Investigating groundwater abstraction impacts on surface water levels and surface water – groundwater interaction with analytical and numerical models

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DECLARATION

I, Ntsako Victor, hereby declare that this dissertation submitted for the Master of Science Degree to the Institute for Groundwater Studies, Faculty of Science, University of the Free State, Bloemfontein, South Africa, is my own independent work. This thesis has not been submitted to any other institution of higher education. I further declare that all sources cited have been acknowledged by means of a list of references.

Ntsako Victor Maluleke

2018414074

Date: November 2023

ACKNOWLEDGEMENTS

I want to start by giving thanks to the Almighty God for giving me the constant fortitude and skill I needed to finish my master's dissertation.

Philippians 4:13: "I can do all thing through him who strengthen me".

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Last but not least, and most importantly, Ndolivhuwa (Thank you) to my parents, Thomas, and Gloria Maluleke, for their unwavering faith in me, their prayers for me, and their belief that I can be the best version of myself. I am so grateful to my sweetheart, Zwivhuya Ratshilingani, for her tremendous help with this dissertation as well as for her tolerance and support.

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ABBREVIATIONS

ANOVA	: Analysis of Variance
Df	: Degrees of Freedom
Equ.	: Equation
GW-SW	: Groundwater-Surface water
MS	: Mean Squares
NWA	: National Water Act
SDF	: Stream Depletion Factor
SS	: Sums of Squares

ABSTRACT

Groundwater is an important source of water for many human needs, including agriculture, water supply and industry. However, the use of groundwater may have adverse consequences. Various studies have been conducted to analyse the effect of groundwater pumping on nearby streams, and many analytical models and numerical models have been developed for various aquifer and stream conditions. Numerical models are time-consuming and require a significant amount of data input but can simulate complex systems. On the other hand, analytical models are suitable to use since they are ease to use, and have minimum data requirements and exact solutions, but they are limited to less complex problems. Streamflow impacts from groundwater pumping continue to be one of the main issues with groundwater resource development. The primary objective of this dissertation is to investigate GW-SW interaction and the impact of groundwater abstraction on surface water bodies. A secondary objective is to then develop a conceptual model that highlights the natural system behaviour of groundwater abstraction in nearby streams to evaluate alternative approaches for managing and understanding streamflow depletion.

The developed hypothetical model of an abstraction borehole near the stream is incorporated by realistic available data from the previous UFS/WRC test site along the Modder River, Free State, South Africa. Four analytical solutions, namely Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003) have been simulated using the developed hypothetical models (simple stream-aquifer conceptualisation to improved stream-aquifer conceptualisation) which aid in the quantification of streamflow depletion rates. To identify parameters that greatly influence streamflow depletion, parameters are varied to understand their sensitivity and effect on the rate of streamflow depletion. Furthermore, statistical analysis is used to visualise parameters that greatly influence streamflow depletion using ANOVA (Analysis of Variance). Groundwater abstraction is found to be the most sensitive parameter that greatly affects the streamflow depletion; therefore, groundwater abstraction rates should be monitored, and data collection should be accurately acquired to give out the best recommendation as this can greatly impact the surface water. Distance separation is also considered sensitive and greatly influences streamflow depletion. The period of pumping is less sensitive due to the fact that only time simulation is being extended with no variation of any parameter. The sensitivity analyses improve the understanding of how the analytical models simulate streamflow depletion rates and provide information to water managers on which parameters are most sensitive and require focused data collection.

Numerical models are applied and designed to represent the analytical solutions to assess how the solution would be implemented in a numerical model. Two approaches have been used within this dissertation namely, 1) analytical model condition which aims at creating a numerical model to simulate analytical condition as closely as possible to understand the implication of the analytical model assumptions and 2) complex numerical model conditions considering a real-world complex numerical model and assess how analytical solutions are able to represent streamflow depletion. The approximation in the numerical model, limited aquifer extent of the numerical model domain, and differences in parameter and boundary condition applied are the most significant factors that cause differences between the analytical and numerical model results. Analytical solutions overestimate and underestimate streamflow depletion when addressing a complex setting, although they fail to address complex hydrogeological settings, these analytical solutions can act as conservative tools which aid water managers in decision-making on the quantification of streamflow depletion and its mitigation.

This study's thorough investigation using different models highlights the complexities of groundwater pumping's impact on streamflow. By examining factors like borehole distance, aquifer hydraulic parameters, and model conceptualisations, it exposes limitations in simplistic analytical models while emphasising the complex nature of aquifer-stream dynamics. Understanding time delays, geological variations, and model assumptions provides crucial insights for water resource management. These findings offer valuable guidance for balancing groundwater use and preserving streamflow, urging a holistic approach integrating the strengths of numerical simulations. This research advocates sustainable strategies to address streamflow depletion, promoting responsible groundwater utilisation for the future.

Chapter 1 INTRODUCTION

Groundwater is a vital water source catering to various human needs such as agriculture, water supply, and industry. However, its utilisation comes with potential adverse consequences. The hydrological cycle tightly intertwines groundwater and surface water, making them hydraulically connected (Winter et al., 1998). Groundwater discharge to surface water constitutes a significant portion of total streamflow. Pumping groundwater from an aquifer connected to a surface water system can diminish flows and, in certain instances, induce streamflow into the underlying aquifer (induced infiltration). This phenomenon has become a critical water-resource management issue as excessive groundwater abstraction can deplete surface water, impacting water rights and harming aquatic ecosystems.

Numerous studies have explored the impact of groundwater pumping on nearby streams, resulting in the development of both analytical and numerical models tailored to diverse aquifer and stream conditions (Barlow and Leake, 2012; Rathfelder, 2016). While numerical models are time-consuming and demand substantial data inputs, they excel at simulating intricate systems. Conversely, analytical models, with their ease of use, minimal data requirements, and instantaneous solutions, are suitable for less complex problems.

Despite extensive research on groundwater abstraction and stream flow depletion, there remains a gap in understanding the comparative effectiveness and applicability of different models under varying conditions. The literature lacks comprehensive studies that evaluate multiple models simultaneously to provide insights into their relative strengths and weaknesses in predicting streamflow depletion. Additionally, while many studies refer to the impact of groundwater pumping, there is often a lack of citations, leading to questions about evidence supporting these claims.

Addressing these gaps is crucial for several reasons. A deeper understanding of how different models perform can guide water resource managers in selecting appropriate tools for their needs. This can lead to more effective management strategies that balance groundwater use with the preservation of streamflow, thereby protecting aquatic ecosystems and ensuring sustainable water use. By identifying the most accurate and reliable models, policymakers can better predict and mitigate the adverse effects of groundwater abstraction on surface water bodies.

This study aims to bridge the knowledge gap by systematically comparing the performance of various analytical models and numerical models in predicting streamflow depletion. By doing so, the study aims to provide a clearer picture of the conditions under which each model is most effective. The hypothesis is that different models will vary in their accuracy and reliability depending on the specific hydrogeological conditions and parameters used in the simulations.

The primary objective of this dissertation is to investigate the impact of groundwater abstraction on surface water bodies, in terms of streamflow depletion. A secondary objective is to develop a conceptual model illuminating the natural system behaviour of groundwater abstraction in nearby streams, aiming to evaluate alternative approaches for managing and understanding streamflow depletion.

1.1 Aim and objectives

The main aim of the study is to investigate groundwater abstraction impacts on streamflow depletion with analytical and numerical models to facilitate an improved understanding and utilisation of these methods.

The aim is planned to be achieved using the following objectives:

- Conducting a literature review on streamflow depletion and the influencing factors.
- Review analytical and numerical methods for quantifying streamflow depletion.
- Develop hypothetical conceptual models to evaluate selected analytical models.
- Analyse analytical solutions of streamflow depletion with STRMDEPL08 analytical computer program tool.
- Evaluate how varying analytical modelling complexities (depending on varying aquifer properties) affects streamflow depletion using four analytical solutions.
- Perform sensitivity analysis to assess the influencing factors on streamflow depletion.
- Perform statistical analyses to investigate parameters which greatly influence streamflow depletion.
- Developing numerical models to further assess the analytical models for streamflow depletion.

1.2 Dissertation structure

Chapter 2 provides an overview of streamflow depletion and how it can be modelled using both analytical solutions (from simple stream-aquifer conceptualisation to improved stream-aquifer conceptualisation) and numerical models.

Chapter 3 describes the methodology applied to the dissertation.

Chapter 4 comprises of the hypothetical case and the parameters incorporated into the hypothetical model which aid in quantifying stream depletion under different scenarios. To further ensure that the hypothetical model is realistic, parameters (available data) from selected study area have been used.

Chapter 5 describes the investigation of streamflow depletion using simple to improved analytical solutions to analyse how changing to improved analytical solutions and how varying parameters affect streamflow depletion. The streamflow depletion will be tested by using four different analytical solutions using STRMDEPL08 computer-based programmes, namely Jenkins (1968), Hantush (1965), Hunt (1999), and Hunt (2003). Sensitivity analysis is used to improve the understanding of how parameters incorporated into the hypothetical model influence streamflow depletion due to the variance of the data set.

Chapter 6 outlines numerical models to simulate analytical conditions as closely as possible to understand the implications of the analytical model assumptions and how analytical models represent complex real-world problems.

Chapter 7 outlines in detail the summary, conclusion, and recommendations made from the results of this dissertation.

Chapter 2 STREAMFLOW DEPLETION OVERVIEW

Streamflow reduction (or depletion) caused by groundwater abstraction has become a significant water-resource management issue because of the detrimental impact that streamflow reduction or rather streamflow depletion can have on the aquatic ecosystem, quantity and quality of streams and rivers (Barlow and Leake, 2012).

Streamflow depletion is a term where two processes are combined, in both losing and gaining streams adjacent to groundwater abstraction (Barlow and Leake, 2012) namely:

- Induced infiltration when the reduction in groundwater level (large pumping rates) near the hydraulic connected stream area is sufficient enough such that the stream-aquifer interface's hydraulic gradient will be reversed as shown in Figure 2-1,
- As well as captured groundwater that would otherwise have discharged to stream due to cone of depression expanding outward from the well during pumping as shown in Figure 2-2.

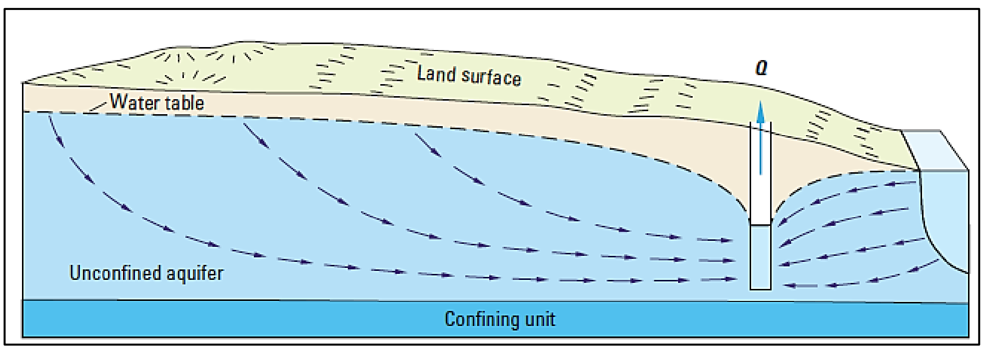


Figure 2-1 Effects of pumping rate from hypothetical unconfined aquifer that may be sufficiently large to result in groundwater to flow from the stream to aquifer (i.e. induced infiltration) (Alley et al., 1999).

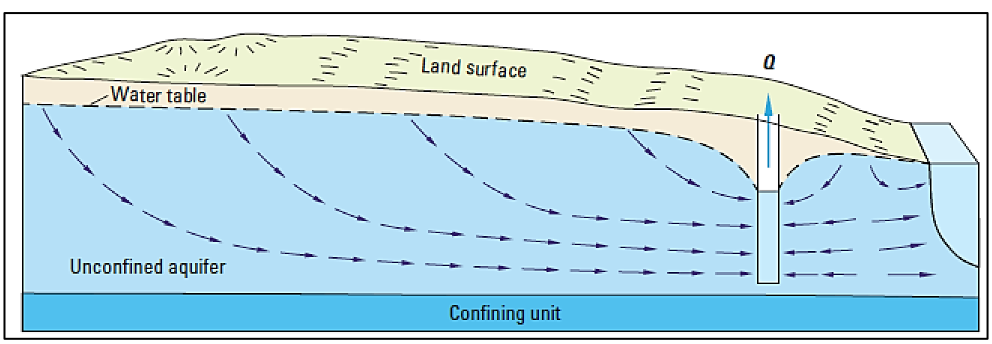


Figure 2-2 Effects of pumping rates from a hypothetical unconfined aquifer, in which the pumping well starts to capture groundwater that would have otherwise discharged to the stream as a cone of depression extends outward from the well (Alley et al., 1999).

Streamflow depletion can result from groundwater abstraction from wells near connected losing stream through direct capturing of groundwater that would otherwise have discharged to the stream as shown in Figure 2-2. It is important to note that, this occurs if the stream and aquifer are hydraulically connected by a fully saturated zone (Rathfelder, 2016). Pumping can easily cause induced filtration near losing stream reaches since the flow gradient is already oriented from stream to aquifer. as shown in Figure 2-1 (Brunner et al., 2011). In disconnected streams, groundwater abstraction can extend the length of the stream reach that is disconnected to the groundwater table, which can reduce streamflow to downstream reaches (Rathfelder, 2016).

2.1 Factors that control the time response of streamflow depletion

Change from aquifer storage during groundwater abstraction to streamflow depletion can be a slow process. This slow process depends on the distance of the pumping well to a nearby stream and the rate at which discharge of groundwater to the stream can be captured (which is termed hydraulic diffusivity of an aquifer), these two are the most important variables in controlling the time response of streamflow depletion (Barlow and Leake, 2012).

During pumping of a hypothetical well in a nearby stream, the only source of groundwater to the well immediately after pumping commences can be from groundwater released from aquifer storage. After some time, the contribution of water will be from the depletion of nearby stream (Barlow and Leake, 2012). When the well pumps for an extended period of time, a new equilibrium will be reached. In which there will be no further contribution from groundwater storage, but rather source of water pumped will eventually be from stream capture or depletion (Rathfelder, 2016) as shown in Figure 2-3.

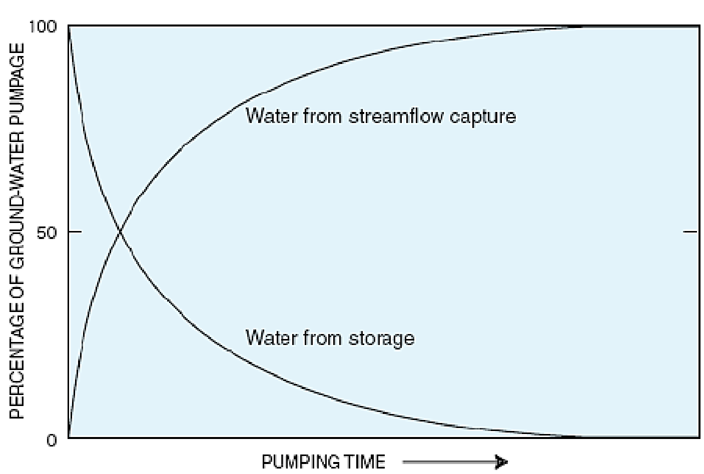


Figure 2-3 Diagrammatic representation of how the water source changes over time in response to pumping (Alley et al., 1999).

When a new equilibrium is reached, the well's pumping rate equals the amount of stream depletion, thus the water level stops declining in response to pumping and the cone of depression stops expanding. (Barlow and Leake, 2012).

2.2 Factors affecting streamflow depletion

Streamflow response to pumping can vary over time, from rapid response in minutes to delayed response processes in years or decades. Factors that are mainly responsible for affecting the rate of streamflow depletion are as follows (Barlow and Leake, 2012):

- Horizontal and vertical distance of wells from the streams,
- Hydraulic parameter of the aquifer (i.e. transmissivity and storativity), and
- Permeability of streambed sediments.

Key parameters which determine stream depletion effects as briefly discussed above, the key detailed parameters which indicate stream depletion effects are as follows (Pattle Delamore Partners Ltd, 2000):

- Q - abstraction rate from a well.
- L - the separation distance between the well and stream.
- T - transmissivity of the aquifer (i.e. the rate at which water is transmitted through the aquifer or measure of how permeable the aquifer is in transmitting water through the aquifer).
- S - storage coefficient of the aquifer (i.e. total volume of water that the aquifer releases or stores per horizontal surface and per unit change in head normal to that surface, or it is rather how much water is released from pore space of aquifer as water pressure fall).
- λ - Streambed conductance (a measure of permeability and dimension of streambed and hydraulic gradient across the streambed).

The following is the description of these parameters and the typical range that can be used as an initial screening procedure to determine whether stream depletion effects are likely to occur in any setting.

Q **- Pumping rate (m^3/day)**

Averaging of abstraction rates over a longer time provides useful estimates of stream depletion in many cases. Due to the principle of superposition to stream depletion rates, the effect of pumping at regular intervals can be simulated by the addition of effects from a sequence of pumping and recovery. Typical pumping rates that are likely to affect stream flow depletion range from 100 to 10 000 m^3/day, and it is important to note that stream depletion increases with large pumping rates. The source of pumping data can be direct measurements from pumping tests (Pattle Delamore Partners Ltd, 2000).

***L* - Separation distance (m)**

Separation distance is the distance between an abstraction well to the edge of the stream, measured perpendicular to the stream flow. Typically values can range from 1 m to 2 000 m from a well to a nearby stream. It is important to note that a smaller separation between abstraction wells and the stream will increase stream depletion. If the boreholes are placed in a distance larger than 2 000 m, it is unlikely to cause stream depletion effects. The source of data can be from a) topographical maps, b) aerial photographs and c) direct measurements from the ground (Pattle Delamore Partners Ltd, 2000).

***t* - Pumping period (days)**

The pumping period is the duration of the pumping period. A typical value would be 1 - 120 days for irrigation wells and continuous pumping for industrial and public supply boreholes. As the pumping period increases, then streamflow depletion effects decrease. The source of data can be from a) direct measurement and b) data from industrial supply requirements (Pattle Delamore Partners Ltd, 2000).

***T* - Transmissivity (m^2/day)**

This is the transmissivity of the aquifer from which the abstraction of groundwater occurs.
Mathematically can be defined as:

$$T = Kb \qquad \textbf{Equ. 2-1}$$

As shown in Equ. 2-1, *K* is the hydraulic conductivity of the aquifer and *b* is the aquifer thickness.
Typical values can range from 10 – 100 000 m^2/day. Stream depletion increases with higher values of transmissivity. Transmissivity lower than 10 m^2/day is mostly impermeable which restricts the flow of groundwater and will unlikely cause significant changes to stream depletion. The source of data can be from a) pumping data (constant rate and step-drawdown test) and b) slug test. Reliable data comes from the well under investigation, provided that the test has used a nearby observation well. Taking into account the nearby stream, Pattle Delamore Partners Ltd (2000) also suggested that there should be multiple measurements of transmissivity from surrounding wells, where it is appropriate to use the geometric mean calculation of the values (Pattle Delamore Partners Ltd, 2000).

***S*-Storage coefficient (dimensionless)**

A typical value of storage would be 0.0005 - 0.3, where stream depletion increases with a smaller value of storage coefficient. Storage values less than 0.0005 are likely to be a confined aquifer. However, confined aquifers are unlikely to be hydraulic connected to the stream to cause sufficient stream depletion, unless the stream penetrates the confining layer. The source of data can be from a) pumping test data with observation wells and b) if no sufficient data S = 0.1 is used to indicate the presence of an unconfined aquifer and is applicable when modelling transient conditions (Pattle Delamore Partners Ltd, 2000).

λ-Stream conductance (m/day)

As shown in Equ. 2-2, the stream conductance measures the vertical hydraulic conductance through the streambed to the underlying aquifer. Mathematically can be defined as:

$$\lambda = \frac{K'W}{M} \qquad \textbf{Equ. 2-2}$$

where *K'* is the hydraulic conductivity of streambed material, W is the width of streambed (m), and M is the thickness of the streambed across which K is measured (m).

Typical values range from 0.01 – 5 000 m/day. Larger values of streambed hydraulic conductance cause the streambed effects to increase. It is unlikely that stream depletion will occur when streambed conductance is less than 0.01 m/day. The source of data can be from seepage meters or infiltration tests (Pattle Delamore Partners Ltd, 2000).

To quantify variables in controlling the time response of streamflow depletion, Jenkins (1968) introduced a term as defined in Equ. 2-3 termed streamflow depletion factor (SDF). *SDF* is a measure of how rapidly a stream responds to new pumping stresses in time that would cause streamflow depletion (Rathfelder, 2016). SDF for a well pumping at a specific point in an aquifer is expressed mathematically as:

$$SDF = \frac{d^2}{D} \qquad \textbf{Equ. 2-3}$$

where, d is the shortest distance between the well and stream nearby, and D is the hydraulic diffusivity of aquifer.

From Equ. 2-2, Streamflow depletion will occur relatively quickly in response to pumping from wells with low SDF value and slow response to streamflow depletion to well with high SDF value (Barlow and Leake, 2012). The high value of hydraulic diffusivity will result in low *SDF* which then results in a fast response of streamflow depletion during pumping. Hydraulic diffusivity is a ratio of transmissivity to storativity (Rathfelder, 2016) as defined in Equ. 2-4.

$$D = \frac{T}{S}$$ **Equ. 2-4**

Hydraulic diffusivity is a measure of aquifer sensitivity to pumping (i.e. hydraulic stress). An unconfined aquifer has a larger storage capacity than a confined aquifer. Stream depletion will generally occur much more in confined aquifers than in unconfined aquifers (Barlow and Leake, 2012), as the cone of depression due to storage factor will generally be larger with pumping well and develop faster in confined aquifers rather than in an unconfined aquifer (Rathfelder, 2016).

The effect of SDF is shown in Figure 2-4, showing the effect of a well pumping for 30 days and then the pump is shut off for 30 days. The assumption is that the streambed offers no resistance to the flow of water between the stream and aquifer (i.e. λ = 100 m/day), which shows an increase in stream conductance as discussed above.

As discussed previously for settings with low *SDF*, most water from pumping is derived from stream aquifer interaction within a short time from the start of pumping. However, as the SDF becomes larger, stream depletion effects are delayed and become smaller.

In some management situations, a value of 100 days *SDF* has been used to differentiate between groundwater abstractions that are connected to the stream (*SD*F < 100 days), and abstractions less connected to the stream (*SDF* > 100 days). Figure 2-4 shows that with *SDF* greater than 100 days, the stream depletion due to groundwater abstraction is 20% less than the well abstraction rate after 30 days and the effect of shutting off pumping at SDF greater than 100 days does not have an immediate impact on the stream.

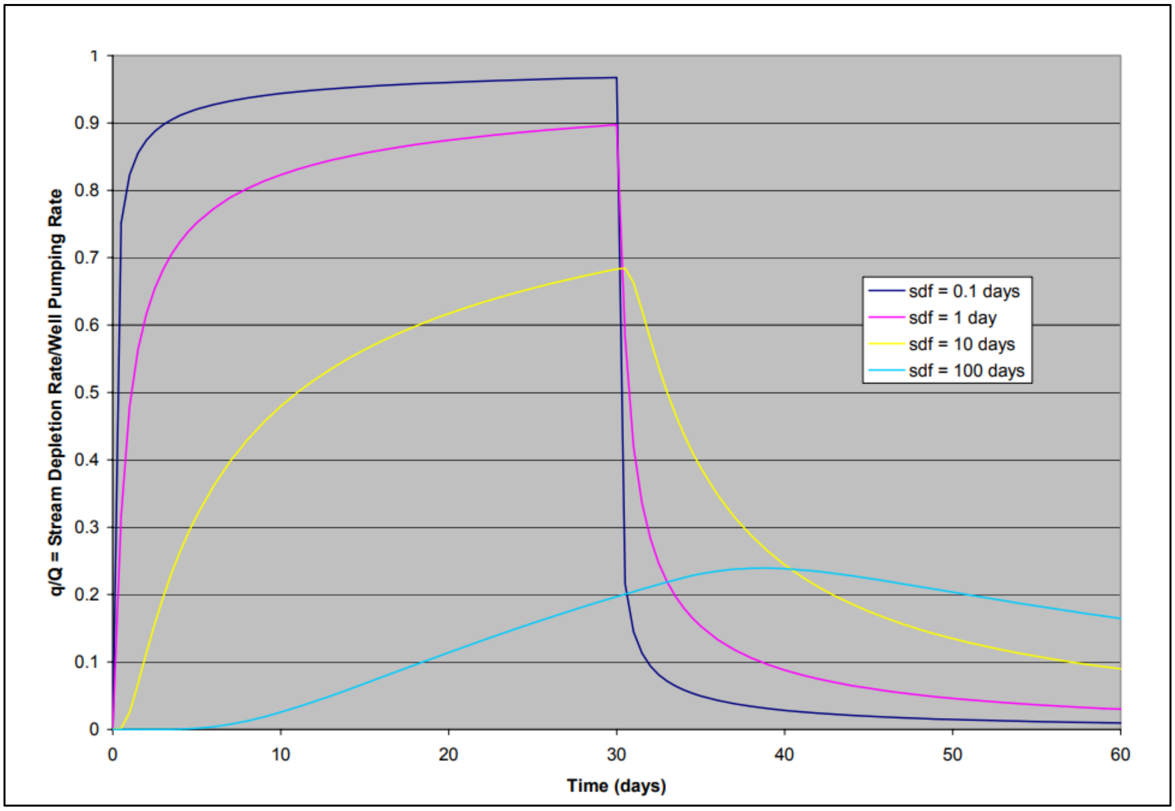


Figure 2-4 Range of SDF relative to stream depletion rate for 30 days pumping and 30 days shutoff, with stream conductance (λ=100 m/day) (Pattle Delamore Partners Ltd, 2000).

