharge and discharge. This conclusion demonstrates that the model is flexible enough to be applied in a different spatial context, in the way that it is applied to include additional spatial detail if this is considered necessary or appropriate in a specific situation. Some of the structural uncertainties identified in the model are associated with process averaging at the catchment scale; therefore spatial disaggregation can potentially reduce some of these uncertainties.

From a surface water perspective the Pitman model is one of the most generally accepted models with full functionality in terms of representing surface hydrological processes explicitly. The question is whether the new groundwater components incorporated into the model, adequately represent the groundwater environment to satisfactorily simulate surface and groundwater interaction systems? Secondly, do the new components add to the functionality of the model in terms of carrying out water resource analysis? While model structural uncertainties are not necessarily straightforward to resolve, the Pitman model was deemed to be structurally sound as it represents most processes known to exist in South Africa, and seems to adequately represent the non-linearity's or thresholds that occur within the relationships between storages and fluxes. While it is not suggested that the new components are a totally realistic representation of groundwater flow in a drainage basin, they have the right effects and the majority of the parameter values of the model should be approximately quantifiable from existing information. The strength of the Pitman model's 'compromise' structure, described in Chapter 3, is its ability to be parameterised fairly easily, while still retaining sufficient detail to represent a variety of processes. The uncertainty version of the model can be used to explore different parameter values and their effects on a range of outputs such as evapotranspiration, recharge, stream flow, etc. This allows the parameter space to be explored in a more efficient way than the single run version of the model allows and it seems the majority of the groundwater parameters can be quantified within relatively narrow bands of uncertainty. In attempting to determine the most behavioural model setup, the uncertainty model together with other model functions such as sensitivity analysis can aid in guiding the model user in formulating and streamlining conceptual ideas. The absolute validation of any model's outputs is not possible in a data scarce country, and it is rarely possible to confirm a single hypothesis in a catchment. However, the Pitman model has demonstrated that it is possible to reject or accept as possible certain hypothesis in many situations. Both the Sabie and Elands River case studies demonstrated the models ability to provide relevant insights into a system. In the case of the Sabie River, the model was able to reliably identify a large portion of the baseflow consisted of unsaturated zone flow or interflow. The Elands River case study demonstrated the model's ability to narrow the possible range of riparian evapotranspiration within the catchment and reject the hypothesis reached by a different model. The incorporation of surface and groundwater interaction routines seems to have resulted in a more robust and realistic model of basin hydrology. DWAF (2004a) describe the essential components of any technique developed for application in integrated water resource management and the modified Pitman model certainly fulfils most if not all the components, 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.

The overall conclusion is that the model, although simplified, is capable of representing the catchment scale processes that occur under most South African conditions, but that there are some specific situations that cannot be represented by the model formulation. In addition, there are some situations where quantifying appropriate parameter values for the surface and groundwater components of the model is relatively straightforward, while there are others where further field-based information or additional conceptual understanding is required to reduce relatively large uncertainty in the estimation of appropriate parameter values. Further development of the model base on improved data availability and process understanding could include:

- Improved representation of transmission losses (these have been included in the model according to current understanding which is uncertain).
- The incorporation of regional groundwater flows.
- Improved representation of the effects of groundwater abstraction on the groundwater contribution to stream flow.

While the model was deemed to be robust based on the behavioural results obtained in the majority of the case studies, in many cases a quantitative validation of the outputs was just not possible based on the available data. In these cases, the model was judged on its ability to represent the conceptualisation of the processes occurring in the catchments. The validation of quantitative model outputs is critical for water resource management decisions and suggests that the identification and inclusion of uncertainty in the outputs of a model is essential. Many water resource decisions are still made without adequate account being taken of the uncertainties inherent in assessing the response of the hydrological systems. It is fairly clear that achieving the optimum model of a hydrological system may be fraught with difficulty. Instead Beven (2009) states that there may be many different model structures and parameter sets within model structures that are consistent in some sense with the uncertainties in the available data and many different ways of estimating uncertainties in the predictions. This makes it very difficult from a practitioner's point of view to decide which model and uncertainty estimation method to use. According to Beven (2009), this may be a transitional problem and a future consensus about which techniques to use may become clearer as we learn more about how to estimate the uncertainties associated with hydrological systems.

While it would be reassuring to be able to conclude this thesis by making some more concrete recommendations or providing sets of guidelines about which methods are best to use for different types of cases, we are not yet at that stage. Beven (2009) argues that we simply do not know enough about the content and limitations of observations to be able to provide such guidelines. Why is this all so difficult? Largely because when we compare a model prediction with an observation, we can quantify the residual uncertainty, but we cannot disaggregate all the different sources of uncertainty (model structure uncertainty, input data uncertainty, parameter estimation uncertainty and observation uncertainty) that contribute to the residual uncertainty. Thus it becomes essential to incorporate conceptual thinking into the modelling process, so that at the very least we are able to conclude that a model generates estimates that are consistent with, and reflect, our understanding (however limited) of the catchment processes.

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