2.3 Common misconception about streamflow depletion

Comprehending the fundamentals of streamflow depletion is necessary to understand groundwater withdrawals connected to surface water.

2.3.1 Misconception 1 Streamflow depletion stops when groundwater pumping ceases.

Pumping involves removing water from the storage area surrounding the well (Barlow and Leake, 2012). A cone of depression is created during pumping periods, and it can enhance recharge to and discharge from the aquifer as shown in Figure 2-5.

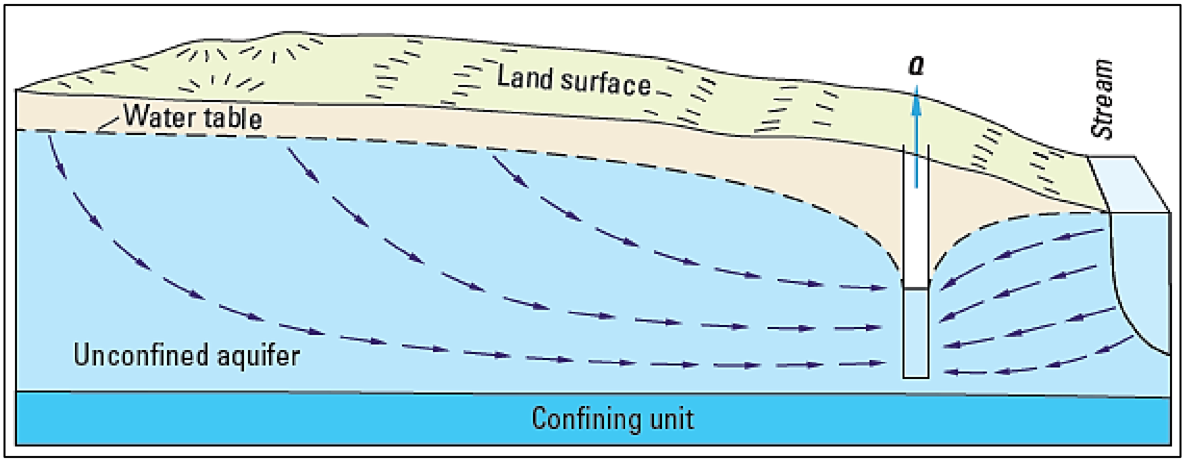


Figure 2-5 Groundwater pumping before the borehole is shut, consisting of groundwater discharge capturing and induced infiltration of streamflow (Barlow and Leake, 2012).

When pumps are shutoff shut off, groundwater levels will begin to recover to the position that existed before pumping, and this is done when the cone of depression that was created during pumping gradually fills in. During the recovery period, when pumping is stopped, groundwater that would normally flow to a stream would instead replenish the aquifer storage, which causes streamflow depletion to continue. as shown in Figure 2-6. Eventually, the system may return to pre-existing conditions, where aquifer storage and streamflow depletion are no longer changing (Figure 2-7) (Barlow and Leake, 2012).

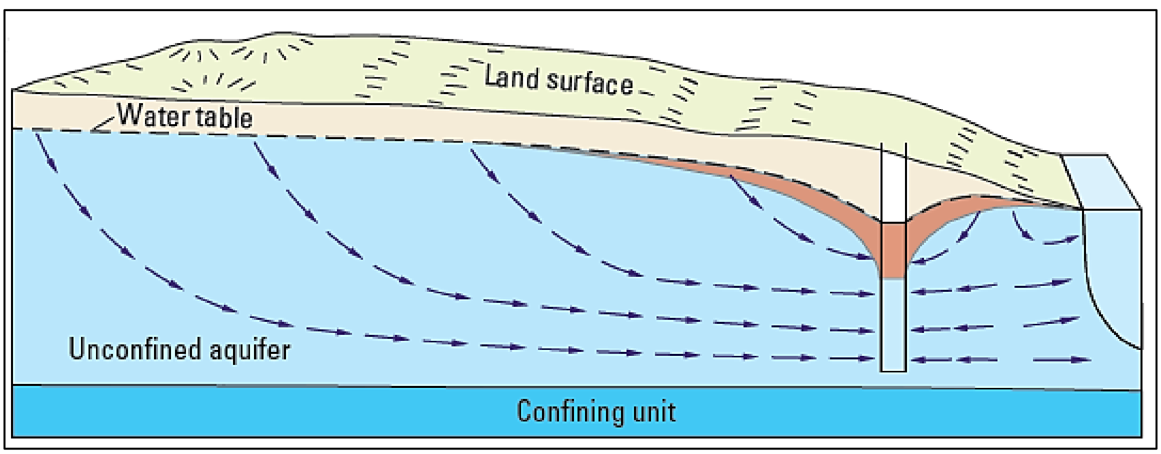


Figure 2-6 Following the cessation of groundwater pumping, groundwater levels rise and water flows into aquifer storage to replenish the cone of depression that was created during pumping. (Barlow and Leake, 2012).

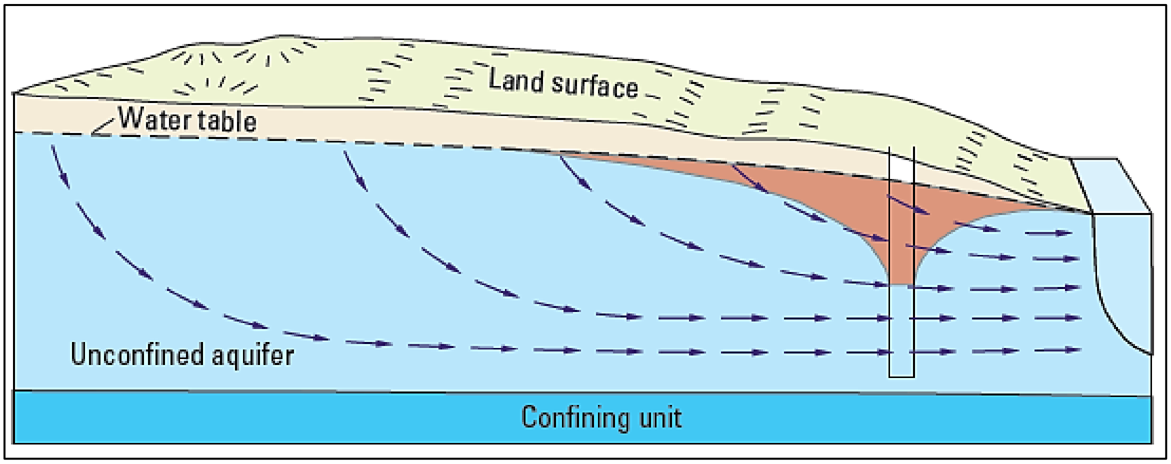


Figure 2-7 Aquifer storage and streamflow depletion remain unchanged when the system returns to its pre-pumping state (Barlow and Leake, 2012).

2.3.2 Misconception 2 Pumping from confined aquifers will not cause streamflow depletion.

Confining layers serve as a groundwater flow barrier, influence when streamflow diminishes, and determine which streams are most likely to be impacted by pumping. Pumping from the confining layer slows down the progression of depletion as compared to aquifers without confining layers (Barlow and Leake, 2012).

Confining material generally consists of silt and clay with low permeability (i.e. transmitting a small amount of water). Aquitards also called leaky aquifers, have low permeability in comparison to aquifer material.

The hydraulic connection between a confined aquifer and surface water may occur in three ways (Rathfelder, 2016):

- Confined aquifer can outcrop from the ground surface and along surface water to create a direct connection between surface water and aquifer (GW-SW interaction) and pumping from this aquifer contributes to streamflow depletion.
- Aquifers may have low permeability, but they can allow vertical movement of water between aquifer and surface water (i.e. vertical hydraulic conductivity).
- The confining layer can be discontinuous, and this creates a preferential flow path between the confined aquifer and surface water (Nielsen and Locke, 2012). Pumping from confined layers that are discontinuous can either increase or decrease the rate of streamflow depletion, as the streamflow depletion will depend on the location of discontinuous strata and pumping well to

stream (Rathfelder, 2016). The location of clay layers and well positioning to stream has an effect on decreasing or increasing streamflow depletion (Barlow and Leake, 2012).

2.3.3 Difference between streamflow reduction and depletion

Many articles have well documented that stream depletion is when there is direct pumping from a well adjacent to a stream that can cause reduced streamflow depletion (Barlow and Leake, 2012; Glove and Balmer, 1954; Hantush, 1965; Hunt, 1999; Hunt, 2003; Jenkins, 1968; Rathfelder,2016). Although reduction and depletion are applied the same in various articles, the Water Act (Act 36 of 1998) National Water Act guideline defines stream reduction as trees reducing the amount of water flowing into the river with no direct abstracting of water from the river.

2.4 South African legislative context for streamflow depletion

Historically in South Africa before the establishment of the Water Act (Act 36 of 1998), surface water and groundwater were isolated in the regulation and the policy under the previous South African Water Act (Act 56 of 1956) (Tanner, 2013). Groundwater could be used by borehole owners with no restriction, and it was not included in the national resource estimations. Surface and groundwater practitioners largely worked in isolation and not often deal with interconnectivity between two resources (Hughes, 2004), even though they were aware of their significance.

The recent Water Act (Act 36 of 1998) requires that water resources be managed in a holistic and integrated manner (Tanner, 2013). The new legislation also established the maintenance of the natural flow of a river to ensure sustainability and to preserve biodiversity (Tanner, 2013). All water uses are subjected to a system of allocation general authorization and licenses (Levy and Xu, 2012). The ecological reserve applies to surface and groundwater (i.e. the impact of one can affect the other due to the hydraulic connection between the two water bodies), this encouraged surface and groundwater practitioners to begin to work together in an integrated manner on both resources (Tanner, 2013).

Many years have passed since the establishment of NWA (Act 36 of 1998) but many factors contribute to the challenge of successful implementation, which can also be referred to as gaps in the legislation. One of these is the uncertainty that surrounds the quantifying of GW-SW interaction (Tanner, 2013). The effective management of water resources where the aquifer is in hydraulic connection with the river system requires an in-depth understanding of groundwater abstraction's impact on surface water bodies.

Under the NWA (Act 36 of 1998), based on the 28 principles for water management in South Africa, the water resource is regarded as “invisible”, and there is no distinction between abstractions from a river (surface water), which may have an impact on groundwater, especially where there is a connection between groundwater and surface water (i.e. where surface water feeds the aquifers or where the aquifer contributes to the baseflow in the river). In both cases, a proper hydrological and geohydrological study is required, which will indicate whether such impact exists, and if they do, what are the risks focusing on the probability of occurrence and magnitude of the effect. If the risk of such impacts occurring is unacceptable, proper mitigation measures to prevent the risk from occurring have to be proposed and implemented to prevent the risk from occurring. The setting of buffer zones (i.e. the closeness of boreholes near a river) is a redundant water governance approach, as all impacts and systems depend on site-specific circumstances (Levy and Xy, 2012).

2.5 Modelling streamflow depletion

Groundwater and surface water are usually hydraulically connected, and as such they should be treated as one single source (Baalousha, 2016). As discussed previously, groundwater pumping in the nearby stream may capture the flow that would otherwise discharge to the stream. It is then important to manage surface water flow to protect stream habitat through proper integrated resource management (Baalousha, 2016).

This section describes how these interactions can be modelled, and coupling GW-SW interaction can be done through analytical and numerical models (Baalousha, 2016). Available methods for quantifying streamflow depletion from groundwater pumping range from easy and cost-effective analytical tools to more complex and detailed numerical modelling (Rathfelder, 2016).

2.5.1 Analytical models

Analytical solutions are based on simplified approaches for assessing stream depletion from groundwater abstraction (Rathfelder, 2016). It can be from a single well. Several analytical solutions have been developed for groundwater flow equations, for the estimation of time-varying rates of streamflow depletion caused by groundwater abstraction (Barlow and Leake, 2012). Analytical models are the simplified representation of real-world and can provide a simple and quick estimation of streamflow depletion by nearby stream to pumping wells (Rathfelder, 2016), partly because they are simple to utilise with user-friendly software or spreadsheets, with less specific data requirements to implement than numerical models (Barlow and Leake, 2012).

Although analytical models are based on simplified representations of the real world, there exist modified analytical equations that represent the real-world problems of stream depletion. Historically analytical models have been used by regulatory agencies for making management of water rights (Sophocleous et al., 1995).

Numerous studies developed analytical models for quantifying the effect of groundwater abstraction on nearby streams. Analytical models are useful tools for gaining insights into system behaviour and for estimating depletion under idealised conditions (Rathfelder, 2016).

2.5.2 Numerical models

Numerical models are more popular than analytical models due to their ability to handle many hydrogeological complexities, especially on a catchment scale (Baalousha, 2016). It is also important to note that these hydrological complexities affect the timing of streamflow depletion which includes non-uniform soil, different stream geometry and non-ideal aquifer geometry (Barlow and Leake, 2012). For complex problems, numerical models are a more robust approach tool in evaluating the effect of pumping on stream flow (i.e. involves approximation of partial differential equations).

2.5.3 Analytical versus numerical models

Analytical models are the most useful tools for gaining insights into system behaviour and estimating streamflow depletion into idealised conditions. However, these analytical solutions remain unable to address many of the complicated factors such as aquifer heterogeneity, multiple streams or complex stream geometry and non-uniform soil and streambed properties (Barlow and Leake, 2012). These complex factors can have substantial effects on streamflow depletion that limit the use of analytical tool applications to address such conditions, particularly in the case of multiple wells pumping simultaneously (Barlow and Leake, 2012). For these complex conditions, numerical models are used and provide a robust (numerical approximation) approach for evaluating groundwater pumping effects on surface water flows (Rathfelder, 2016). Despite solving more complex problems, the use of numerical models is complicated by the significant time and effort required to create, calibrate, and validate the models (Li et al., 2016).

Analytical models of streamflow depletion assume a system with a single straight (Figure 2-8) stream, yet in a numerical approach (real-world setting), the system includes a stream with tributaries, each with complex stream geometry and stream network (Barlow and Leake, 2012).

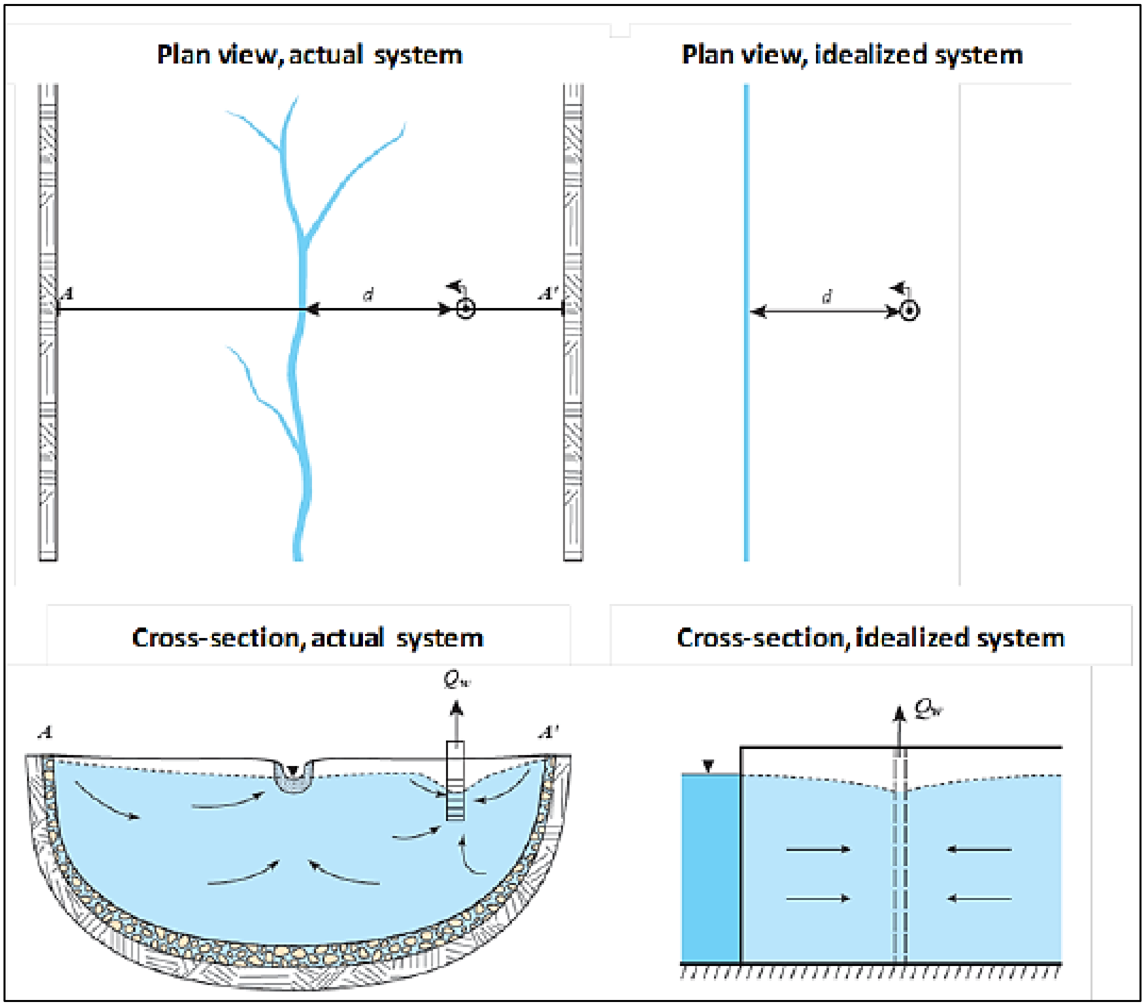


Figure 2-8 Hypothetical aquifer-stream conceptualization used in the Glover model (*d* is the distance from the well to the nearest stream and Q_w is the pumping rate at the well) (Rathfelder, 2016).

Analytical models would not be able to account for these effects of irregular aquifer dimensions. When faced with these complexities (Barlow and Leake, 2012), the numerical modelling tools are the most powerful tool to address complex physical and hydrogeological conditions for understanding streamflow depletion from groundwater pumping and to also test the assumptions made by various analytical solutions (Sophocleous et al., 1995).

Some common complexities in the real system that require a numerical modelling approach include (Barlow and Leake, 2012):

- Irregular aquifer dimensions (lateral and vertical boundaries of the aquifer).
- Irregular geometry of stream and variable surface water features such as meandering stream, seasonal streamflow, and stream network.

- Heterogeneous aquifer properties with varying parameters.
- Time-varying pumping schedule.
- Seasonally dependent recharge, evapotranspiration and irrigation returns (irregular sources that are not constant over time).
- Difference in boundary conditions at different scales.
- If there is uncertainty about the potential impacts or if regulatory requirements demand a quantitative analysis, then a numerical model becomes essential to provide a more detailed and precise assessment.

Chapter 3 METHODOLOGY

This dissertation will focus on the development of a hypothetical model for four analytical solutions i.e. Jenkins (1968), Hantush (1965), Hunt (1999), and Hunt (2003), aiming to develop an understanding of streamflow depletion. The analytical solutions will then be further assessed within the frame of numerical groundwater modelling.

3.1 Analytical models

3.1.1 Foundational analytical solutions

Theis equation is the most well-known and earliest solution of unsteady state groundwater flow (Theis, 1941). Theis developed a hypothetical stream-aquifer system as shown in Figure 2-8 (Rathfelder, 2016). The full set of assumptions for analytical solution are as follows (Jenkins, 1968):

- Aquifer is homogeneous, isotropic and extends infinitely in all directions from the stream.
- The aquifer is confined. Transmissivity and saturated thickness are not time-dependent. The solution is regularly applied to the unconfined aquifer, which can be assumed that the drawdown caused by pumping is small compared to the initial saturated thickness of the aquifer.
- Streams that form the boundary with the aquifer are straight, fully penetrated the thickness of the aquifer, infinitely long, remain flowing at all times and are in perfect hydraulic connection with the aquifer (thus, the solution assumes no streambed and streambed material do not impede the flow between stream and aquifer).
- The temperature of stream and aquifer is homogenous and does not vary with time, this is a significant assumption since the variation of temperature affects the hydraulic conductivity of streambed and aquifer sediments.
- Pumping is from a single well that pumps from the full saturated thickness of the aquifer at a constant rate and maximum discharge.
- Stream is the only source of possible recharge to the aquifer.

3.1.2 Jenkins (1968) model

The simplest and most widely applied analytical solution of streamflow depletion was expressed by Glover and Balmer (1954) and Jenkins (1968) further extended the model by introducing SDF (stream depletion factor) which accounts for proportions of streamflow depletion relative to total groundwater pumping over time. Jenkins (1968) modified the Glover solution by extending the model to determine the response of streamflow depletion after a pump has stopped (i.e. residual streamflow depletion). To then calculate residual streamflow depletion, is the difference between the rate of streamflow depletion from a pumping well that is pumping continuously and an imaginary injection well at the same location (Rathfelder, 2016). The solution is easier to solve and expressed in different forms and developed independently (Rathfelder, 2016). The initial solution is for a system with a stream that fully penetrates the aquifer with no streambed resistance between the stream and the aquifer and may be expressed as (Glover and Balmer, 1954; Jenkins, 1968)

$$Q_s = Q_w \, erfc(z)$$ **Equ. 3-1**

The Jenkins solution as defined mathematically in Equ. 3-1, provides an expression of stream depletion flow rate as a function of time (defined mathematically as Q_S), and is equal to the product of constant flow rate abstracted at well, Q_w, and mathematical function, which is called complementary error function, $erfc(z)$.

Variable Z in Equ. 3-1 is equal to $\sqrt{\frac{d^2S}{4Tt}}$, but as described in Equ. 2-4, $\frac{S}{T}$ is the inverse of the hydraulic diffusivity of the aquifer($D = \frac{T}{S}$). Then Jenkins equation is mathematically defined as:

$$\frac{Q_s}{Q_w} = erfc\left(\sqrt{\frac{d^2S}{4Tt}}\right)$$ **Equ. 3-2**

Where,

Q_s is stream depletion flow rate (L^3/T);

Q_w is constant flow rate abstracted at well (L^3/T);

S is the aquifer storage coefficient (dimensionless);

T is the aquifer transmissivity (L^2/T);

t is the duration of the pumping period (T);

d is the shortest distance between the well and stream edge (L), based on the approach that the stream is assumed to be a straight-line feature.

For example, the Jenkins model for groundwater pumping in an unconfined aquifer, as explained by the stream depletion factor ($SDF = \frac{d^2}{D}$), the rate of groundwater pumping causes stream depletion to increase as the distance between the well and the stream decreases and *SDF* becomes low, as shown in Figure 3-1. No matter how far the well distance to the stream is, the well location will eventually drive all water from the stream (Rathfelder, 2016), but the difference will be the rate of streamflow depletion. The rate of streamflow depletion depends on SDF (Rathfelder, 2016).

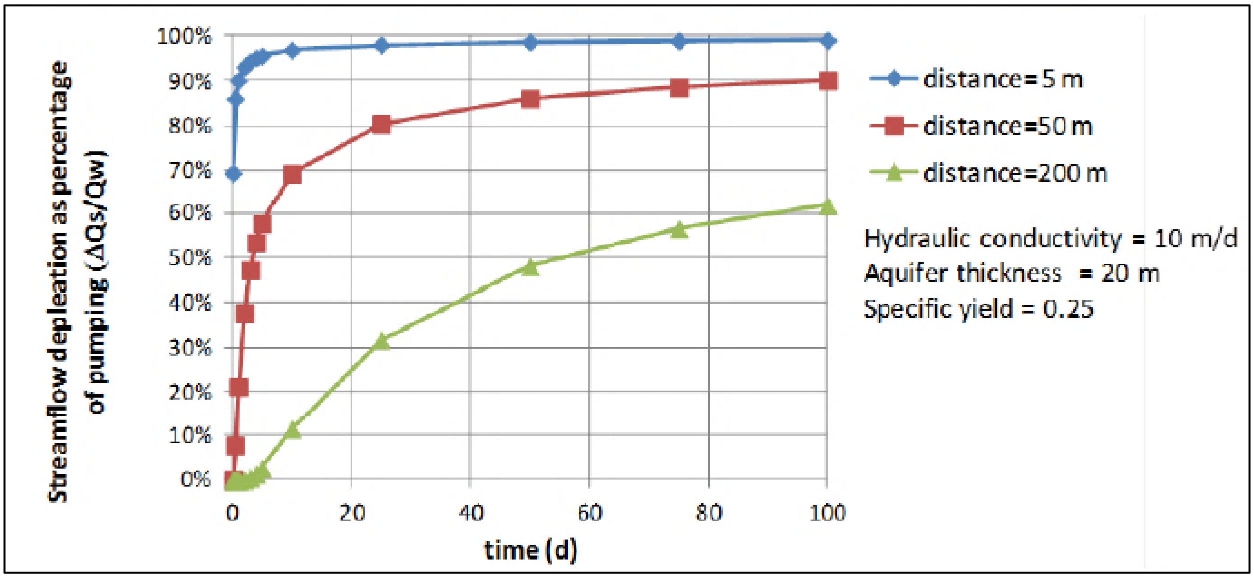


Figure 3-1 Graphical representation showing the Jenkins model solution, as the distance between the well and stream decreases the rate of stream depletion increases (Rathfelder, 2016).

3.1.3 Hantush (1965) model

Since the Glover equation does not account for hydrological complexities (Rathfelder, 2016), Several authors have shown that the assumption of a perfect hydraulic connection between the aquifer and stream (penetrating the full thickness of the aquifer) can result in errors in determining the time response of streamflow depletion (Barlow and Leake, 2012).

Since streambed material has lower hydraulic conductivity and only partially penetrates the stream, the analytical model was expanded to account for flow resistance at the stream-aquifer interface to address these inaccuracies. (Rathfelder, 2016). These are the two conditions that strongly affect the prediction accuracy of streamflow depletion.

Hantush (1965) extended the Glover solution to include resistance to flow at stream-aquifer boundary (i.e. effects of a semi-impervious streambed layer). Hantush's (1965) model was still based on a fully penetration stream, with low permeability streambed sediments than adjacent aquifers as shown in Figure 3-2. All Other assumptions of Glover's solution still hold.

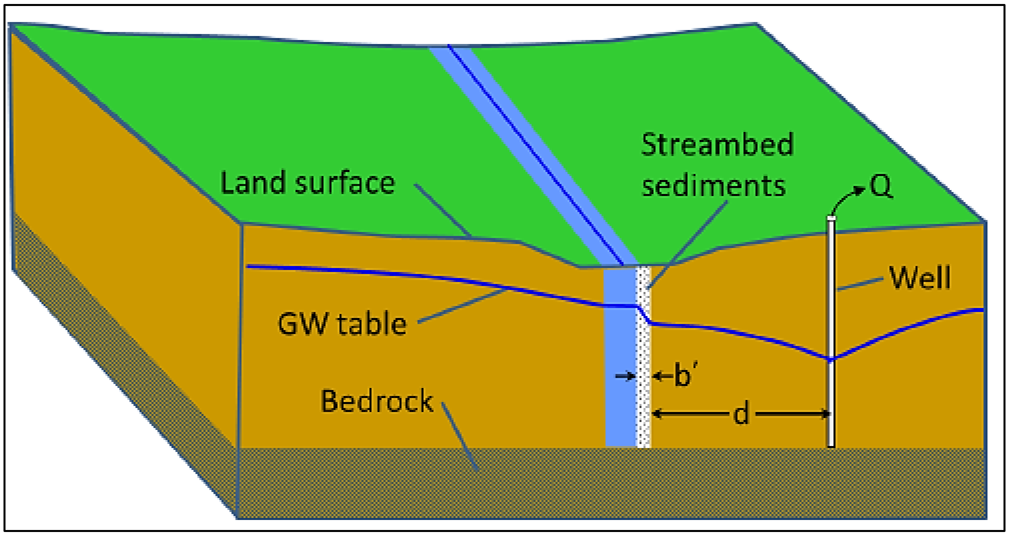


Figure 3-2 GW-SW interaction (aquifer-stream conceptualization) in the Hantush model (Rathfelder, 2016).

Mathematically based on the conceptualization, Hantush developed the following analytical solution:

$$\frac{Q_s}{Q_w} = erfc\left(\sqrt{\frac{d^2S}{4Tt}}\right) - exp\left(\sqrt{\frac{Tt}{SL^2}} + \frac{d}{L}\right) erfc\left(\sqrt{\frac{Tt}{SL^2}} + \sqrt{\frac{Sd^2}{4Tt}}\right)$$ **Equ. 3-3**

where, all terms are previously defined with an addition of the following term:

L is the streambed leakance that represents the effect of streambed resistance to flow due to streambed material.

Streambed leakance is expressed as:

$$L = \frac{K}{K'} b'$$ **Equ. 3-4**

where,

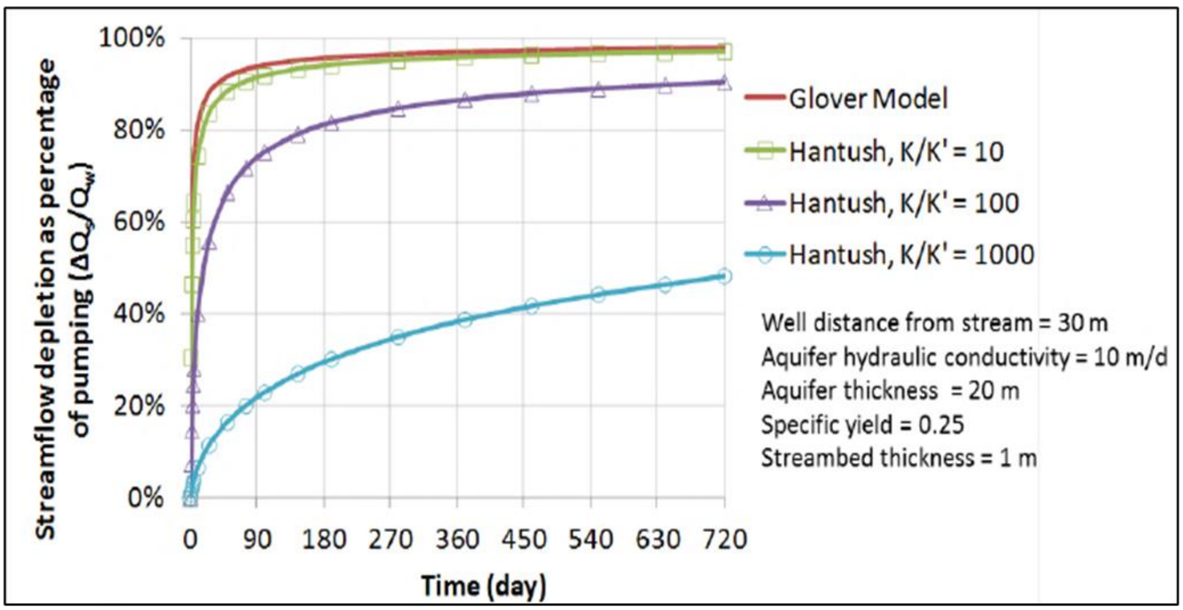
K is hydraulic conductivity of aquifer;

K' is the hydraulic conductivity of streambed material, which is lower than the hydraulic conductivity of adjacent aquifer, and;

b' which is the thickness of the streambed as illustrated in Figure 3-2.

If the streambed leakance is zero ($L = 0$) then Hantush's (1965) solution is equivalent to Glover's solution, and this occurs when streambed thickness is relatively small and can be ignored or when the hydraulic conductivity of the streambed is equal to the hydraulic conductivity of the adjacent aquifer (Rathfelder, 2016).

An example illustrated in Figure 3-3 illustrates the results where the streambed conductivity affects the rate of streamflow depletion. As the conductivity difference between streambed and adjacent aquifer increases, the rate of resistance in streamflow depletion increases (Rathfelder, 2016). It is important to note that streambed conductivity is four to five orders of magnitudes lower than aquifer conductivity as shown in Figure 3-3 ($K > K'$) (Calver, 2001).



Explanation:

Aquifer properties		Streambed Properties		Conductivity ratio (K/K')
Conductivity (k) (m/d)	Representative texture	Conductivity (K') (m/d)	Representative texture	
10	Coarse sand	1	Fine sand	10
		0.1	Silty sand	100
		0.01	Silt	1000

Figure 3-3 Graphical representation of the Hantush model for different hydraulic conductivity ratios (Rathfelder, 2016).

3.1.4 Hunt (1999) model

Hunt (1999) further extends the solution of Hantush (1965) to accommodate both streambed resistance and a partially penetrated stream, instead of fully penetrated the aquifer, as shown in Figure 3-4. An important assumption is that the groundwater level in the aquifer at the stream must remain above the streambed so that the stream is not disconnected from the underlying aquifer (Barlow and Leake, 2012). Thus, this equation can only simulate a gaining stream. Model conceptualisation is large enough such that the stream width is zero and the aquifer has an infinite extent (Rathfelder, 2016).

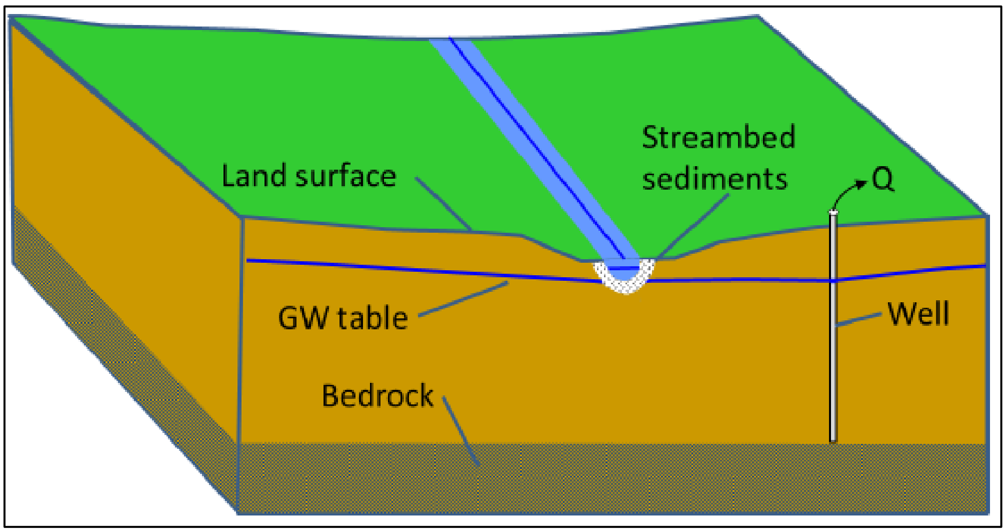


Figure 3-4 GW-SW interaction (aquifer-stream conceptualisation) in the Hunt model (Rathfelder, 2016).

Hunt (1999) derived the following mathematical expression (Equ. 3-5) in which λ (lambda) is an additional term that represents streambed conductance:

$$\frac{Q_s}{Q_w} = erfc\left(\sqrt{\frac{d^2S}{4Tt}}\right) - exp\left(\sqrt{\frac{\lambda^2}{4ST}} + \frac{\lambda}{2T}\right) erfc\left(\sqrt{\frac{\lambda^2 t}{4ST}} + \sqrt{\frac{Sd^2}{4Tt}}\right) \qquad \textbf{Equ. 3-5}$$

3.1.5 Hunt (2003) model

Hunt (2003) is the improved analytical solution from Hunt (1999) that provides advances in terms of improved representation of hydrogeologic systems and depth insights into specific conditions (Rathfelder, 2016). This analytical solution is for a system with a stream partially penetrating the aquitard with pumping from an underlying leaky aquifer (Reeves, 2008). It is important to note that the following assumption holds where the model conceptualisation is similar to Hunt (1999), the distance between the abstracted borehole and the modelled stream is large enough such that the width of the stream is zero and the aquifer has an infinite extent (Rathfelder, 2016).

Hunt (2003) includes more information requirements, namely (Figure 3-5);

- Aquifer and aquitard thickness.
- Aquifer and aquifer hydraulic conductivity.
- Aquifer and aquifer storativity.
- Streambed thickness and its hydraulic conductivity.
- Stream width.
- Borehole distance from the modelled stream.

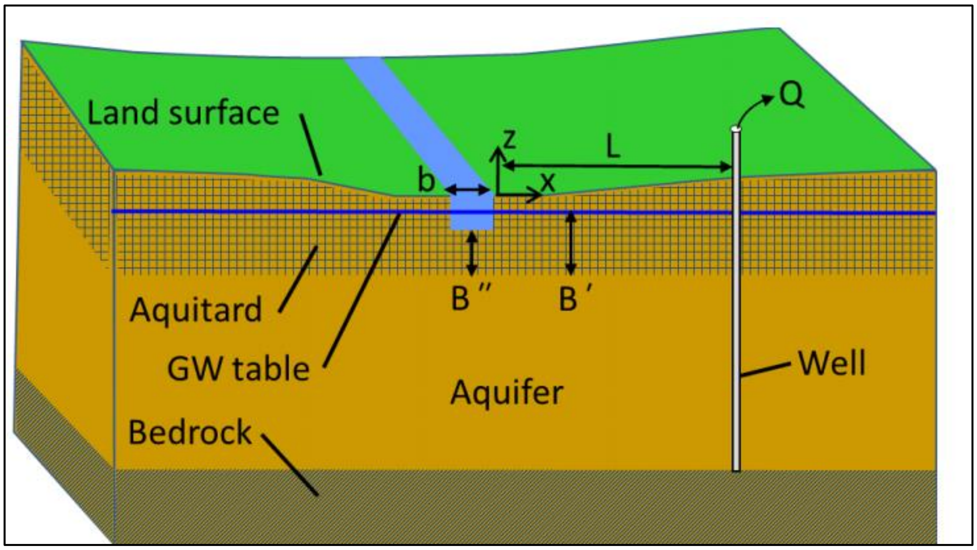


Figure 3-5 GW-SW interaction (semi-confined aquifer and stream conceptualisation) in the Hunt (2003) model (Rathfelder, 2016).

Hunt (2003) plotted stream depletion against time and found a delayed aquifer response. At an early time of pumping, the rate of streamflow depletion is small as pumped water comes mainly from aquitard storage. As time increases, pumped water comes from both aquifer storage and stream and eventually causes the rate of streamflow depletion to increase (Reeves, 2008). At a large pumping rate, stream flow depletion approached the pumping rate. The analytical solution for a stream penetrating the aquitard with pumping from the underlying leaky aquifer is expressed as follows (Hunt, 2003):

$$\frac{Q_s}{Q_w} = \left[erfc\left(\sqrt{\frac{d^2S}{4Tt}}\right) - exp\left(\sqrt{\frac{\lambda^2 t}{4ST}} + \frac{\lambda d}{2T}\right) erfc\left(\sqrt{\frac{\lambda^2 t}{4ST}} + \sqrt{\frac{Sd^2}{4Tt}}\right) - \lambda \int_0^1 F(\alpha,t)G(\alpha,t)d\alpha\right] \quad \textbf{Equ. 3-6}$$

In this equation, $F(\alpha, t)$ and $G(\alpha, t)$ are the function that may be calculated as (Hunt, 2003):

$$F(\alpha,t) = exp\left(-\frac{Sd^2}{4Tt \propto^2}\right)\sqrt{\frac{Tt}{Sd^2\pi}} - \frac{\alpha t\lambda}{2Sd}\exp\left(\frac{\lambda d}{2T} + \frac{t\alpha^2\lambda^2}{4ST}\right) erfc\left(\frac{\alpha\lambda}{2}\sqrt{\frac{t}{ST}} + \frac{d\sqrt{S}}{2\alpha\sqrt{Tt}}\right) \quad \textbf{Equ. 3-7}$$

$$G(\alpha,t)=\frac{1}{2}\left[1-e^{-(a+b)}Io\left(2\sqrt{ab}\right)+\left(\frac{b-a}{a+b}\right)\sum_{n=0}^{\infty}\binom{2n}{n}P(2n+1,a+b)\left(\frac{\sqrt{ab}}{a+b}\right)^{2n}\right] \quad \textbf{Equ. 3-8}$$

where,

$$a=\frac{(K''/B')}{\sigma}(1-\alpha^2) \quad \textbf{Equ. 3-9}$$

$$b=\frac{(K''/B')}{S}\alpha^2 \quad \textbf{Equ. 3-10}$$

α is the variable of integration (dimensionless);
Io is the modified Bessel function of zero order (dimensionless);
$\binom{2n}{n}$ is the Binomial Coefficient (dimensionless);
$P()$ is an incomplete Gamma function (dimensionless);
K'' is the hydraulic conductivity of the aquitard (L/T);
B' is the thickness of the aquitard (L), and;
σ is the specific yield of the aquitard (dimensionless).

3.1.6 Overview

This study will focus mainly on how streamflow depletion is affected by analysing four analytical equations that address the following hydro-geological factors:

- Fully penetrating stream with no streambed resistance,
- Fully penetrating stream with streambed resistance,
- Partially penetrating stream with streambed resistance, and
- Partially penetrating stream with drawdown in the aquifer below streambed.

Although have many studies have been done on the modification of analytical solutions, for example, a system that addresses streamflow depletion in the presence of other sources. These solutions will not be discussed in this dissertation but important to appreciate their existence in addressing streamflow depletion on a complex system.

To facilitate the understanding and sensitivity analysis, parameters are varied from magnitude lower to magnitude higher. Statistical analysis using ANOVA (Analysis of Variance) is used to understand the influence of parameters in streamflow depletion.

3.2 Analytical modelling software

Analytical models discussed are easy to use and require minimal data input and assume to provide a conservative estimate. Many analytical models have been developed to address various studies on stream flow depletion, but these studies focused mainly on four analytical solutions that have been discussed above. These solutions have been modified such that it can be concluded that stream deletion is affected by many factors not just pumping rate in a simple system but also parameters like aquifer hydraulic conductivity, aquifer geometry, and strength of hydraulic connection between aquifer and surface water body. To analyse these solutions together with their effect on stream depletion, USGS computer program STRMDEPL09 will be used which accounts for all four analytical solutions to understand how streamflow depletion varies from one solution to another. STRMDEPL08 is a web interface tool that quantifies streamflow depletion by an abstracting borehole with the four analytical solutions discussed (Reeves, 2008) (https://mi.water.usgs.gov/software/groundwater/CalculateWell/index.html). This program is an extended program of STRMDEPL, a one-dimensional model using two analytical solutions namely, Jenkins (1968) and Hantush (1954) to calculate streamflow depletion by a nearby abstracting borehole. STRMDEPL incorporates solutions for stream that fully penetrates the aquifer with and without streambed resistance and the extended and modified program (STRMDEPL08) allows for solving solutions for partially penetrating stream with streambed and for a stream in an aquitard with pumping groundwater from an underlying aquifer (Reeves, 2008). Verification and performance of STRMDEPL08 was tested using both constant pumping and time-varying pumping rates.

3.3 Numerical models

Streamflow depletion may be simulated using any licensed groundwater modelling tool, provided that the program includes multiple system water balance component computations. (Barlow and Leake, 2012), including inflow to the aquifer (i.e. movement of water from area of recharge), change in aquifer storage and outflow from the aquifer to area of discharge (Barlow and Leake, 2012). For this research, the groundwater modelling software MODFLOW, with the use of the PMWIN interface, has been selected for use.

MODFLOW is used worldwide to quantify the aspect of groundwater flow also including streamflow depletion (Harbaugh, 2005). MODFLOW is a three-dimensional difference model for simulating the flow of water through porous media. This model uses a rectangular square grid shape to represent the model domain (it can be to represent aquifer and stream interaction) (Barlow and Leake, 2012). The movement of water between cells is calculated through the iterative solution to the sequence of finite difference equations (Pattle Delamore Partners Ltd, 2000).

It is beyond the scope of this dissertation to give detailed instructions on the use of MODFLOW. However, the following are the key features that must be present in MODFLOW to assess streamflow depletion effects. Packages required within MODFLOW to simulate stream depletion assessment are described below:

3.3.1 Packages

Input data for basic packages include the following characteristics (Pattle Delamore Partners Ltd, 2000):

- Defining model grid (i.e. rows, columns, and model layers).
- Time steps and stress periods for simulation (a stress period is a period where the external model boundary remains constant, within each simulation, a stress period is divided into several smaller time steps).
- Initial head distribution over the whole model within the aquifer at the start of simulation; and
- Boundary condition for the model grid.

Block-centred Flow package

Input data for the centred Flow package defines grid geometry, storage coefficient, hydraulic conductivity parameter of each cell and top and bottom elevation of the layers. These input parameters should be from the simple conceptual model (Pattle Delamore Partners Ltd, 2000).

RIVER package

River package and STREAM package are the two MODFLOW packages that can be used to specify the stream (Barlow and Leake, 2012). Both these packages identify cells in which the stream occurs, the stage height of water in the stream, the height of the streambed and the conductance of the streambed.

RIVER package specifies a constant stage height of the stream for each stress period and operates under the assumption that the stage height remain constant, even when the movement of water between aquifer and stream occurs. When the elevation of water in the stream is constant, this package is appropriate for simulation where the stream has a steady flow. If the groundwater level drops below the bottom of streambed sediment, the seepage rate to the aquifer becomes steady (Pattle Delamore Partners Ltd, 2000).

STREAM package provides an option for specifying the initial elevation of water in the stream (stage height) and stream flow at the start of the model simulation (Pattle Delamore Partners Ltd, 2000). The stage height and streamflow are determined by the movement of water in between the aquifer and the stream throughout the model simulation. If there are significant aquifer losses, the STREAM package may permit the stream to dry up (Pattle Delamore Partners Ltd, 2000). The package keeps track of the flow in a stream.

Well package

The Borehole package specifies the location of the borehole and the pumping rate as input data. A constant pumping rate is specified for each stress period.

3.3.2 Overview

Numerical models are applied and designed to represent the analytical solutions to assess how the solution would be implemented in a numerical model, two approaches are used namely:

- Analytical model condition: creating a numerical model to simulate analytical conditions as closely as possible to understand the implications of the analytical model assumptions.
- Complex numerical model conditions considering a real-world setting and assessing how analytical solutions can represent the streamflow depletion in this case.

Chapter 4 STUDY SITE CONCEPTUALISATION

The conceptual model comprises a hypothetical case of an abstraction borehole near a river to illustrate the effects of stream depletion and the analytical solutions for this. To ensure the hypothetical model is realistic, available data from a selected study area has been used to define the hypothetical conceptual model and model parameters. The selected study area is a previous UFS/WRC test site along Modder River, Free State, South Africa. Figure 4-1 shows a schematic diagram of the hypothetical model. The developed hypothetical model will form the base case, which will aid in quantifying stream depletion under different scenarios.

The aquifer parameters defined for the hypothetical model, specifically designed for streamflow depletion analysis are presented and tabulated in Table 4-1. The parameters included in the hypothetical model are designed for the analytical models selected for investigation, as discussed in Section 4.1 to Section 4.4.

4.1 Conceptual model - Jenkins (1968) model

The hypothetical conceptual model for the Jenkins (1968) model is illustrated in Figure 4-1, and the main components are discussed.

4.1.1 Distance to pumping borehole from the stream (D)

The distance between the abstraction borehole and the nearest margin of the river for the hypothetical model base case is 500 m (Table 4-1). This distance is arbitrarily selected from the average distance of boreholes from the river in the study area. From this base case, the distance will be varied to see how this parameter impacts streamflow depletion.

4.1.2 Transmissivity (T)

The transmissivity of the hypothetical conceptual model is assigned at a value of 70 m^2/d (Table 4-1). This value is based on multi-borehole aquifer experiments conducted by Gomo (2011) in the selected case study area. The transmissivity will be varied from this initial base case to determine the sensitivity of streamflow depletion to this parameter.

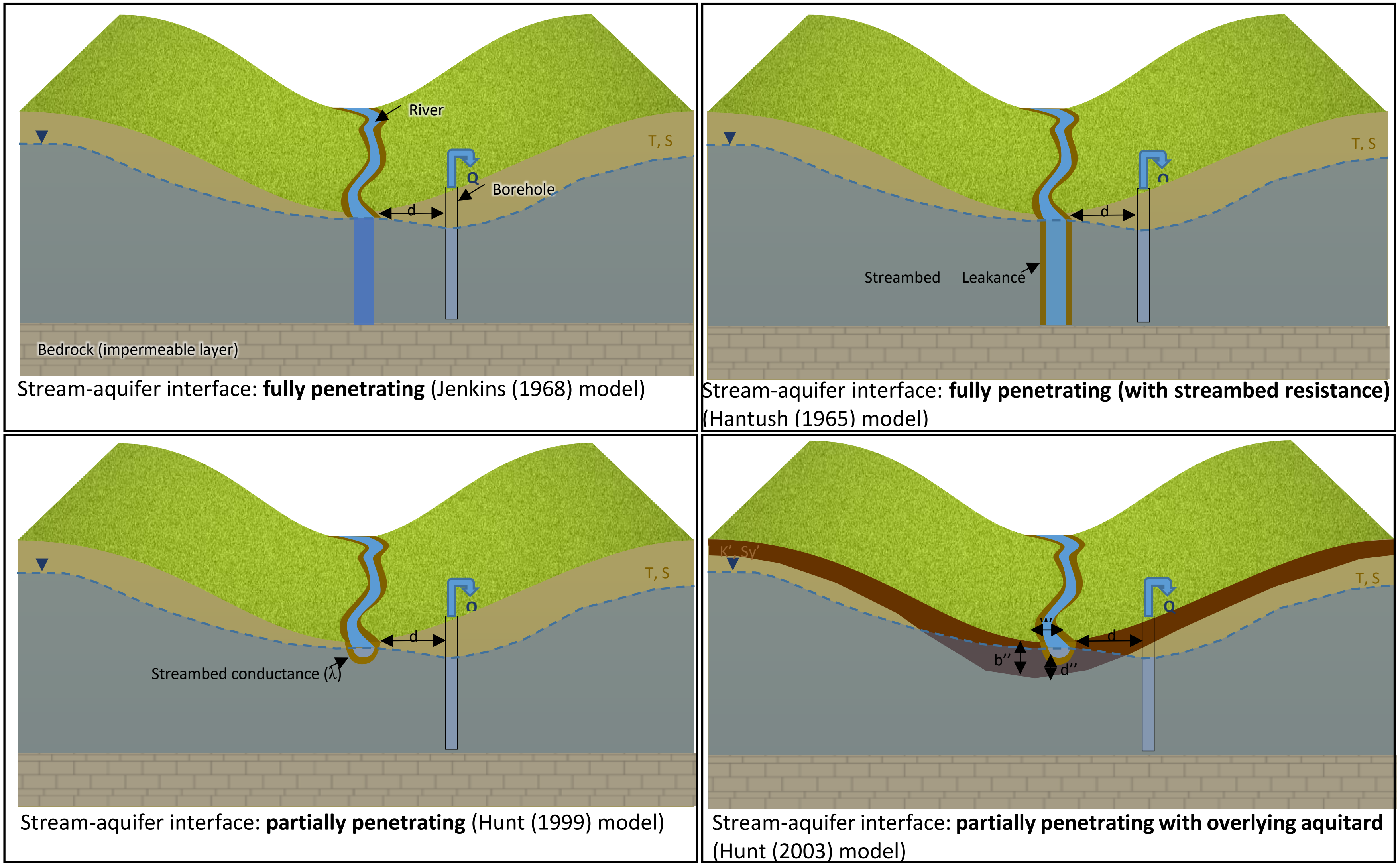


Figure 4-1: Hypothetical model depicting a simple aquifer-stream conceptualisation for streamflow depletion.

4.1.3 Storativity (*S*)

The hypothetical conceptual model will be defined using the average storativity of 0.02 (Table 4-1) to model the geohydrological setting and parameters of the study area. This initial value is based on work performed by Gomo (2011), where observation boreholes were used to conduct multi-borehole studies to establish storativity at the selected case study area. The storativity will be varied from this initial base case to determine the sensitivity of streamflow depletion to this parameter.

4.1.4 Abstraction rate (*Q*)

The defined abstraction rate for the hypothetical conceptual model is 2 l/s (Table 4-1). This value is based on the average yield of boreholes investigated in the study area by Gomo (2011). The abstraction rate should have a strong impact on streamflow depletion, and thus this parameter will also be varied in the sensitivity analysis.

4.1.5 Pumping period (*t*)

For the hypothetical model base case, the duration of the pumping period is selected at 365 days (1 year). From this base case, the duration of pumping will be varied to see how the pumping period influences streamflow depletion.

4.2 Conceptual model – Hantush (1965) model

The same parameters as discussed for the Jenkins (1968) model also apply to the Hantush (1965) model conceptualisation. Additionally, the following parameters are considered for this model.

4.2.1 Hydraulic conductivity for the aquifer (*K*)

Hydraulic conductivity for the aquifer (*K*) is required to calculate streambed leakance. A hydraulic conductivity of 1 m/d will be used to define the aquifer in the model. This parameter is based on work done by Gomo (2011), where alluvial samples were analysed using the grain size analysis method.

4.2.2 Streambed thickness (*b'*)

Streambed thickness is required to calculate streambed leakance. The conceptual model is defined with a streambed thickness of 0.5 m (Table 4-1), based on general South African conditions (Moseki, 2013).

4.2.3 Hydraulic conductivity of streambed (K')

The hydraulic conductivity of the streambed is required to calculate streambed leakance. The streambed of the hypothetical conceptual model is defined using an average hydraulic conductivity of 0.01 m/d, describing the movement of water through a semi-pervious streambed material. The study area does not have specific measurements for the streambed, and thus literature values are used. Moseki (2013) estimated that semi-pervious streambed material in most South African conditions has a low average hydraulic conductivity. The vertical hydraulic conductivity of the streambed is an order of magnitude smaller than the horizontal hydraulic conductivity. The hydraulic conductivity of the streambed is a highly uncertain parameter because it is often not directly measured.

4.2.4 Streambed leakance (*L*)

Streambed leakance is calculated using Equ. 4-1.

$$L = \frac{K}{K'} b' \qquad \text{Equ. 4-1}$$

where,

K	is the hydraulic conductivity of aquifer;
K'	is the hydraulic conductivity of streambed material; and
b'	is the thickness of streambed.

From the previously defined base case values for the hypothetical model, the streambed leakance is calculated at 50 m. The streambed leakance is an uncertain parameter due to the difficulty of measuring the streambed thickness and hydraulic conductivity. Thus, it will be varied to determine the sensitivity of this parameter.

4.3 Conceptual model – Hunt (1999) model

The same parameters as discussed for the Jenkins (1968) model also apply to the Hunt (1999) model conceptualisation, as well as the hydraulic conductivity of the streambed and streambed thickness. Additionally, the following parameters are considered for this model.

4.3.1 Stream width (*W*)

The stream width for the hypothetical conceptual model is defined at an average value of 50 m (Table 4-1). This represents base case conditions, based on the study area, from which variations will be made to test the sensitivity of this parameter.

4.3.2 Streambed conductance (λ)

Streambed conductance (λ) is calculated using Equ. 4-2 :

$$\lambda = \frac{K'W}{M} \qquad \textbf{Equ. 4-2}$$

where,

W	is the stream width, and
M	is the streambed thickness.

From the previously defined base case values for the hypothetical model, the streambed conductance is calculated at 1 m/d. The streambed conductance is an uncertain parameter due to the difficulty of measuring the streambed thickness and hydraulic conductivity. Thus, it will be varied to determine the sensitivity of this parameter.

4.4 Conceptual model – Hunt (2003) model

The same parameters as discussed for the Jenkins (1968) model also apply to the Hunt (2003) model conceptualisation, as well as stream width. Additionally, the following parameters are considered for this model.

4.4.1 Specific yield of the aquitard (Sy'')

The specific yield for the hypothetical conceptual model is defined using a conservative average specific yield value of 0.1 (Table 4-1). This value is based on literature from the study area on which the typical model is based (Gomo, 2011). For unconfined aquifers, the specific yield typically ranges from 0.1 to 0.3 (Freeze and Cherry, 1979), thus why the lower range value is considered conservative.

4.4.2 Hydraulic conductivity of aquitard (K'')

Hydraulic conductivity of the underlying aquitard (clay) is determined in order of 10^{-6} m/d, based on the literature from the study area (Gomo, 2011). This value will be varied to assess the impact of the assigned hydraulic conductivity of the aquitard on the computed streamflow depletion.

4.4.3 Thickness of the aquitard (b'')

The conceptual model has an aquitard thickness of 30 m based on analysis of geological logs from the study area (Gomo, 2011). This parameter will vary from site to site and is also uncertain due to the difficulty of measurement. Thus, it will be varied to assess the influence it has on the simulated streamflow depletion.

4.4.4 Distance from streambed to bottom of aquitard (d'')

The distance from the streambed to the bottom of the aquitard is selected at an initial distance of 5 m (Table 4-1). The study area on which parameters are based does not have any specific information on this particular parameter. Thus, a test conducted by Hunt (2003) has been used to establish a baseline distance from the streambed to the bottom of the aquitard (Hunt, 2003). From this base case, the distance from the streambed to the bottom of the aquitard will be varied to assess how changing this parameter will affect the simulated streamflow depletion.

Table 4-1: Summary values of aquifer and stream parameters (Note: Internet-based interface for STEMDEPL08 uses imperial units to quantity streamflow depletion hence provided in this table)

Parameter	Symbol	Value (SI)	Value (Imperial units)	Source
Distance to pumping borehole from the stream	*d*	500 m	1640,4 ft	Google Earth
Transmissivity	*T*	70 m^2/d	753.4 ft^2/d	Pumping test results
Hydraulic conductivity for the aquifer	*K*	1 m/d	3.3 ft/d	Gomo (2011)
Storativity of the aquifer	*S*	0.02	0.02	Gomo (2011)
Abstraction rate	*Q*	2l/s	31.7 gal/min	Gomo (2011)
Pumping period	*t*	365 days		
Streambed thickness	*b′*	0.5 m	1.64 ft	Moseki (2013)
Stream width	*W*	50 m	164.04 ft	
Hydraulic conductivity of streambed	*K′*	0.01 m/d	0.03 ft/d	Moseki (2013)
Streambed leakance	*L*	50 m	164.04 ft	Calculated (Equ. 4-1)
Streambed conductance	λ	1 m/d	3.3 ft/d	Calculated (Equ. 4-2)
Hydraulic conductivity for the aquitard	*K′′*	0.09 m/d	0,28 ft/d	Gomo (2011)
Specific yield of Aquitard	*Sy*	0.1	0.1	Gomo (2011)
Aquitard thickness	*b′′*	30 m	98.4 ft	Gomo (2011)
Distance from streambed to bottom of aquitard (m)	*d′′*	5 m	16.4 ft	Hunt (2003)

Chapter 5 ANALYTICAL MODELS

Streamflow depletion is simulated using four analytical solutions based on the developed hypothetical conceptual model as described in Chapter 4 and Table 4-1. The four analytical solutions are Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003), as discussed in Chapter 4. Firstly, the base case is presented for each analytical model, and then a sensitivity analysis is performed on the various model input parameters to gauge the influence of varying parameters.

5.1 Base case

The base case hypothetical model is applied for each analytical solution to determine streamflow depletion. The simulated streamflow depletion over time for each solution is presented in Figure 5-1.

5.1.1 Jenkins (1968)

Jenkins's (1968) analytical approach is based on the need for a model with the fewest possible parameters and ease of implementation, as is demonstrated in the literature study. Maximum streamflow depletion after a 365-day pumping period is quantified to be 130 m^3/day (1.5 l/s) which is 75% of the maximum abstraction rate. The following presumptions may contribute to the overestimation of streamflow depletion with this solution:

- In the Jenkins (1968) solution, streambed resistance is disregarded, which has an impact on the accuracy of the streamflow depletion estimate.
- Stream geometry in the Jenkins (1968) equation is represented as a fully penetrated system (i.e., the depth of the stream is equivalent to the aquifer thickness).

5.1.2 Hantush (1965)

Hantush (1965), which was previously discussed in the literature review, is the extended Jenkins (1968) solution that takes into account the effects of a semi-impervious streambed layer, which in the developed conceptual model, is defined by clay with a streambed conductivity (*K'*) two orders of magnitude lower than the aquifer conductivity (Table 4-1).

According to the Hantush solution, the maximum amount of water withdrawn from the stream after 365 days of pumping is 126 m^3/day (Figure 5-1), which is less than the presented conservative Jenkins (1968) solution.

5.1.3 Hunt (1999)

Hunt (1999) further refined the Hantush (1965) analytical approach, to enhance aquifer-stream representation, as it has been previously described in the literature. According to Figure 5-1 with the assumption that all parameters are held constant during the same pumping duration, streamflow depletion volume decreases to 119 m^3/d, as compared to the Jenkins (1968) and Hantush (1965) solutions. The Hunt (1999) analytical solution inclusion of streambed impedance has a significant impact in delaying the response to streamflow depletion within the study's improved stream-aquifer conceptualisation.

5.1.4 Hunt (2003)

As discussed in the literature review, Hunt's (2003) analytical solution is the improved solution of stream-aquifer conceptualisation. As shown in Figure 5-1, there is a delay in streamflow depletion response for five (5) days. The delayed response of streamflow depletion is the result of pumped water initially coming from aquifer storage which does not affect streamflow. Due to the continuous pumping over time, the drawdown cone of depression expands and eventually after five (5) days draws water that would rather flow to the stream in the absence of pumping. This continuous pumping reduces streamflow and causes streamflow depletion. After 365-day pumping periods, the volume of water induced through the streambed into the aquifer is 56 m^3/d.

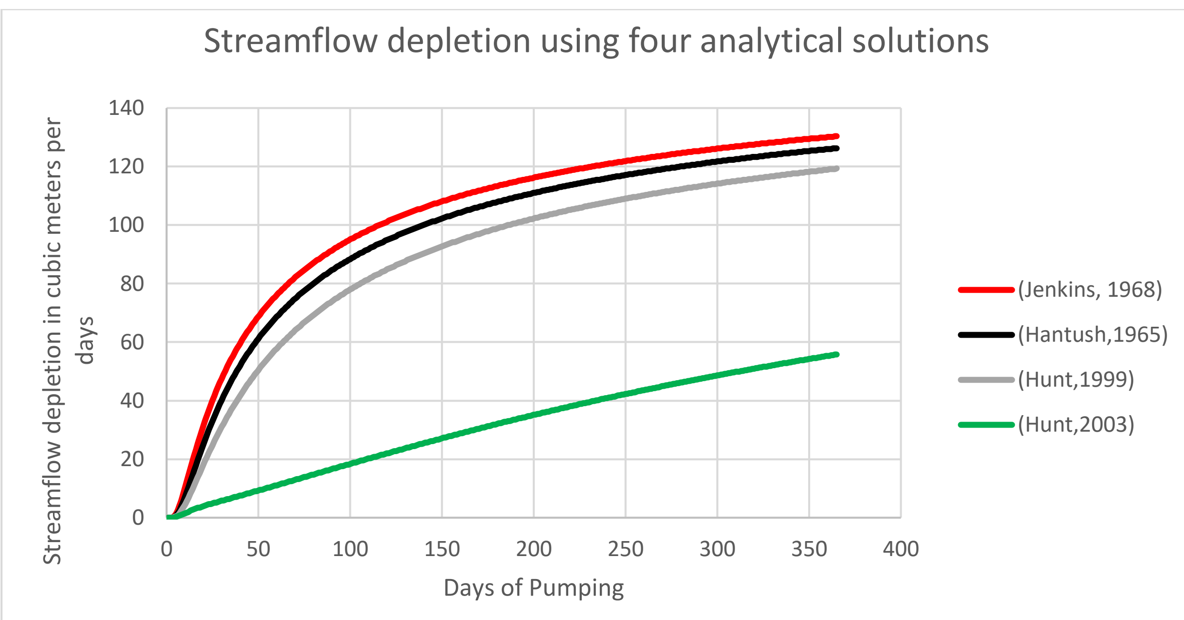


Figure 5-1: Modelled streamflow depletion from continuous pumping at developed hypothetical model determined by the STRMDEPL08 tool using four (4) analytical solutions.

5.1.5 Streamflow depletion as fraction/factor of pumping rate

Streamflow depletion eventually approaches the pumping rate with an assumption that the river is the only head-dependent aquifer boundary present in the developed conceptual model.

Streamflow depletion expressed as a fraction of the pumping rate at the borehole for both pumping rate scenarios is shown in Figure 5-2. The graph's curve asymptotically approaches 80% (value of 0.8) of the maximal pumping rate, and the reporting unit is dimensionless. As illustrated by the streamflow depletion curves by Jenkins's (1968) analytical model, the borehole captures more than 70% of its withdrawal from streamflow depletion within more than 200 days at the start of pumping. The time to reach a depletion-dominated supply is around 79 days (> 50%) of its withdrawal from streamflow depletion (storage-dominated supply is shown by a red dotted line (Figure 5-2). Over time as shown in Figure 5-2, the rate of depletion will tend to approach the pumping rate of the borehole. It is shown in Figure 5-2, no matter the abstraction rates at the borehole, the quantified streamflow depletion will always approach the maximum abstraction rates over time (value of 1).

At the beginning of pumping, the borehole captures more than 70% of its withdrawal from streamflow depletion within more than 290 days, as illustrated by the streamflow depletion curves using Hantush's (1965) analytical solution. It takes approximately 95 days (more than 50%) from the point of streamflow depletion to achieve a depletion-dominated supply.

As illustrated by the streamflow depletion curves by Hunt's (1999) analytical model, the borehole captures more than 70% of its withdrawal from streamflow depletion in 365 days of continuous pumping (i.e. due to extrapolating the streamflow depletion curve). The time to reach a depletion-dominated supply is around 127 days (> 50%) of its withdrawal from streamflow depletion which is longer than it takes for the previously quantified analytical solutions. However, the time it takes for streamflow depletion to approach value 1 depends on the parameter variation and improved analytical solutions.

The curve of the graph asymptotically approaches 40% of the maximum pumping rate (value of 0.4) for Hunt's (2003) model. A borehole captures less than 50% of its withdrawal from streamflow depletion within 365 days of continuous groundwater abstraction (storage-dominated supply). Therefore, time to reach a depletion-dominated supply does not exist within 365 days of continuous pumping (Figure 5-2).

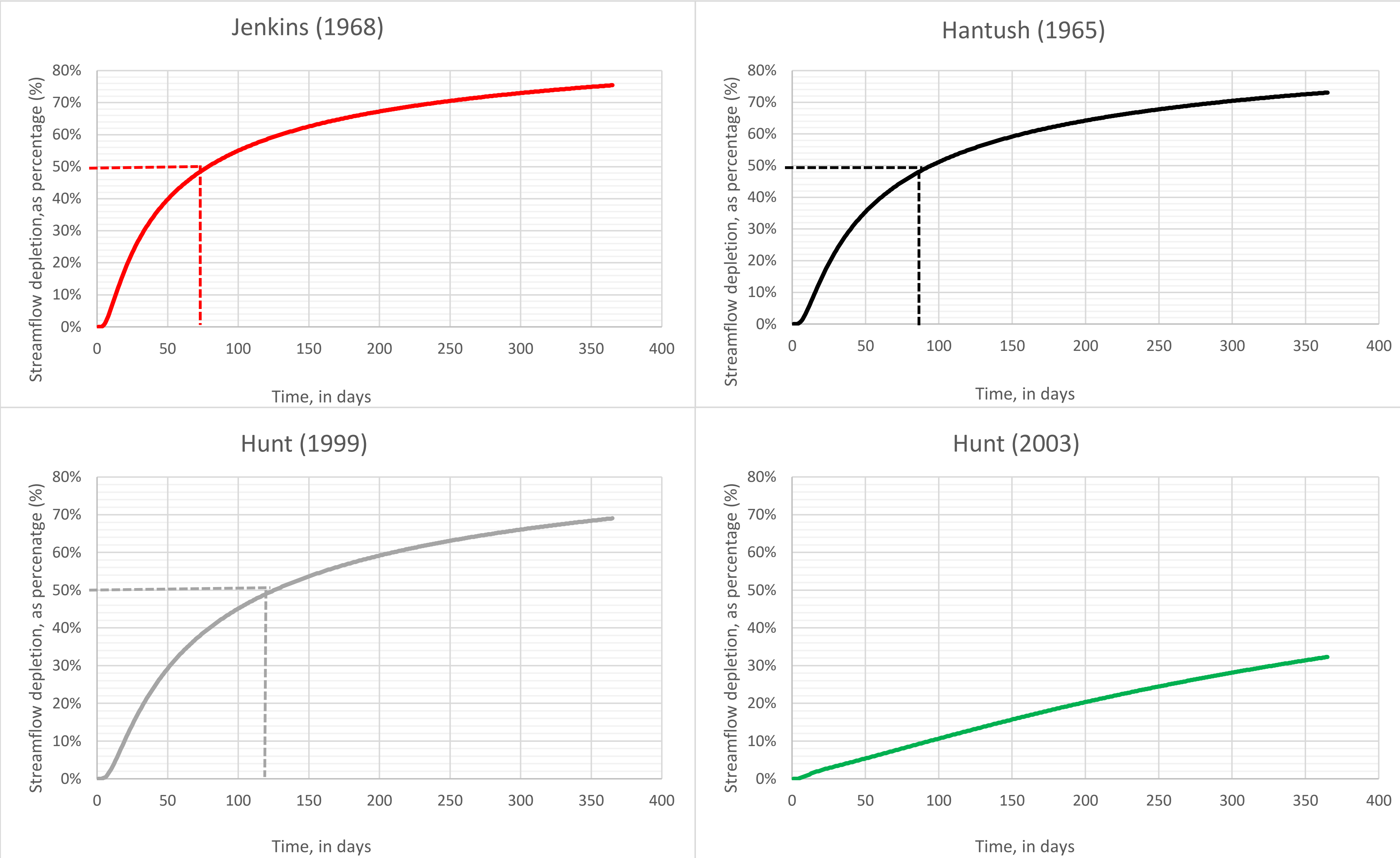


Figure 5-2: Streamflow depletion expressed as fraction of the groundwater pumping based on the developed conceptual model for the Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003) analytical models.

5.2 Sensitivity analysis

Sensitivity analysis is used to determine how changing parameters will impact the simulated streamflow depletion. The sensitivity analysis will help decision makers (water management bodies) to focus on parameters that greatly impact streamflow depletion and identify those parameters which require specific data collection. Parameters to be varied with analytical solutions to be used are outlined in Table 5-1.

Table 5-1: List of parameters to be varied within the analytical solutions.

Parameters to be varied	Analytical solutions used
Distance to pumping borehole from the stream	Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003)
Pumping rate	Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003)
Pumping period	Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003)
Transmissivity	Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003)
Storativity	Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003)
Streambed leakance	Hantush (1965)
Streambed conductance	Hunt (1999)

5.2.1 Distance to pumping borehole from the stream

In the conceptual model, the distance between the stream and the pumping borehole is 500 meters. It is noted from the literature review that as separation distance increases, stream depletion decreases, and vice versa. The selected distances to be tested, in reference to the study area, range from 10 m to 6 000 m about the base case scenario (500 m). The results from varying this parameter for the four analytical models are presented Figure 5-3**.**

Scenario: Distance 10 m

Figure 5-3 illustrates the solution of Jenkins's (1968) approach with the pumping borehole at a 10 m distance away from the stream within the developed conceptual model. It is noticed that there is no delay in streamflow depletion during the early time of groundwater abstraction. The borehole closer to the stream showed a rapid response in stream depletion, with the peak magnitude reaching the pumping rate. (i.e., 172 m^3/day volume, which is 99% of the actual pumping rate and 76 % streamflow depletion rate higher than the base case scenario at maximum continuous pumping time). It is also noticed that after 75 days of continuous pumping, the volume of groundwater discharge from the borehole equals the volume of water that is induced from the stream. It can therefore be deduced that after continuous pumping of the borehole closer to the stream allows early time equilibrium to be reached. The borehole too close to the stream takes a shorter time to draw water from the stream depending as well on the type of aquifer material and discharge rate.

Based on the assumption of the Hantush (1965) approach, the analytical solution is an extension of Jenkins (1968) to consider partial penetration. Initially during pumping, the cone of depression immediately extent steeply towards the river to affect the streamflow by inducing the volume of the stream by 94 m^3/day which is 55% of the maximum discharge abstraction rate (Figure 5-3). Over time as the cone of depression extends further, the maximum induced volume of the stream is 168 m^3/day which is 97% closer to the maximum groundwater discharge rate (173 m^3/day) and 75% higher than the streamflow depletion rate predicted by base case scenario (Figure 5-3).

Figure 5-3 illustrate the solution of Hunt's (1999) analytical solution, which is an extension from Hantush (1965) to further consider partial penetration of the stream. Initially during pumping, the cone of depression immediately extent steeply towards the river to affect the streamflow by inducing the volume of the stream by 50 m^3/day which is lower than Hantush's (1965) stream depletion model prediction due to the introduction of partially penetrating stream conditions and streambed impendence that alternatively causes an increase impedance in streamflow depletion. Over time as the cone of depression extends further, the maximum induced volume of the stream is 160 m^3/day which is 93% closer to the maximum groundwater discharge rate (173 m^3/day).

Figure 5-3 illustrates the solution of Hunt's (2003) analytical solution, which is an extension of Hunt's (1999) to consider an aquitard affecting the connection between the stream and groundwater. Initially at early times during pumping, the cone of depression immediately extends steeply towards the river to affect the streamflow by inducing the volume of the stream by 43.5 m^3/day which is lower than that quantified by Hunt (1999). The reason for the small rate of stream depletion at an early time is due to the water that initially comes from aquitard storage. Over time (immediate times) as the cone of depression extends further, the maximum induced volume of the stream increases as pumped water is derived from aquifer storage and stream. Streamflow depletion (138 m^3/day) eventually reaches the pumping rate (173 m^3/day).

Scenario: Distance 100 m

Figure 5-3 illustrates the solution of Jenkins's (1968) approach with the borehole 100 m away from the existing stream within the developed conceptual model. During the early time of pumping, the cone of depression expands to the river and induces a 40 m^3/day volume of water from the river. There is no delay response of streamflow depletion during the early time of pumping within the 100 m distance range. After some time of continuous pumping, the streamflow depletion reaches a pseudo-state state where the rate of change of streamflow depletion is constant.

Based on the assumption for the Hantush (1965) approach, the cone of depression immediately extent steeply towards the river to affect the streamflow by inducing the volume of the stream by 19 m^3/day which is 11% of the maximum discharge abstraction rate at an early period of groundwater abstraction. Over time as the cone of depression extends further, the maximum induced volume of the stream is 160 m^3/day which is 93% closer to the maximum groundwater discharge rate (173 m^3/day) to reach the equilibrium.

According to Hunt (1999), the cone of depression immediately extends steeply towards the river to affect the streamflow by inducing the volume of the stream by 9 m^3/day which is 0,1% of the maximum discharge abstraction rate during early pumping periods. Over time as the cone of depression extends further, the maximum induced volume of the stream is 152,40 m^3/day which is 88% closer to the maximum groundwater discharge rate (173 m^3/day) to reaching the equilibrium **(**Figure 5-3**)**.

Hunt (2003) further reduced streamflow depletion by 2,69 m^3/day which is 1,6 % of the maximum discharge abstraction rate during the early period of pumping as compared to Hunt (1999). Over time as the cone of depression extends further, the maximum induced volume of the stream is 76.8 m^3/day approaching the pumping rate (173 m^3/day) to reach the equilibrium, Figure 5-3**.**

Scenario: Distance 1 000 m

Figure 5-3 illustrates the solution of Jenkins's (1968) approach with the borehole at 1 000 m away from the stream within the developed conceptual model. During early pumping time, there exists a delay in response to streamflow depletion observed from the first 12 days of pumping (Figure 5-3). Delay in response to streamflow depletion implies that as groundwater is initially abstracted from the aquifer, the borehole initially extracts water from the aquifer storage which does not affect the stream. However, with continual pumping, the cone of depression expands outward from the borehole, and the pumping eventually reaches the zone of impact and affects the streamflow. After some time of continuous pumping, streamflow depletion volume continues to increase over time. However, the equilibrium (volumes of streamflow depletion equals to volume of groundwater abstraction over time) is not reached. The maximum induced volume of the stream rate after 365 days of continuous pumping is 92 m^3/day which is 53% of the maximum discharge rate of groundwater pumping. During early pumping time, there exists a delay in response to streamflow depletion observed from the first 13 days of pumping as observed by Hantush (1965). The maximum induced volume of stream rate after 365 days of continuous pumping is 89 m^3/day which is 53% of the maximum discharge rate of groundwater pumping.

Figure 5-3 illustrates the solution of Hunt's (1999) approach. During early pumping time, there exists a delay in response to streamflow depletion observed from the first 14 days of pumping. The maximum induced volume of stream rate after 365 days of continuous pumping is 83 m^3/day which is 48% of the maximum discharge rate of groundwater pumping.

During early pumping time, there exists a delay in response to streamflow depletion observed from the first 31 days of pumping. The maximum discharge rate of the stream is 9 m^3/day which is 5,3% of the pumping rate (172,8 m^3/day) to reach equilibrium.

Scenario: Distance 2 000 m

Figure 5-3 illustrates the solution of Jenkins's (1968) approach with the borehole at 2 000 m away from the existing stream within the developed hypothetical model. During early pumping time, there is a delay in response to streamflow depletion observed from the first 48 days of pumping (Figure 5-3). The maximum induced volume of stream rate after 365 days of continuous pumping is 47 m^3/day which is 21,5% of the maximum discharge rate of groundwater pumping.

Hantush (1965) quantifies that at early time groundwater pumping, there exists a delay in response to streamflow depletion observed from the first 51 days of pumping, which is 3 days delay higher than that quantified by Jenkins (1968) (Figure 5-3). The maximum induced volume of stream rate after 365 days of continuous pumping is 35 m^3/day which is 20% of the maximum discharge rate of groundwater pumping.

At the early time of pumping, Hunt (1999) shows an increase in delay time in response to streamflow depletion observed from the first 54 days of pumping. The maximum induced volume of stream rate after 365 days of continuous pumping is 32 m^3/day which is 19% of the maximum discharge rate of groundwater pumping **(**Figure 5-3**)**. The introduction of aquitard on Hunt's (2003) solution further increases the delay in response to streamflow depletion from the first 238 days of pumping, which is a 183-day delay than that quantified by Hunt (1999). The maximum induced volume of stream rate after 365 days of continuous pumping is 0,5 m^3/day which is 0,3% of the groundwater pumping rate **(**Figure 5-3**)**.

Scenario: Distance 5 000 m

Figure 5-3 illustrates the solution of Jenkins's (1968) approach. During early pumping time, there exists a delay in response to streamflow depletion observed from the first 305 days of pumping (Figure 5-3). The maximum induced volume of stream rate after 365 days of continuous pumping is 0,24 m^3/day which is 0,14% of the maximum discharge rate of pumping to reach equilibrium.

Figure 5-3 illustrates the solution of Hantush's (1965) approach. During early pumping time, there exists a delay in response to streamflow depletion observed from the first 310 days of pumping, which is 5 days higher in delay than that quantified by Jenkins (1968) (Figure 5-3). The maximum induced volume of stream rate after 365 days of continuous pumping is 0,24 m^3/day which is 0,14% of the maximum discharge rate of groundwater pumping.

Figure 5-3 illustrates the solution of Hunt's (1999) approach. During early pumping time, there exists a delay in response to streamflow depletion observed from the first 320 days of pumping, which is 10 days higher in delay than that quantified by Hantush (1965). The maximum induced volume of stream rate after 365 days of continuous pumping is 0,24 m^3/day which is 0,1% of the maximum discharge rate of pumping to reach equilibrium.

Figure 5-3 illustrates the solution of Hunt's approach with the borehole at 5 000 m away from the stream within the developed conceptual model. During continuous pumping over the distance of 5 000 m (5 km) away from the river, the streamflow is not affected over time (Figure 5-3). Consequently, the length of the cone of depression will take a lot longer to reach the stream as the distance from the stream grows.

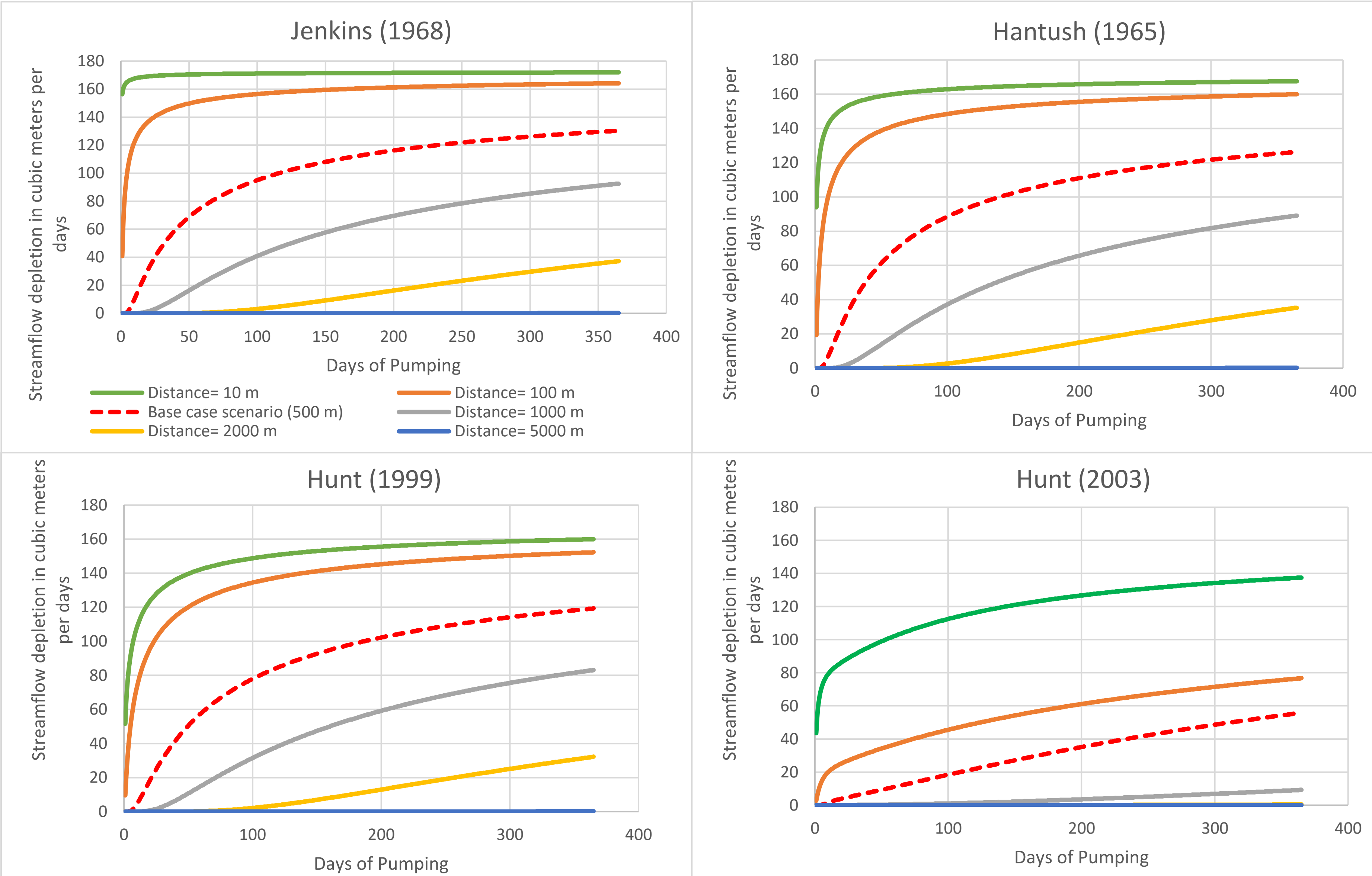


Figure 5-3: Modelled streamflow depletion from tested distance variation of continuous pumping borehole based on the developed conceptual model for the Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003) analytical models.

Summary

The following can be concluded when varying the distance parameter in the observation of streamflow depletion response:

- Streamflow depletion reduces with increasing distance (i.e., drawdown induced at the stream-aquifer boundary by the boreholes closest to the stream is greater than that created by a borehole farther away from the stream). The simulated graphs indicate that streamflow depletion rates are largest if the borehole is closer to the stream (i.e., 1 m away from the stream) and decrease as the distance of the pumped borehole from the stream increases.
- As distance increases the time delay response of streamflow depletion increases.
- The stream depletion from the pumped borehole within various distances approaches the pumping rate of the borehole over time (172,8 m^3/day).

It must be noted that streamflow depletion is parameter dependent, therefore all parameters outlined within the quantification of the streamflow depletion play a role in the behaviour of streamflow depletion, as such sensitivity of each parameter should be outlined in depth in water management and mitigation measures.

5.2.2 Pumping rate (m^3/day)

Four pumping rate scenarios were simulated in relation to the base case scenario (2 l/s), using the four analytical solutions for a 1-year period of continuous pumping rates. Scenario A simulates an order of magnitude lower constant pumping rate of 0.2 l/s in relation to the base case scenario, scenario B simulates with pumping rate of 1 l/s (half of the base case scenario), scenario C simulates with pumping rate of 4 l/s (double base case scenario pumping rate), and scenario D simulate with pumping rate of 20 l/s (an order of magnitude higher to the base case scenario). The objective is to evaluate the potential effects of pumping rates impact on streamflow depletion within the study area. The simulated streamflow depletion due to the effects of groundwater pumping is illustrated in Figure 5-3 for the four analytical solutions.

Scenario A: Abstraction rate 0.2 l/s

In this scenario, the pumping rate of the borehole is 17,28 m^3/day (0,2 l/s), an order of magnitude lower than the base case. For the Jenkins (1968) model, the streamflow depletion rate is delayed by four days as all water pumped by boreholes during this early time comes from water stored in the aquifer (i.e., storage-dominated supply). After four days of continuous abstraction, the cone of depression progressively widens laterally with time. It is shown in Figure 5-4 that as the borehole abstraction continues for an extended maximum period of time, the rate of water from stream depletion (source of water pumped by the well is from depletion) is 13 m^3/day, which is 75% of the maximum groundwater abstraction rate. The streamflow depletion is also reduced by an order of magnitude when the abstraction rate is reduced by an order of magnitude (Section 5.1.1) with the same percentage of the abstraction rate. This highlights that there is a linear relationship between the abstraction rate and streamflow depletion. For the Hantush (1965) model, the simulated streamflow depletion is delayed by five days. The cone of depression gradually widens laterally during five days of continuous pumping. It is shown in Figure 5-4 that as the borehole abstraction continues for an extended maximum period of time, the rate of water from streamflow depletion is 12,7 m^3/day, which is 73% of the maximum groundwater abstraction rate. Despite the introduction of the streambed resistance factor to Hantush's (1965) solution, maximum streamflow depletion is similar to the one predicted by Jenkins (1968).

For the Hunt (1999) model, the simulated streamflow depletion is delayed by six days. It is shown in Figure 5-4 that as the borehole abstraction continues for an extended maximum period of time, the rate of water from stream depletion (source of water pumped by the well from depletion) is 12 m^3/day, which is 69% of the maximum groundwater abstraction rate. Despite the introduction of streambed resistance factor and partially penetrating stream to Hunt's (1999) solution, maximum streamflow depletion is similar to the one predicted by Hantush (1999), there is no sudden change to streamflow depletion with varying the analytical solution but maintaining the same pumping rate.

The streamflow depletion rate is delayed by nine days for the Hunt (2003) model. It is shown in Figure 5-4 that as the borehole abstraction continues for an extended maximum period of time, the rate of water from stream depletion is 6 m^3/day, which is 35% of the maximum groundwater abstraction rate. This is lower than streamflow depletion predicted by Hunt (1999), in this instance, the decrease in aquifer storage is the main source of water (or supply) to the borehole. A similar trend to the base case is seen for all four analytical solutions, with an order of magnitude reduction in streamflow when the abstraction rate is reduced by an order of magnitude.

Scenario B: Abstraction rate 1 l/s

In this scenario, the pumping rate of the borehole is 86.4 m^3/day (1 l/s). During the early time, the Jenkins (1968) model shows a delay in stream flow depletion response by three days as all water pumped by boreholes during this early time comes from water stored in the aquifer (i.e., storage-dominated supply). It is shown in Figure 5-4 that as the borehole abstraction continues for an extended maximum period of time, the rate of water from stream depletion (source of water pumped by the well is from depletion) is 65 m^3/day, which is 75% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year).

Hantush's (1965) model predicts the longer lag time of response to streamflow deletion during early time of groundwater pumping by four days as compared to Jenkins (1968). It is shown in Figure 5-4 that as the borehole abstraction continues for an extended maximum period of time, the rate of water from stream depletion (source of water pumped by the well from depletion) is 63 m^3/day, which is 73% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year).

The Hunt (1999) model predicts a similar lag time response to streamflow depletion during early time of groundwater pumping by five days as compared to the prediction made by Hantush (1965). It is shown in Figure 5-4 that as the borehole abstraction continues for an extended maximum period of time, the rate of water from stream depletion (source of water pumped by the well from depletion) is 60 m^3/day, which is 69% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year).

The Hunt (2003) model also predicts a similar lag time response to streamflow depletion during early time of groundwater pumping by five as compared to the prediction made by Hantush (1965). It is shown in Figure 5-4 that as the borehole abstraction continues for an increased maximum time, the rate of water from stream depletion is 28 m^3/day, which is 32% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year), in this case source of water is influenced by reduction in aquifer storage.

Scenario C: Abstraction 4 l/s

In this scenario, the pumping rate of the borehole is 345.6 m^3/day (4 l/s). During the early time, the Jenkins (1968) model shows a delay in stream flow depletion response by two days as all water pumped by boreholes during this early time comes from water stored in the aquifer (i.e., storage-dominated supply). After two days of continuous pumping, the streamflow starts to be influenced by simulated streamflow depletion by the model. It is shown in Figure 5-4 that as the borehole abstraction continues for an extended maximum period of time, the rate of water from stream depletion (source of water pumped by the well from depletion) is 261 m^3 /day, which is 76% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year) which shows depletion dominated supply.

The Hantush (1965) model predicts the longer lag time of response to streamflow deletion during early time of groundwater pumping by three days as compared to Jenkins (1968). It is shown in Figure 5-4 that as the borehole abstraction continues for an extended maximum period, the rate of water from stream depletion (source of water pumped by the well from depletion) is 252 m^3/day, which is 73% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year) which shows depletion dominated supply.

The Hunt (1999) model predicts a similar lag time of response to streamflow deletion during the early time of groundwater pumping by three days, as compared to Hantush (1965). As borehole abstraction continues for a maximum period, the rate of water coming from streamflow is 238 m^3/day, which is 69% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year). This shows a depletion-dominated supply.

The Hunt (2003) model still predicts a similar lag time of response to streamflow deletion during the early time of groundwater pumping by three days, as compared to Hunt (1999). The total rate of water from stream depletion is 112 m^3/day, which is 32% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year) which), in this case, the source of water to the borehole is influenced by a reduction in aquifer storage.

Scenario D: Abstraction 20 l/s

In this scenario, the pumping rate of the borehole is 1 728 m^3/day (20 l/s). During the early time, the Jenkins (1968) model shows a delay in stream flow depletion response by two days as all water pumped by boreholes during this early time comes from water stored in the aquifer (i.e., storage-dominated supply). After two days of continuous pumping, the zone of influence increases and begins to interact with the river and streamflow. As the borehole abstraction continues, the maximum streamflow depletion rate is 1 303 m^3/day, which is 75% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year) which shows a depletion-dominated supply.

The Hantush (1965) model predicts a similar lag time of response to streamflow deletion during the early time of groundwater pumping by two days as compared to Jenkins (1968). However, the maximum streamflow simulated by the Hantush (1965) model 1 263 m^3/day, which is 73% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year). This is slightly lower than the Jenkins (1968) model and can be accounted for by the inclusion of a partially penetrating system which reduces the simulated streamflow depletion.

The Hunt (1999) model predicts a similar lag time of response to streamflow deletion during the early time of groundwater pumping by two days as compared to Hantush (1965). Yet, the simulated total streamflow depletion rate is 1 193 m^3/day, which is 69% of the maximum groundwater abstraction rate. Again, this rate is slightly lower than the previous models because this solution considers more of a partially penetrating stream.

The Hunt (2003) model still predicts a similar lag time of response to streamflow deletion during the early time of groundwater pumping by two days as compared to Hunt (1999). However, the simulated streamflow depletion rate is much lower at 558 m^3/day (32% of the maximum groundwater abstraction rate). This is due to the inclusion of the aquitard which further reduces the interaction between surface water and groundwater.

Summary

The following can be concluded when varying the abstraction rate in observation of streamflow depletion response:

- A linear relationship between the abstraction rate and streamflow depletion is shown in the simulation results
- The improved conceptualisation of aquifer–stream dynamics demonstrates that it requires more time to obtain steady-state streamflow depletion (i.e., Hunt, 2003 does not reach a steady-state rate of streamflow depletion than Jenkins, 1968 assuming all other factors are equal).
- The Hunt (2003) model has an improved conceptualisation of aquifer-stream dynamics because it accounts for the streambed influence; and therefore, a delay in the response time is seen.

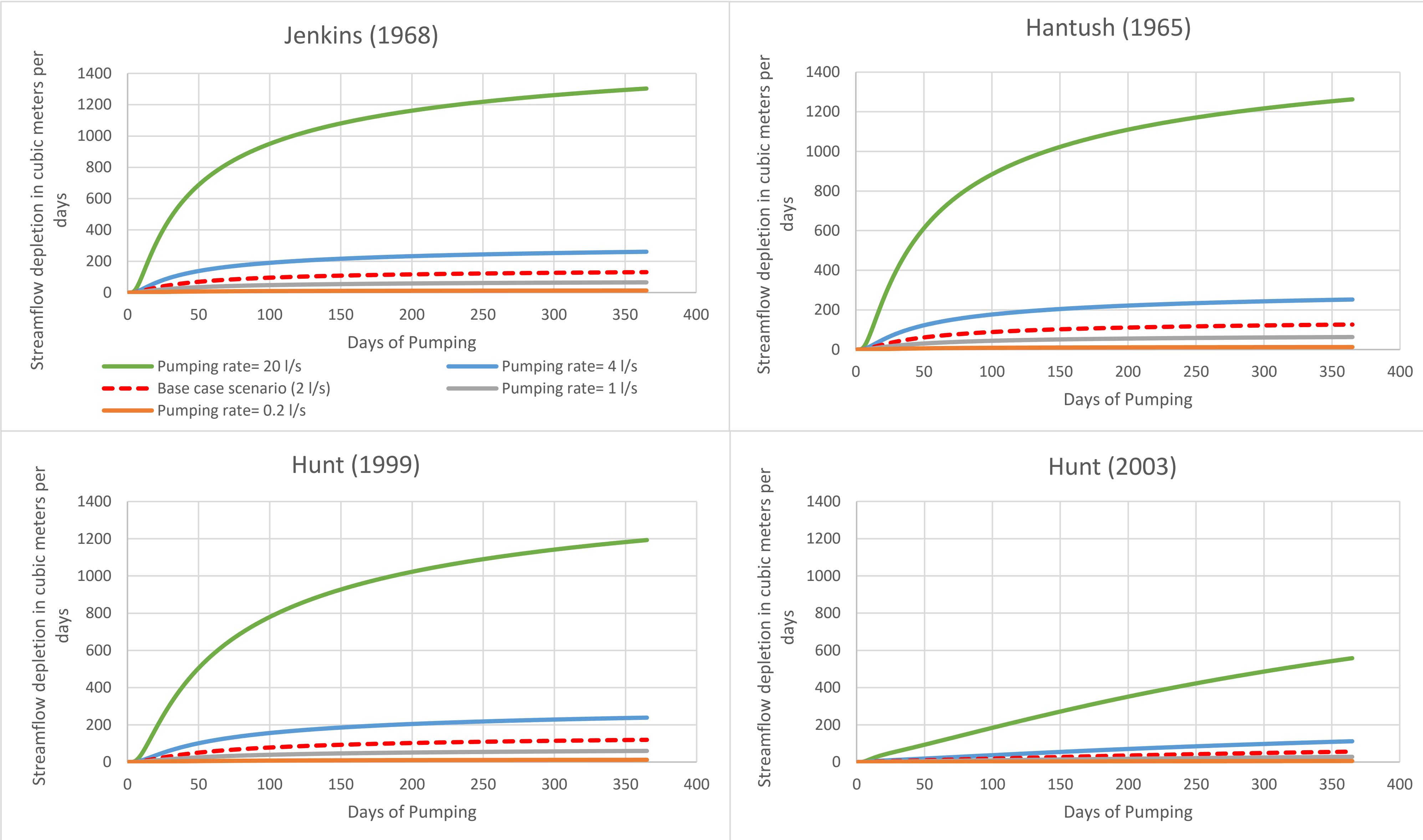


Figure 5-4: Stream depletion as a function of time for four scenarios of groundwater pumping based on the developed conceptual model for the Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003) analytical models.

5.2.3 Transmissivity (m^2/day)

Transmissivity is one of the significant parameters that influence the timing and rates of streamflow depletion. From this base case transmissivity value, a few scenarios are simulated to determine the influence of varying this parameter on the predicted streamflow depletion. The transmissivity scenarios will include the base case (70 m^2/d) with additional variations at 7, 35, 150, and 700 m^2/day. The simulated streamflow depletion for each transmissivity scenario and each of the four analytical solutions is presented in Figure 5-5.

Scenario A: Transmissivity 7 m^2/day

As illustrated in Figure 5-5, the transmissivity variation quantified by Jenkins (1968) is directly proportional to the rates of streamflow depletion. Relatively low transmissivity within the homogenous material results in 31 days of lag time response between pumping and initial streamflow depletion. It is shown in Figure 5-5 that after 365 days of continuous pumping, the Jenkins (1968) model results show 32% (56 m^3/day) of normalised streamflow depletion, which shows storage-dominated supply within the simulation period. This is greatly reduced from the base case scenario of 130 m^3/day (75% of abstraction rate).

The Hantush (1965) model further reduces the lag time response between pumping and streamflow depletion by 35 days (4 days higher than the one predicted by Jenkins, 1968). During the early time, for 35 days water initially comes from the aquifer. After 35 days of groundwater pumping, the cone of depression expands to induce streamflow causing streamflow depletion. It is shown that the Hantush (1965) model simulates a lower streamflow depletion rate of 48 m^3/day (28% of abstraction rate). The slight reduction of streamflow depletion rate predicted by Hantush (1965) is a result of the introduction of the semi-permeable layer at the boundary between the aquifer and stream bank.

The introduction of a partially penetrating, semi-pervious layer by the Hunt (1999) model reduces the lag time response of streamflow depletion to 32 days of groundwater pumping as groundwater initially comes from the aquifer. After 32 days, the rate of streamflow depletion increases. Maximum stream flow depletion after a 365-day simulation period is 54 m^3/day (31% of the groundwater abstraction rate).

Hunt (2003) increases the lag time response between pumping and streamflow depletion by a magnitude higher (101 days) due to the introduction of an aquiclude layer. During early, the rate of streamflow depletion is small, as pumped water comes mainly from aquitard storage. After 101 days of continuous pumping, the rate of streamflow depletion increases and after 365 of continuous pumping the maximum rate of streamflow depletion lowers to 3 m^3/day (Figure 5-5) which is 2% of groundwater abstraction.

Scenario B: Transmissivity 35 m^2/day

Increasing the transmissivity value to 35 m^2/day from the previous low scenario illustrates that as the transmissivity value increases, the lag time response between pumping and streamflow reduces. For the Jenkins (1968) model, the lag time is 6 days, as compared to the 31-day lag in the previous scenario. After 6 days of pumping, the streamflow depletion rate increases to a maximum of 114 m^3/day (Figure 5-5) after a 365-day simulation period, which is 66% of groundwater abstraction. Therefore, time to reach a depletion-dominated supply exists within 365 days of continuous pumping.

Hantush's (1965) analytical solution results in a lag time response between pumping and streamflow depletion of seven days (one day more than predicted by Jenkins (1968)). During the early time of seven days, water initially comes from the aquifer. After seven days of groundwater pumping, the cone of depression expands to induce streamflow causing streamflow depletion. It is shown that Hantush (1965) reduces maximum streamflow depletion to 108 m^3/day (63% of subtraction rate) of the groundwater abstraction after the 365-day simulation period, which shows a depletion-dominated supply within the total simulation period.

Hunt (1999) predicts that early, the lag time response between pumping and streamflow is seven days. After seven (7) days of groundwater pumping, the streamflow depletion rate increases to a maximum streamflow depletion rate of 106 m^3/day (61% of the groundwater abstraction), which shows a depletion-dominated supply.

The lag time response predicted by Hunt (2003) increases to 9 days, with a maximum streamflow depletion of 38 m^3/day which is 22% of the groundwater abstraction. Therefore, time to reach a depletion-dominated supply does not exist within 365 days of continuous pumping (storage - depletion supply was reached) because of the influence of the aquitard and partially penetrating, semi-pervious stream.

Scenario C: Transmissivity 150 m^2/day

An increase in transmissivity to 150 m^2/d from the base case scenario, results in a one-day lag response in the simulated streamflow depletion by the Jenkins (1968) model (Figure 5-5). The Jenkins (1968) model predicts a short lag time response of time for a cone of depression to expand and induce the stream to cause streamflow depletion. As pumping continues for the total simulation period, the maximum streamflow depletion rate is 144 m^3/day, which is 83% of the abstraction rate. Eventually, streamflow depletion approaches the pumping rate at a large time (365 days). Consequently, depletion-dominated supply is reached within the total simulation period.

The Hantush (1965) model reduced the maximum streamflow depletion rates to 141 m^3/day, which is 81% of the abstraction rate.

The Hunt (1999) model increases the lag time response between pumping and depletion to two days. After continuous pumping within the total simulation period, the streamflow depletion rate is reduced to 127 m^3/day, which is 73% of groundwater abstraction (Figure 5-5).

Due to the presence of an aquitard in the Hunt (2003) model, the maximum streamflow depletion is reduced to 69 m^3/day, which is 40% of the abstraction rate. Consequently, depletion-dominated supply was not reached within the total simulation period (storage-depletion supply was reached).

Scenario C: Transmissivity 700 m^2/day

For this high transmissivity scenario, representing unconsolidated sand and gravel of the alluvial aquifer, there is no lag time for the Jenkins (1968) and Hantush (1965) models, with a steep increase in the streamflow depletion rate from the beginning of the simulation. Thus, representing a highly connected system where the surface water–groundwater interaction is instantaneous. The Jenkins (1968) model simulates a maximum streamflow depletion rate of 159 m^3/day (Figure 5-5), which is 92% of the maximum groundwater abstraction rate over the total simulation period of 365 days (1 year) which shows a depletion-dominated supply.

Hantush (1965) reduced the maximum streamflow depletion rate to 158 m^3/day (Figure 5-5), which is 91% of the abstraction rate. After continuous pumping within the total simulation period, the streamflow depletion rate predicted by Hunt (1999) is reduced to 123 m^3/day (Figure 5-5), which is 71% of groundwater abstraction. Due to the presence of aquiclude presented by Hunt (2003), the maximum streamflow depletion rate is reduced to 70 m^3/day, which is 41% of the abstraction rate. Therefore, the time to reach depletion-dominated supply was not reached within the total simulation period (storage-depletion supply was reached).

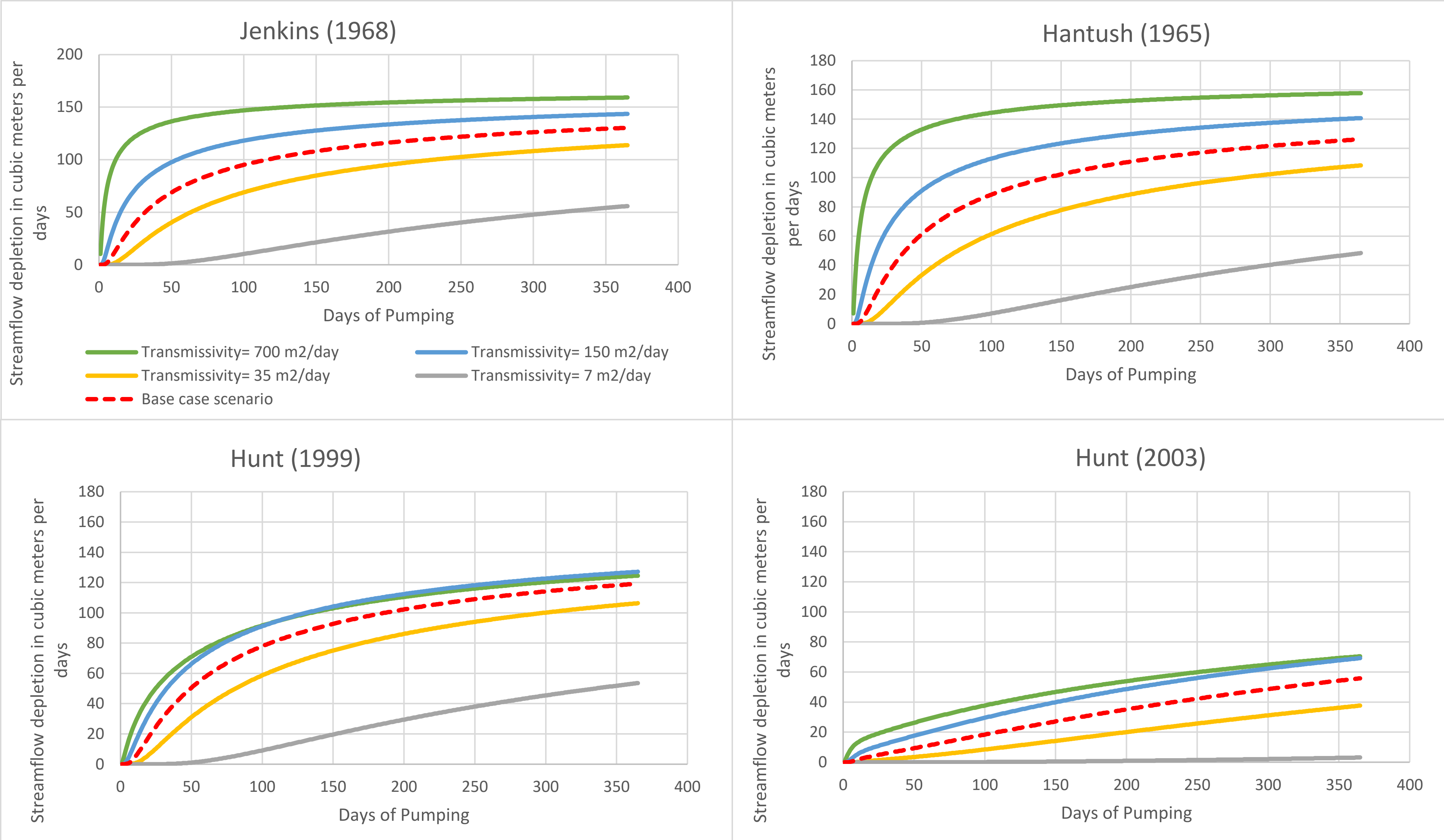


Figure 5-5: Modelled streamflow depletion from continuous pumping with transmissivity variation based on the developed conceptual model for the Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003) analytical models.

Summary

The following can be concluded when varying the transmissivities parameter in the observation of streamflow depletion response:

- For higher transmissivities, the time lag between pumping and depletions is shorter, with $\frac{Q_S}{Q_W}$= 1 being approached more quickly than small transmissivities.
- Streambed leakage in Hantush (1965) and aquiclude introduced in Hunt (2003) has a great impact in reducing streamflow depletion as transmissivity increases.

5.2.4 Storativity (S)

Storativity is also a significant parameter that potentially controls the timing and rates of streamflow depletion depending on aquifer storage. From the base case storativity value, scenarios are simulated to determine the influence of varying this parameter on the predicted streamflow depletion. The storativity scenarios include the base case (0.02) with additional variations at 0.005 and 0.3, to represent confined and unconfined conditions. The simulated streamflow depletion for each storativity scenario and each of the four analytical solutions is presented in Figure 5-6.

Scenario A: storativity 0.005 (confined)

For the Jenkins (1968) model, the reduced storativity values, result in a rapid streamflow depletion with no time delays, due to limited aquifer storage and direct connection with the river in this analytical model. The maximum streamflow depletion rate is 151 m^3/day over the period of simulation, which is 87% of the abstraction rate (Figure 5-6). Therefore, the time to reach depletion-dominated supply was reached within the total simulation period.

The Hantush (1968) model introduced the effect of a semi-impervious streambed layer, which lowers the effect of streamflow depletion as compared to Jenkins (1968), to a maximum streamflow depletion rate of 149 m^3/day over the total simulation period, which is 86% of the total abstraction rate. It can be concluded that the streamflow depletion-dominated supply was reached as 86% of water comes from stream depletion and 14% still comes from storage supply (aquifer).

The Hunt (1999) model further reduced streamflow depletion, due to the inclusion of a partially penetrating stream, streamflow depletion rates substantially increase until the maximum depletion rate of 145 m^3/day is reached, which is 84% of the total abstraction rate. Depletion-dominated supply was reached with 84% of the water coming from streamflow during continuous pumping and 16% of water still coming from aquifer storage supply.

The introduction of the aquitard layer by the Hunt (2003) model slows the process of streamflow depletion, with a simulated lag time between groundwater pumping and streamflow depletion of 1 day. After 1 day of pumping, streamflow depletion increases linearly to a maximum rate of 59 m^3/day, which is 34% of the total abstraction rate (Figure 5-6). Therefore, time to reach a depletion-dominated supply does not exist within 365 days of continuous pumping (storage - depletion supply was reached).

Scenario B: storativity 0.3 (unconfined)

For the increased storativity scenario, representing an unconfined system, Jenkins (1968) simulates a slower the propagation of cone of depletion towards the river. During the early time of groundwater pumping, the lag response time increases to 46 days of storage-dominated supply (Figure 5-6). After 46 days of continuous pumping, the streamflow depletion rate increased linearly to a maximum rate of 39 m^3/day, which is 23% of the total abstraction rate. Therefore, the time to reach a depletion-dominated supply was not achieved within 365 days of continuous pumping (storage-depletion supply was reached).

The Hantush (1968) model introduced the effect of a semi-impervious streambed layer, which lowers the effect of streamflow depletion as compared to Jenkins (1968). During the early time of groundwater pumping, the lag response time increased to 53 days of storage-dominated supply (Figure 5-6). After 53 days of continuous pumping, the streamflow depletion rate increased linearly to a maximum rate of 32 m^3/day, which is 19% of the total abstraction rate. Therefore, time to reach a depletion-dominated supply does not exist within 365 days of continuous pumping (storage - depletion supply was reached).

The Hunt (1999) model further increases the lag response time between streamflow depletion and abstraction rates to about 59 days of storage-dominated supply (Figure 5-6). After 59 days of continuous pumping, the streamflow depletion rate increased linearly to a maximum rate of 24 m^3/day, which is 14% of the total abstraction rate. Therefore, time to reach a depletion-dominated supply does not exist within 365 days of continuous pumping (storage - depletion supply was reached).

The introduction of the aquitard layer by the Hunt (2003) model substantially slows the process of depletion with a lag time between groundwater pumping and streamflow depletion of 66 days. After 66 days of pumping, streamflow depletion increases linearly to a maximum rate of 15 m^3/day, which is 9% of the total abstraction rate (Figure 5-6). Therefore, a depletion-dominated supply is not seen within the 365-day simulation period.

Summary

The following can be concluded when varying the storativity parameter in the observed streamflow depletion response:

- Groundwater abstraction that occurs in a confined aquifer produces substantially more rapid depletion than in an unconfined aquifer, where the systems are in direct connection.
- Unconfined aquifers have a larger storage property than confined aquifers, which causes the confined aquifers to speed up depletion from groundwater pumping than unconfined aquifers.
- Vertical transmissivity is assumed to be negligible as all four analytical solutions encompass that the pumping occurs over the entire thickness of the aquifer (i.e., horizontal transmissivity is higher than vertical conductivity)

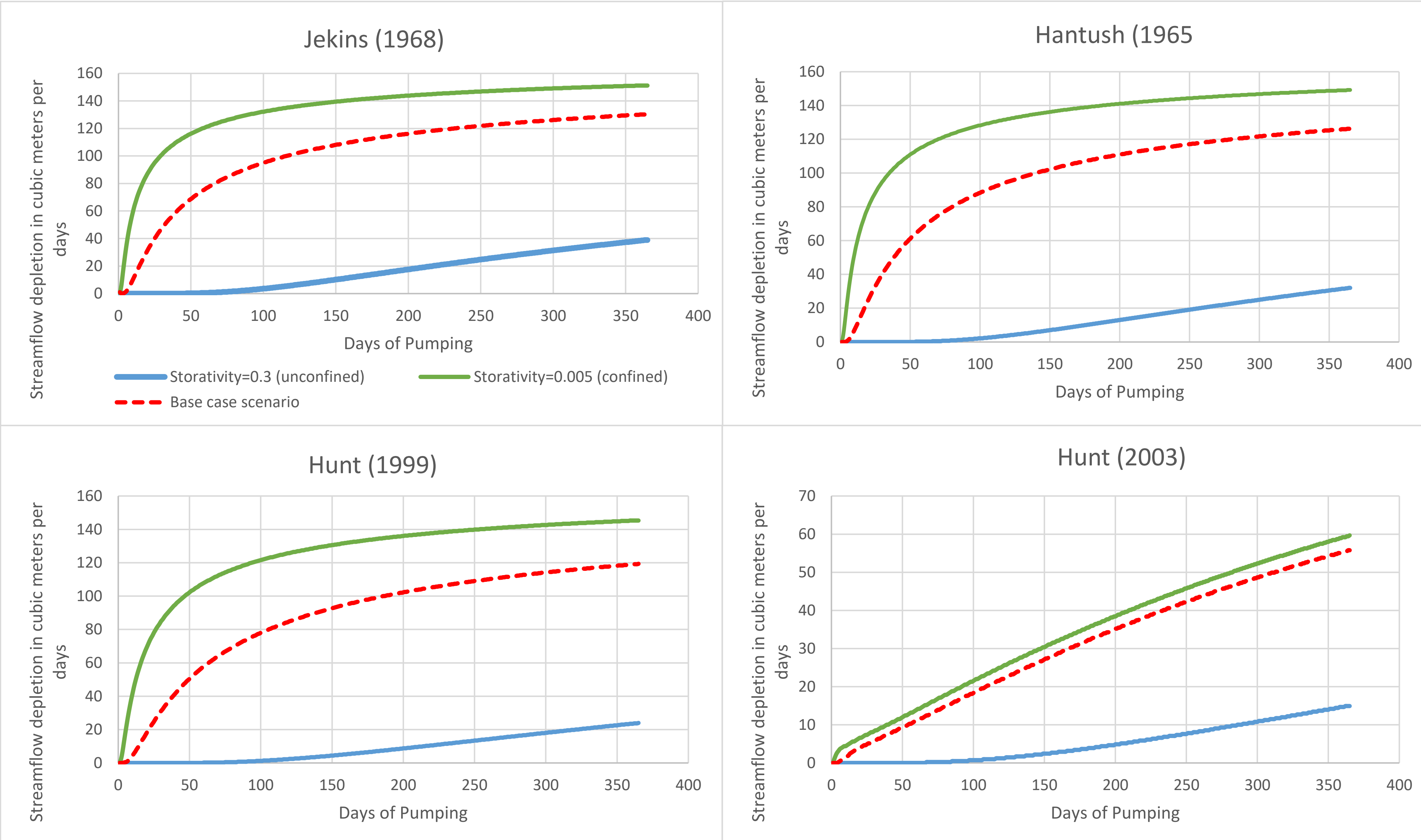


Figure 5-6: Modelled streamflow depletion from continuous pumping with storativity variation based on the developed conceptual model for the Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003) analytical models.

5.2.5 Streambed leakance

Different sediments with a range of conductivities are presented to properly compare the influence of streambed sediments. Based on grain size analysis, as presented by Gomo 2011, three possible sediments, namely silty sand, silt, and clay were used with their estimated hydraulic conductivities with aquifer sediments ranging up to 3-orders of magnitude as presented by Hazen (1911) and as shown in Table 5-2. It is significant to highlight that the conductivity of streambed sediments is orders of magnitude lower than that of aquifer sediments. Three scenarios of streambed conductivity have been used to quantify the sensitive analysis of the rate of streamflow depletion with conductivity ratios (*K/K'*) between the pumped aquifer and streambed material of 10, 100 and 1000.

The Jenkins (1968) model does not have a streambed leakance parameter. Thus, only the Hantush (1965) model was used to vary the streambed leakance parameter. Figure 5-7 presents the simulated streamflow depletion by the Hantush (1965) model for each scenario.

Table 5-2: Summary of different conductivity (*K/K'*) ratio

Aquifer Properties		Streambed Properties		Conductivity ratio (*K/K'*)	Streambed Leakance (*L*) (m)
Conductivity (*K*) (m/d)	Representative texture	Conductivity (*K'*) (m/d)	Representative texture		
1	Gravel-sand	0.1	Silty sand	10	5
		0.01	Silt	100	50
		0.001	Clay	1000	500

Scenario A: Streambed conductivity (0.1; Silty sand)

For this scenario of the highest conductivity for the streambed material, the simulated streamflow depletion for the Hantush (1965) model corresponds to the results for the Jenkins (1968) model. This correlation indicates that the streambed materials in this scenario are not impeding interaction between the surface water and groundwater, mimicking the fully penetrating assumptions of the Jenkins (1968) model.

Scenario B: Streambed conductivity (0.001; Silt)

As the streambed leakance increases due to the decrease in streambed conductivity, maximum streamflow depletion decreases to 126 m^3/day (97% of abstraction rate) (Figure 5-7).

Scenario C: Streambed conductivity (0.0001; Clay)

The influence of low streambed conductivity (clay) is evident with the Hantush (1965) model simulating a maximum streamflow depletion rate of 95 m^3/day (73% of abstraction rate) (Figure 5-7) over the simulation period. There is also a substantial delay in the streamflow depletion and a slower increase in streamflow depletion over time.

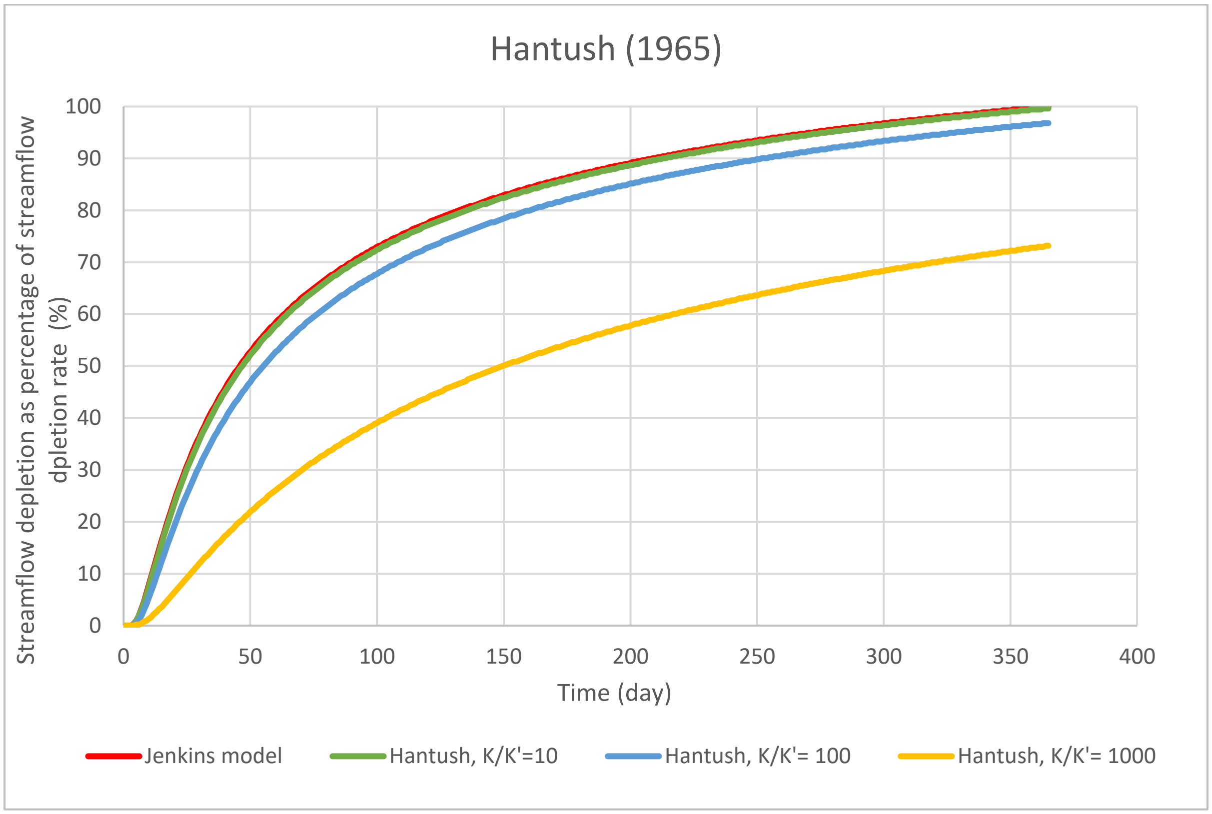


Figure 5-7: Hantush (1965) model solution for various conductivity ratios (*K/K'*) with borehole distance from the stream = 500, aquifer hydraulic conductivity = 1 m/d, aquifer thickness = 30 m, specific yield = 0.1 and streambed thickness = 0.5 m.

Summary

- The comparison indicates that when the difference in aquifer and streambed conductivity increases, so does the impedance on the rate of streamflow depletion.

5.2.6 Streambed conductance

Four streambed conductance scenarios were simulated in relation to the base case scenario (1 m/day) using Hunt (1999) analytical solutions for 1-year period of continuous pumping rates. Hunt (1999) was utilised to simulate streamflow conductance because Jenkins (1968) and Hantush (1965) do not include a streambed conductance component. Scenario A simulates an order of magnitude lower constant streambed conductance of 0.1 m/day in relation to the base case scenario; scenario B simulates a streambed conductance of 0.5 m/day, which is half of the base case scenario; scenario C simulates a streambed conductance of 2 m/day, which is double base case scenario pumping rate; and scenario D simulate a streambed conductance of 10 m/day, an order of magnitude higher than the base case scenario. The objective is to evaluate the impact of groundwater movement through a streambed on streamflow depletion for the developed model.

The Hunt (1999) model can assess the effect of streambed conductance, or how rapidly groundwater flows through the streambed, on streamflow depletion from groundwater abstraction. The simulated streamflow depletion for the streambed conductance scenario for each of the four analytical models is presented in Figure 5-8.

Scenario A: Streambed conductance 0.1 m/day

For this streambed conductance scenario (0.1 m/day), the Hunt (1999) model results show relatively low streambed conductivity (low groundwater movement through the streambed), which results in six days of lag time response between pumping and streamflow depletions at an early time. Over time, as pumping continues, the maximum streamflow depletion rates increase to 61 m^3/day over the period of simulation, which is 35% of the abstraction rate (Figure 5-8). Therefore, time to reach a depletion-dominated supply does not exist within 365 days of continuous pumping (storage - depletion supply was reached).

Scenario B: Streambed conductance 0.5 m/day

For this streambed conductance scenario (0.5 m/day), the Hunt (1999) model results show moderately low streambed conductivity (low groundwater movement through the streambed), which results in four days of lag time response between pumping and streamflow depletions at an early time. Over time, as pumping continuous, maximum streamflow depletion rates increase to 109 m^3/day over the period of simulation, which is 63% of the abstraction rate (Figure 5-8). Therefore, time to reach a depletion-dominated supply exists within 365 days of continuous pumping within groundwater moving through a streambed of 0.5 m/day conductivity.

Scenario C: Streambed conductance 2 m/day

For this streambed conductance scenario (2 m/day), the Hunt (1999) model results show high streambed conductivity (high groundwater movement through the streambed) which results in three days of lag time response between pumping and streamflow depletions at an early time. Over time, as pumping continuous, maximum streamflow depletion rates increase to 125 m^3/day over the period of simulation, which is 72% of the abstraction rate (Figure 5-8). Therefore, time to reach a depletion-dominated supply exists within 365 days of continuous pumping within groundwater moving through a streambed of 2 m/day conductivity.

Scenario C: Streambed conductance 2 m/day

Figure 5-8 shows relatively high streambed conductivity (high groundwater movement through the streambed) for this streambed conductance scenario (2 m/day), with three days of lag time response between pumping and streamflow depletions at an early time. Over time, as pumping continuous, maximum streamflow depletion rates increase to 129 m^3/day over the period of simulation which is 75% of the abstraction rate (Figure 5-8). Therefore, time to reach a depletion-dominated supply exists within 365 days of continuous pumping within groundwater moving through a streambed of 10 m/day conductivity.

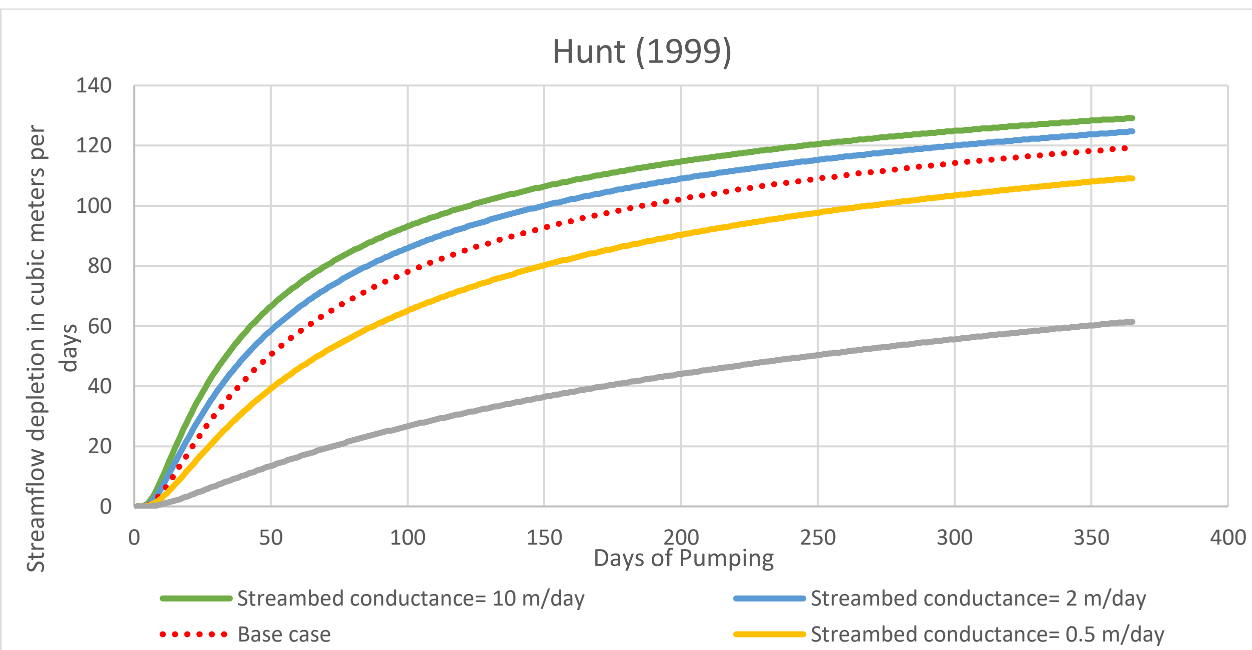


Figure 5-8: Hunt (1999) model solution for various streambed conductance with hydraulic conductivity of streambed, streambed width and streambed thickness model parameters.

Summary

Relatively high groundwater movement through the streambed lowers the impedance of streamflow depletion with rapid streamflow depletion at the early time of pumping.

5.2.7 Pumping period

An extended pumping period of 60 years (21 900 days) was simulated using four analytical solutions, from the base case scenario of 1 year (Figure 5-9). Over time with continuous pumping, the Jenkins (1968) model results depict that the streamflow depletion rate reaches equilibrium with a maximum rate of 166 m^3/day which is 96% of the maximum abstraction rate and designated as depletion-dominated supply. Hantush's (1965) and Hunt's (1999) analytical models also converge to a similar equilibrium rate of streamflow depletion.

The Hunt (2003) steadily increases over the simulation period of pumping with maximum streamflow of 155 m^3/day (Figure 5-9), which is 90% of the maximum. The Hunt (2003) model simulated streamflow depletion taking longer to reach equilibrium. The presence of the aquitard greatly reduces and delays the interaction between the river and the aquifer.

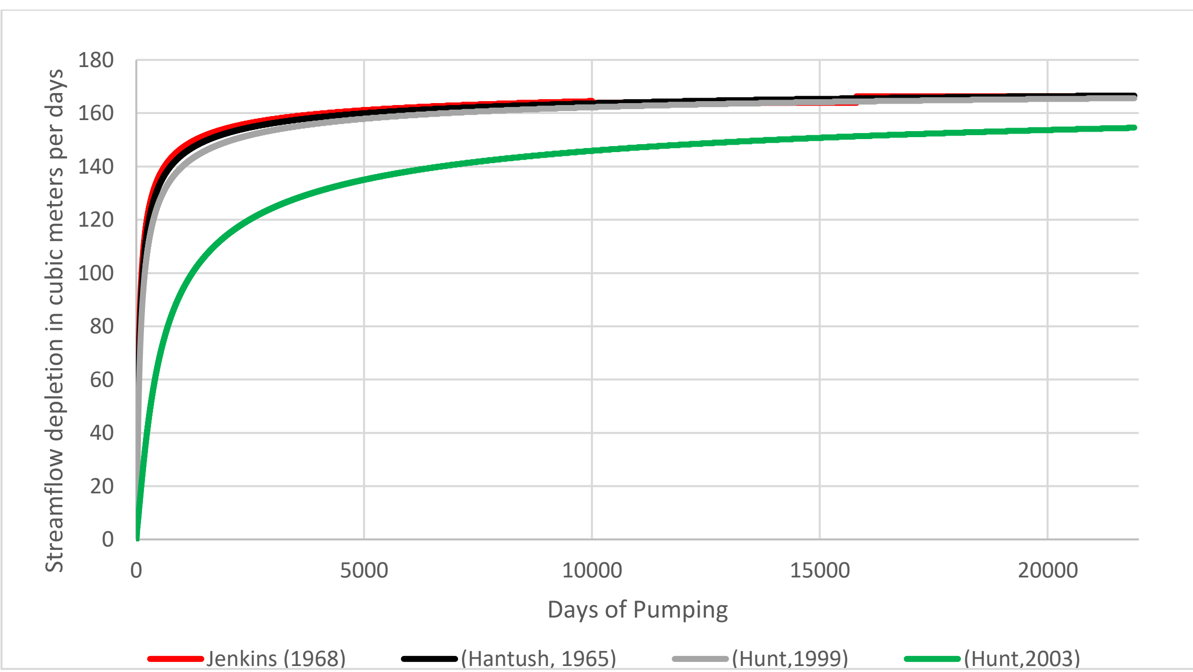


Figure 5-9: Modelled streamflow depletion from 60-year pumping period based on the developed conceptual model for the Jenkins (1968), Hantush (1965), Hunt (1999) and Hunt (2003) analytical models.

5.3 Statistical analysis of sensitivity results

One of the research objectives is to provide information to groundwater managers on which parameters are the most important to focus data collection efforts, where groundwater abstraction takes place near surface bodies. To generate information to inform decisions, the results from the sensitivity analysis are statistically analysed to determine the most significant parameters influencing the quantification of streamflow depletion.

It is observed within the previous section of parameters variability that streamflow depletion varies as the parameters vary. However, this section gives an overview of parameters' significant rankings, from the most significant parameter that is attributed to streamflow depletion to the least significant parameter (**Error! Reference source not found.**). Whisker plots are shown to describe the range of maximum s treamflow depletion and variation at each case scenario within the same axis for comparison. Abstraction rate is shown to be the largest influence on streamflow depletion with an increasing exponential effect as the abstraction rate increased (Figure 5-10).

The variance mean of the spread between streamflow depletion rate for the distance between stream and borehole is shown to increase with decreasing distance from the stream, while for transmissivity the variance for each case scenario slightly increases as transmissivity increases, however, there exists a steady state of maximum streamflow depletion.

The total data set spread distribution for transmissivity is found to be larger than that for the distance, however, it can be shown within the whisker plots that the distance is much more sensitive to small distances and transmissivity is more sensitive to large transmissivity values. The visualization of whisker plots can show how parameters influence streamflow depletion due to the variance of the data set.

Table 5-3: Model scenario containing the relevant variables (e.g., parameters, model results)

Model	Scenario: Distance 10 m	Scenario: Distance 100 m	Base case: Distance 500 m	Scenario: Distance 1 000 m	Scenario: Distance 2 000 m	Scenario: Distance 5 000 m
Jenkins (1968)	172	164	130	92	37	0,2
Hantush (1965)	168	160	126	89	35	0,2
Hunt (1999)	160	152	119	83	32	0,2
Hunt (2003)	137	77	56	9	1	0
Model	**Scenario: Abstraction rate (0.2 l/s)**	**Scenario: Abstraction rate (1 l/s)**	**Base case: Abstraction rate (2 l/s)**	**Scenario: Abstraction rate (4 l/s)**	**Scenario: Abstraction rate (20 l/s)**	
Jenkins (1968)	13	65	130	261	1303	
Hantush (1965)	12.7	63	126	252	1263	
Hunt (1999)	12	60	119	238	1193	
Hunt (2003)	6	28	56	112	558	
Model	**Scenario: Transmissivity 7 m2/day**	**Scenario: Transmissivity 35 m2/day**	**Base case: Transmissivity 70 m2/day**	**Scenario: Transmissivity 150 m2/day**	**Scenario: Transmissivity 700 m2/day**	
Jenkins (1968)	56	114	130	144	159	
Hantush (1965)	48	108	126	141	158	
Hunt (1999)	54	106	119	127	123	
Hunt (2003)	3	38	56	69	70	
Model	**Scenario: Unconfined (0.3)**	**Base case scenario (0.02)**	**Scenario: Confined (0.005)**			
Jenkins (1968)	151	130	39			
Hantush (1965)	149	126	32			
Hunt (1999)	92	119	24			
Hunt (2003)	33	56	15			
Model	**Scenario: Period (21900 days, 60 years)**	**Base case: Period (365 days)**				
Jenkins (1968)	166	130				
Hantush (1965)	167	126				
Hunt (1999)	166	119				
Hunt (2003)	155	56				

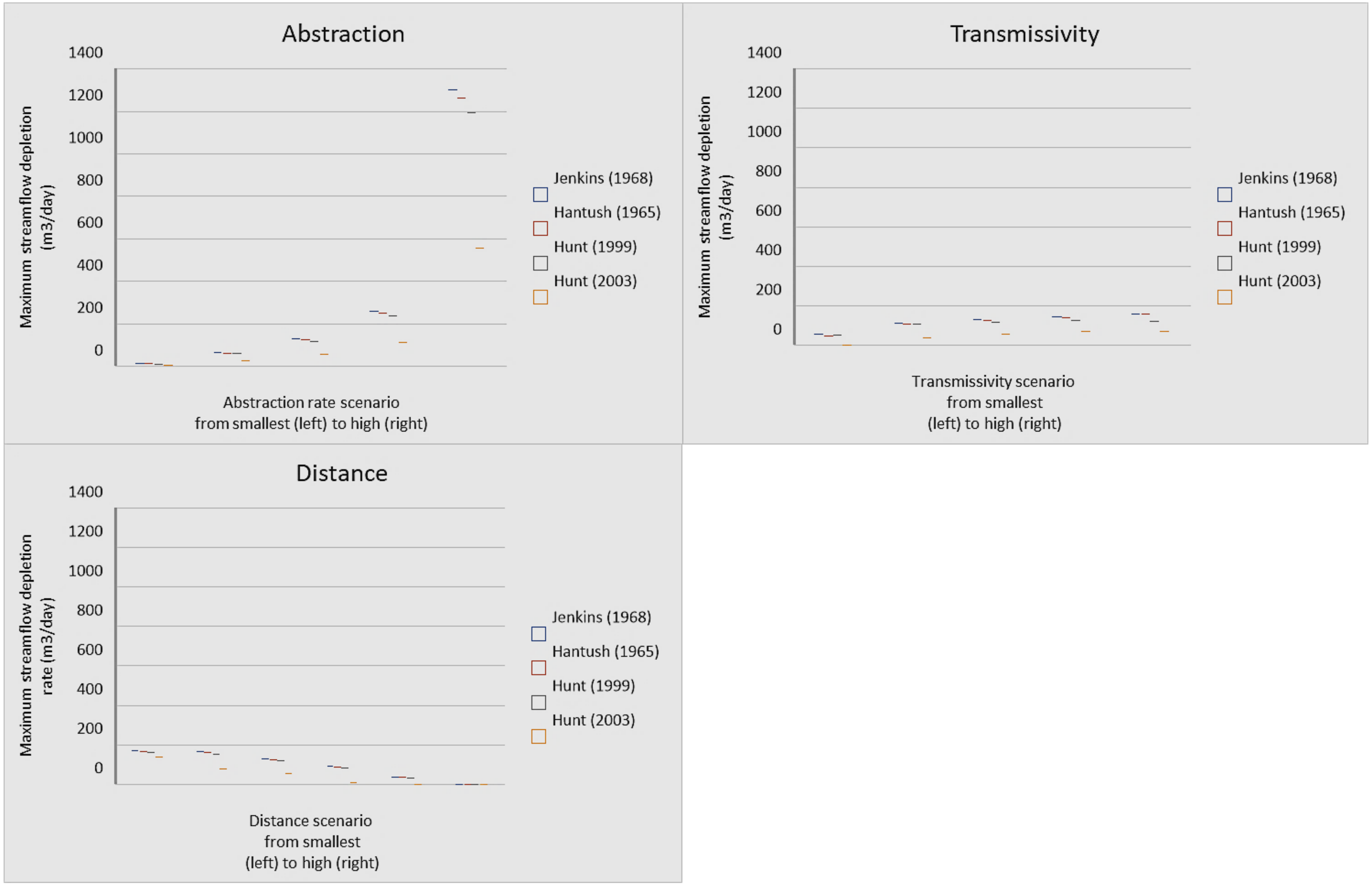


Figure 5-10: Whisker plots showing each scenario and model of the data's distribution and variability.

To further understand the relative sensitivity of each parameter, a statistical approach is considered, using a p-value approach to determine the significant difference between scenarios. If p-value < α (0.05), the null hypothesis is rejected and concludes that there is a significant difference. The ANOVA (Analysis of variance) test is used because multiple parameters need to be assessed (i.e. when comparing more than two scenarios or models). The interpretation of the p-value was used to indicate the probability of observing the observed data or more extreme data if the null hypothesis is true. If p-value < α (chosen significance level of 0.05), the null hypothesis is rejected and concludes that there is a significant difference.

The abstraction parameter is considered to be the most significant (sensitive parameter) to affect the simulated streamflow depletion because it has the lowest p-value (4.54E-07) (Table 5-4). Therefore, the management and recommendation (safe yield) must be taken into consideration for water managers. Groundwater abstraction rates should be monitored, and data collection should be accurately acquired to give out the best recommendation as this can greatly impact the surface water. The abstraction rate is the main force controlling streamflow depletion, which is expected, as the main stress applied to the system.

Table 5-4: ANOVA (Analysis of Variance): comprising of abstraction rates scenarios using four (4) analytical models.

Descriptive statistics						
Groups	Count	Sum	Average	Variance		
Scenario A: Abstraction rate (0.2 l/s)	4	43.7	10.93	10.96		
Scenario A: Abstraction rate (1 l/s)	4	216	54	304.67		
Base case: Abstraction rate (2 l/s)	4	431	107.8	1210.92		
Scenario A: Abstraction rate (4 l/s)	4	863	215.8	4873.60		
Scenario A: Abstraction rate (20 l/s)	4	4317	1079.3	122822.9		
ANOVA						
Source of Variation	SS	df	MS	F	p-value	F crit
Between Groups	3180640	4	795160.1	30.77	4.52E-07	3.06
Within Groups	387669.1	15	25844.61			
Total	3568309	19				

Distance separation is less sensitive compared to abstraction rate, with a p-value of 1.32E-06 (Table 5-5). Distance separation also needs to be considered when drilling boreholes closer to the surface water bodies, as this could potentially affect groundwater. Water managers must buffer surface water bodies in relation to the allocation of groundwater, under the NWA (Act 36 of 1998), based on the 28 Principles for

Water Management in South Africa, the water resource is regarded as "indivisible", and there is no distinction between abstraction from a river (surface water), which may have an impact on groundwater, especially where the surface water feeds the aquifers, and abstraction from an aquifer, where the aquifer may contribute to the baseflow in the river.

In both cases, a proper integrated hydrological and geohydrological study is required, which will indicate whether such impacts exist, and if they do, what the risk (probability of occurrence and severity or magnitude of the effect) will be. If the risk of such impacts occurring is unacceptable, proper mitigation measures to prevent the risk from occurring have to be proposed and implemented to prevent the risk from occurring.

The setting of "buffer zones" is a redundant water governance approach, as all impacts and systems depend on site-specific circumstances. This research gives a basic understanding of how distance proximity could potentially affect streamflow depletion.

Table 5-5: ANOVA (Analysis of Variance): comprising of distance separation scenarios using four analytical models.

Descriptive statistics						
Groups	Count	Sum	Average	Variance		
Scenario: Distance 10 m	4	637	159.25	244.92		
Scenario: Distance 100 m	4	553	138.25	1692.25		
Base case: Distance 500 m	4	431	107.75	1210.92		
Scenario: Distance 1 000 m	4	273	68.25	1574.25		
Scenario: Distance 2 000 m	4	105	26.25	287.58		
Scenario: Distance 5 000 m	4	0.6	0.15	0.01		
ANOVA						
Source of Variation	SS	df	MS	F	P-value	F crit
Between Groups	79123.33	5	15824.67	18.95197	1.32E-06	2.78
Within Groups	15029.78	18	834.99			
Total	94153.11	23				

Transmissivity and storativity are less sensitive because of their p-values (0.02 and 0.03), see Table 5-6 and Table 5-7 are larger than abstraction and distance of separation, however, they still show a significant difference (p-value < 0.05) and thus they are significant in determining streamflow depletion.

Detailed information on site-specific conditions should be known using aquifer testing to properly quantify aquifer parameters (storativity and transmissivity) to understand the effect of aquifer parameters on streamflow depletion.

Table 5-6: ANOVA (Analysis of Variance): comprising of transmissivity scenarios using four analytical models.

Descriptive statistics						
Groups	Count	Sum	Average	Variance		
Scenario A: Transmissivity 7 m^2/day	4	161	40.25	628.25		
Scenario A: Transmissivity 35 m^2/day	4	366	91.5	1283.67		
Base case: Transmissivity (70 m^2/day)	4	431	107.75	1210.92		
Scenario A: Transmissivity 150 m^2/day	4	481	120.25	1222.25		
Scenario A: Transmissivity 700 m^2/day	4	510	127.5	1749.67		
ANOVA						
Source of Variation	SS	df	MS	F	P-value	F crit
Between Groups	19344.7	4	4836.18	3.97	0.02	3.06
Within Groups	18284.25	15	1218.95			
Total	37628.95	19				

Table 5-7: ANOVA (Analysis of Variance): comprising of storativity scenarios using four analytical models.

Descriptive statistics						
Groups	Count	Sum	Average	Variance		
Unconfined (0.3)	4	425	106.25	3132.92		
Base case scenario (0.02)	4	431	107.75	1210.92		
Confined (0.005)	4	110	27.5	107		
ANOVA						
Source of Variation	SS	df	MS	F	P-value	F crit
Between Groups	16858,5	2	8429.25	5.68	0.023	4.26
Within Groups	13352.5	9	1483.61			
Total	30211	11				

The pumping period has no significant difference where the p-value (0.05) (Table 5-8) is more than the chosen significant level of 0.05, therefore; the Null Hypothesis (H0) hold: (There is no significant difference between the scenarios of pumping period variation). This is because no variation of any parameter, but only extending the time simulation. Over time the simulated streamflow depletion with reach a maximum and then equilibrate with the system.

Table 5-8: ANOVA (Analysis of Variance): comprising of pumping period scenarios using four analytical models.

Descriptive statistics						
Groups	Count	Sum	Average	Variance		
166	3	488	162.67	44.33		
130	3	301	100.33	1486.33		
ANOVA						

Source of Variation	SS	df	MS	F	P-value	F crit
Between Groups	5828.17	1	5828.17	7.62	0.05	7.71
Within Groups	3061.33	4	765.33			
Total	8889.5	5				

5.3.1 Summary

Statistical analysis using the ANOVA test and using visualisation of box plots to identify potential outliers concluded that abstraction rate is the largest influencer or rather sensitive parameter that affects streamflow depletion due to its smallest p-value. P-value analysis (quantification of variation) using the ANOVA test further confirms that the period of abstraction has no influence on streamflow depletion as the p-value is greater than significance interval (>0.05) which concludes that alternative hypotheses hold (there exists no significance within the data sets). Period of abstraction does not influence streamflow depletion since no parameter is varied but rather extends the time simulation.

Chapter 6 NUMERICAL MODELS

Numerical models are developed with the aim of improving the understanding of groundwater abstractions on streamflow depletion and evaluating the analytical model used to quantify streamflow depletion. The goal is to quantify the effects of groundwater pumping on streamflow based on the conceptual hypothetical model. in Chapter 5. This is done in two approaches, firstly under analytical conditions, where the numerical models are designed to represent the simplifying assumptions of the analytical models, And, secondly, under a complex model condition that is more representative of a real groundwater system.

The knowledge of analytical models for estimating streamflow depletion is restricted to the study of idealized situations, where a number of the real groundwater system's complexities are either disregarded or approximated through the application of simplifying assumptions. (i.e. hypothetical conceptual models). Compared to analytical solutions, numerical models offer a more accurate depiction of aquifer conditions since they handle numerous complexities and are not as limited by oversimplifying assumptions about the actual groundwater system. Thus, the comparison of both the analytical and numerical models are relative standard of measurements for gaining insight into the predictability and limitations of the analytical models.

6.1 Approach 1 – Analytical model conditions

The numerical model is developed based on simulated analytical solutions to understand the implications of the analytical model assumptions to streamflow depletion. Each conceptual model from Jenkins (1968), Hantush (1965), Hunt (1999), and Hunt (2003) is represented by the numerical model for comparative reasons. The numerical model setup is reliant on the conceptual hypothetical model.

6.1.1 Model Comparison for Jenkins (1968)

Numerical model setup

The numerical model is developed using PMWIN (Processing MODFLOW for Windows) which utilises MODFLOW-2000 to simulate the groundwater system. For details on MODFLOW and the software program used, please refer to McDonald and Harbaugh (1988) and Harbaugh (2005).

The numerical model is designed to represent the hypothetical conceptual model for the Jenkins (1968) analytical model (Figure 4-1). The model set-up is based on the base case scenario of the hypothetical conceptual model with the analytical solution as follows:

- The river is represented as a constant head boundary condition, which will mimic a fully penetrating stream (Figure 6 1).
- The assigned abstraction borehole on the model is situated 500 m away from the constant head boundary condition as it is on the hypothetical conceptual model
- Large model extent to represent the assumption of infinite lateral extent associated with the analytical model.
- Hydraulic properties of the model are the same as assigned for the corresponding analytical solution (Table 4-1).
- The model makes use of a finite difference uniform grid with rows and columns of evenly spaced model elements of 200 by 10 m.
- All parameters assigned are constant one-layer model.
- The aquifer is confined.
- The transient groundwater flow model was developed with an abstraction rate of 172.8 m^3/day. The model runs with a single period number and a total simulation period of 365 days (1 year). Recharge into the model is not taken into consideration.

It should be noted that other boundaries are set as non-flow boundaries because, in modelling a specific area, the model domain can be defined such that beyond a certain point, the groundwater flow is negligible or irrelevant to the study. Setting these boundaries as non-flow ensures that the model focuses only on the significant flow within the domain of interest.

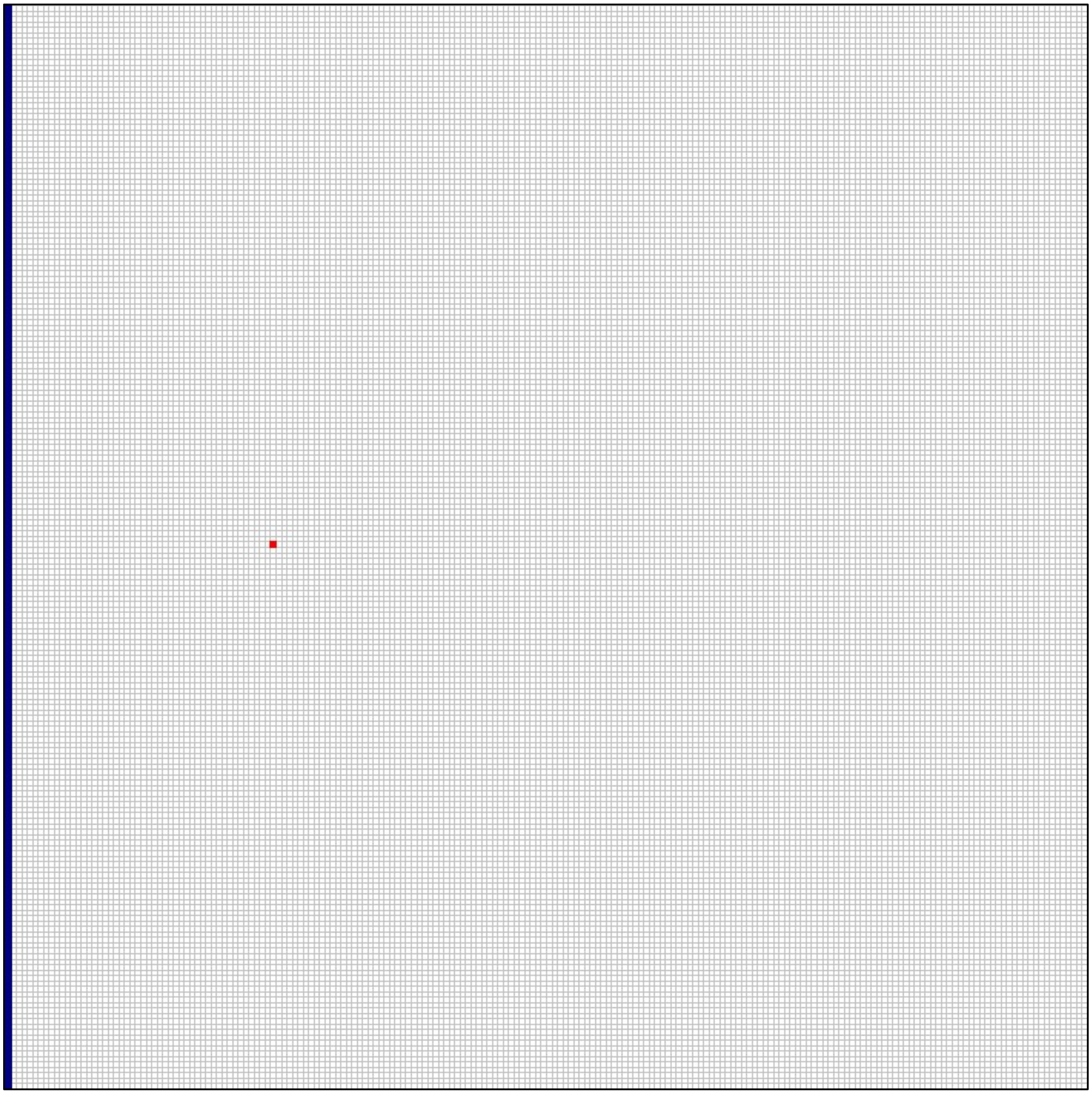


Figure 6-1: Numerical model base setup, showing the finite difference grid, and stream (constant head boundary condition (blue) and abstraction well (red) locations.

Simulation results

The simulated hydraulic head distribution at the end of the simulation is shown in Figure 6-2. The simulated head at the borehole is 46.6 m; lower than 50 m head of the river stage (constant head boundary condition). PMPATH is used to show a simulated water table and drawdown at the pumping borehole (Figure 6-2). The amount of water induced into the model by boundary conduction was measured using the water budget as streamflow depletion rates across the simulation period (365 days).

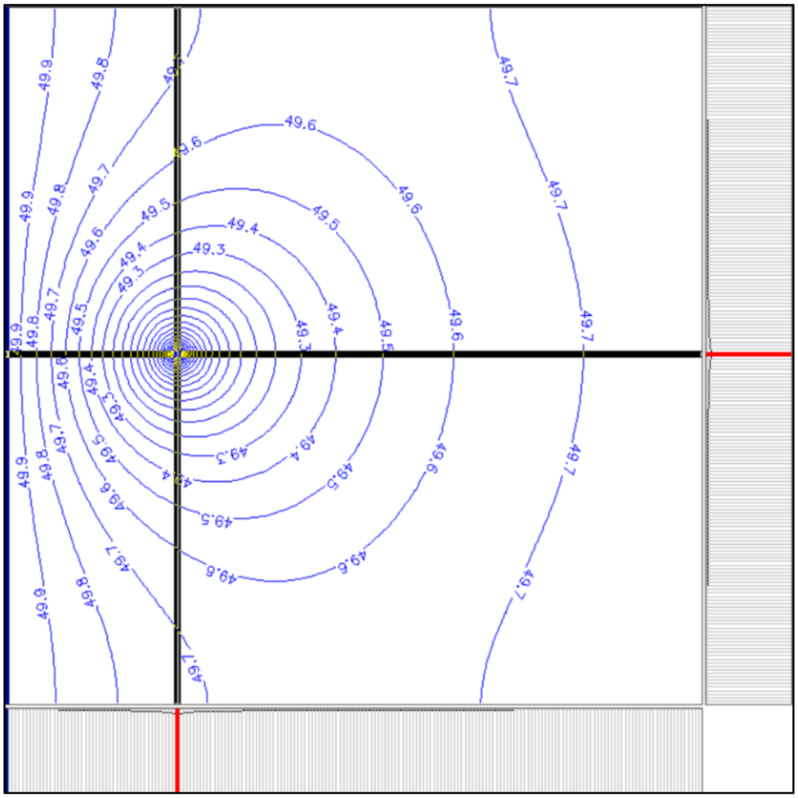


Figure 6-2: PMPATH cross section of simulated water table (model has a flat-water table that is only affected by the abstraction at the well, forming a drawdown cone).

Comparison with analytical model

Figure 6-3 shows the change in water budget rates over the simulation period for the case of pumping borehole which is the streamflow depletion simulated by the numerical model being compared to the analytical solution of Jenkins (1968). It is shown in Figure 6-3 that numerical model as designed to represent the analytical model shows less streamflow depletion rates as compared to the analytical solution and the following assumptions can be the contributing factors:

- Numerical models are approximations and the discrepancy seen could be a function of the discretisation error associated with numerical models for both space (grid size) and time (time steps). Refining the numerical model around the borehole slightly changes the results of the model simulation.
- The analytical model assumes that the aquifer extends infinitely long from the stream, however, a numerical model is bounded by no-flow boundary conditions (north, east, and south), thus less water available from the aquifer and thus technically more streamflow depletion would have been simulated.

- The numerical model uses the confined aquifer layer type, which was originally developed for the Theis equation, however, analytical solutions are applied to unconfined aquifers in practical application, but the model is built upon Theis mathematical equations which assumes a confined aquifer.

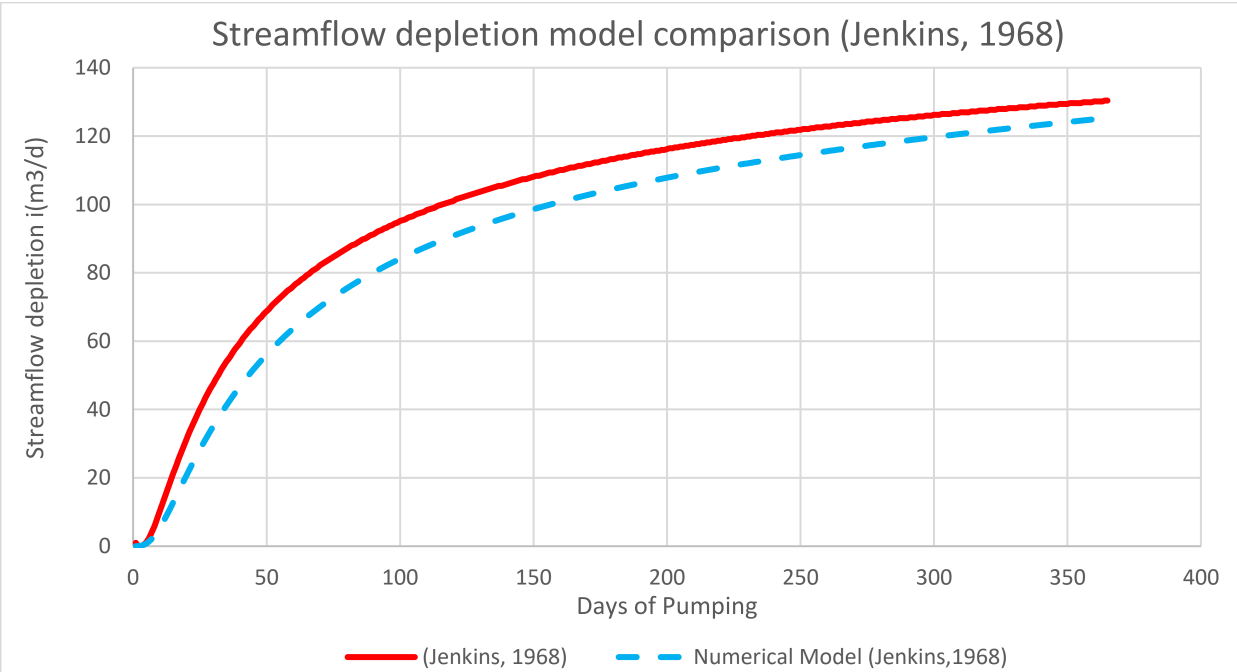


Figure 6-3: Modelled streamflow depletion from continuous pumping for both analytical and numerical model (Jenkins, 1968)

6.1.2 Model Comparison for Hantush (1965)

Numerical model setup

The numerical model is designed to represent the hypothetical conceptual model for the Hantush (1965) analytical model (Figure 4-1). The same setup as the one for Jenkins (1968) was used, however, 3rd type boundary condition (river package) instead of the constant head boundary condition was set up to account for streambed leakance within the model as the Hantush (1965) analytical model includes. A single layer is still used because the analytical model does not yet account for riverbed bottom elevation. The introduction of streambed leakance was incorporated into the river package as an elevation of the riverbed bottom value of 0 m (representing a fully penetrating river).

Simulation results

Figure 6-4 shows the simulated hydraulic heads at the end of the simulation period of 365 days. The simulated head at the borehole is 46.6 m lower than 50 m head of the river stage (constant head boundary condition) Over time, the streamflow is affected by borehole abstraction. The water budget was used to quantify the volume of water induced from the river leakage into the model as streamflow depletion rates over the simulation period (365 days).

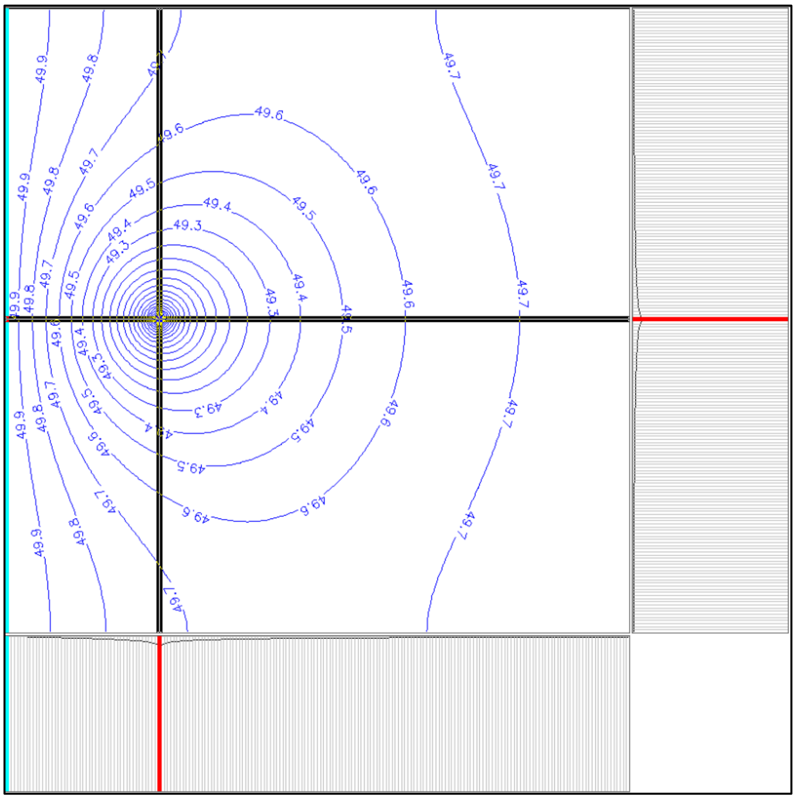


Figure 6-4: PMPATH cross section of simulated water table (model has a flat-water table that is only affected by the abstraction at the well, forming a drawdown cone).

Comparison with analytical model

Figure 6-5 illustrates the streamflow depletion model outcomes from continuous groundwater abstraction in a hypothetical conceptual model.

The numerical solutions show a better match with the analytical solution (Hantush, 1965) than for the Jenkins (1968) model. The two solutions are in close agreement for the first 30 days of pumping. After that, they start to diverge, with the differences diminishing as steady-state conditions are approached after about 1 year (365 days) of pumping.

The 3rd boundary condition better represents a leakance factor than the constant head boundary condition representing fully penetrating steam as a result, the numerical model shows a better comparison with the analytical solution (Hantush, 1965).

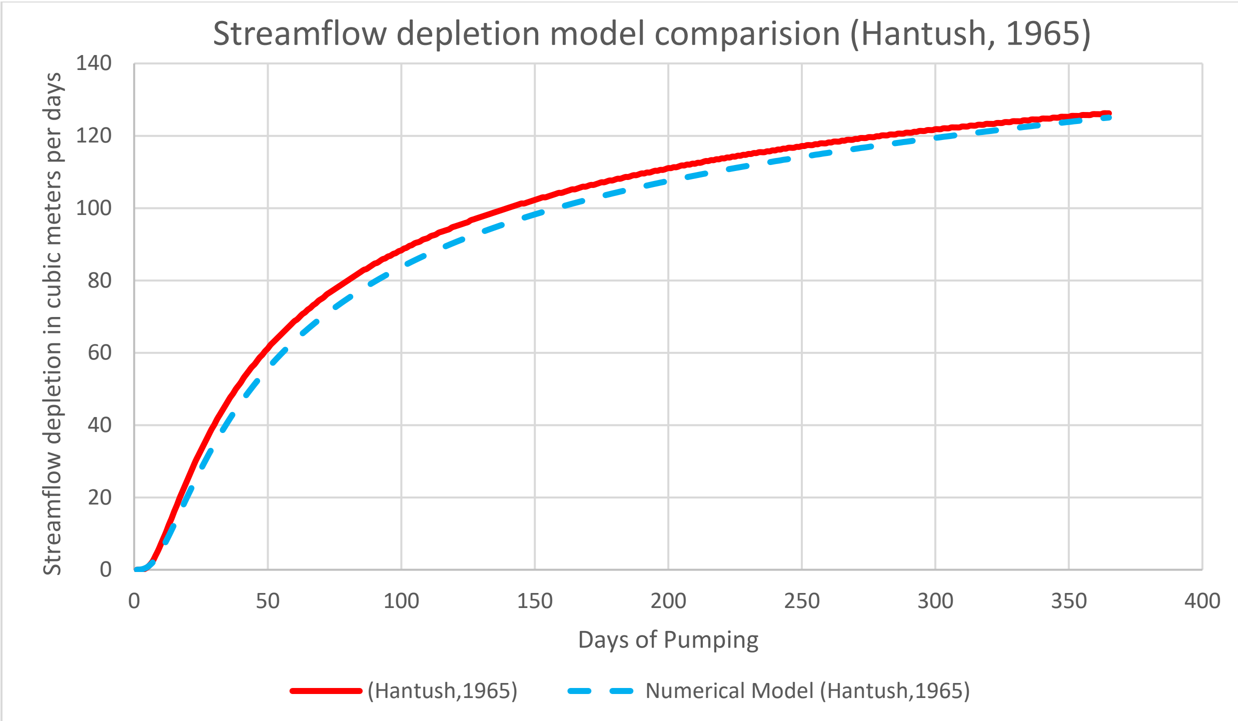


Figure 6-5: Modelled streamflow depletion from continuous pumping for both analytical and numerical model (Hantush, 1965)

6.1.3 Model Comparison for Hunt (1999)

Numerical model setup

The numerical model is designed to represent the hypothetical conceptual model for the Hunt (1999) analytical model (Figure 4-1). The numerical model makes use of a 3rd type boundary condition at the river, similar to the results in Section 6.1.2 however, the model layer is constructed as a two-layer model. Thus, the river package is only assigned to the first layer to imitate the analytical solution assumption of a partially penetrating stream. Furthermore, the abstraction borehole is assigned to the bottom layer. Hydraulic parameters for both layers were constant to represent the hypothetical conceptual model.

Simulation results

Figure 6-6 shows the simulated hydraulic heads at the end of the simulation period of 365 days. Due to borehole abstraction, the simulated gradient is from the boundary representing the river. The simulated head at the borehole is 48 m lower than 50 m head of the river stage (constant head boundary condition)

The output water budget quantification was used to simulate the volume of water from the boundary (river leakage) into the model to quantify streamflow depletion for each simulated time step.

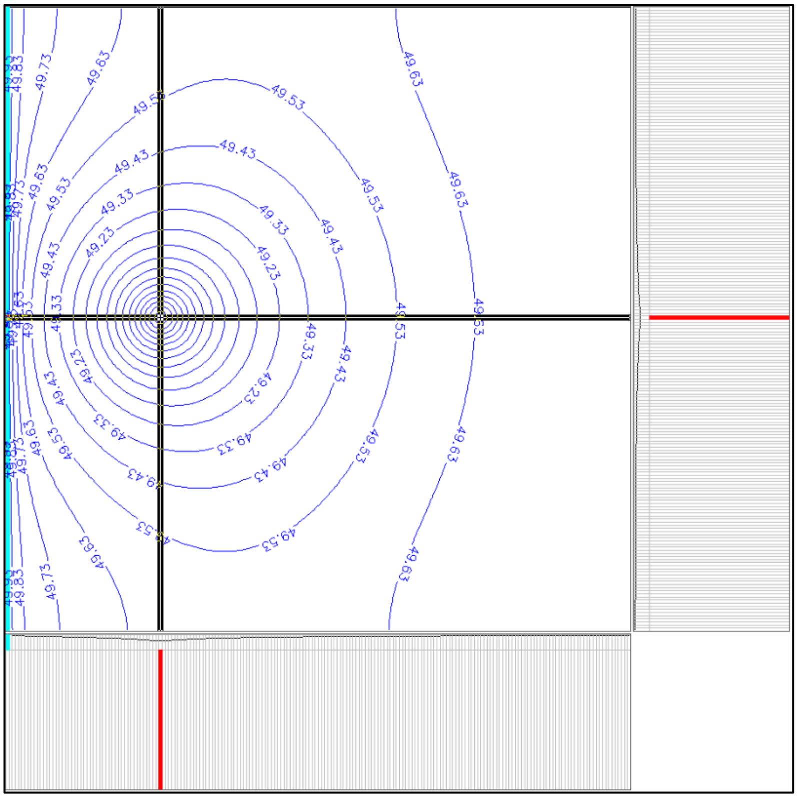


Figure 6-6: PMPATH cross section of simulated water table (model has a flat-water table that is only affected by the abstraction at the well, forming a drawdown cone).

Comparison with analytical model

Figure 6-7 shows that two solutions (numerical and analytical solution) are in close agreement for the first 11 days of pumping, after that, they begin to diverge, with the numerical solution simulating less streamflow depletion than the analytical model at later times. The largest difference, at 365 days (1 year), is about 18 m^3/day, with the numerical model value being about 15% less than the analytically calculated value. The following might be the contributing factor of significant variations in comparison between numerical model and analytical solution:

- The effect of pumping borehole at the lower layer can cause a delay repose of streamflow depletion, thus less streamflow depletion compared to the one simulated by analytical solution (Hunt, 1999).
- Difference in parameters (river package within the numerical model is largely dependent on the length of the river within the cell of the model, where else the analytical solution only considers streambed conductance and stream width to quantify streamflow depletion).

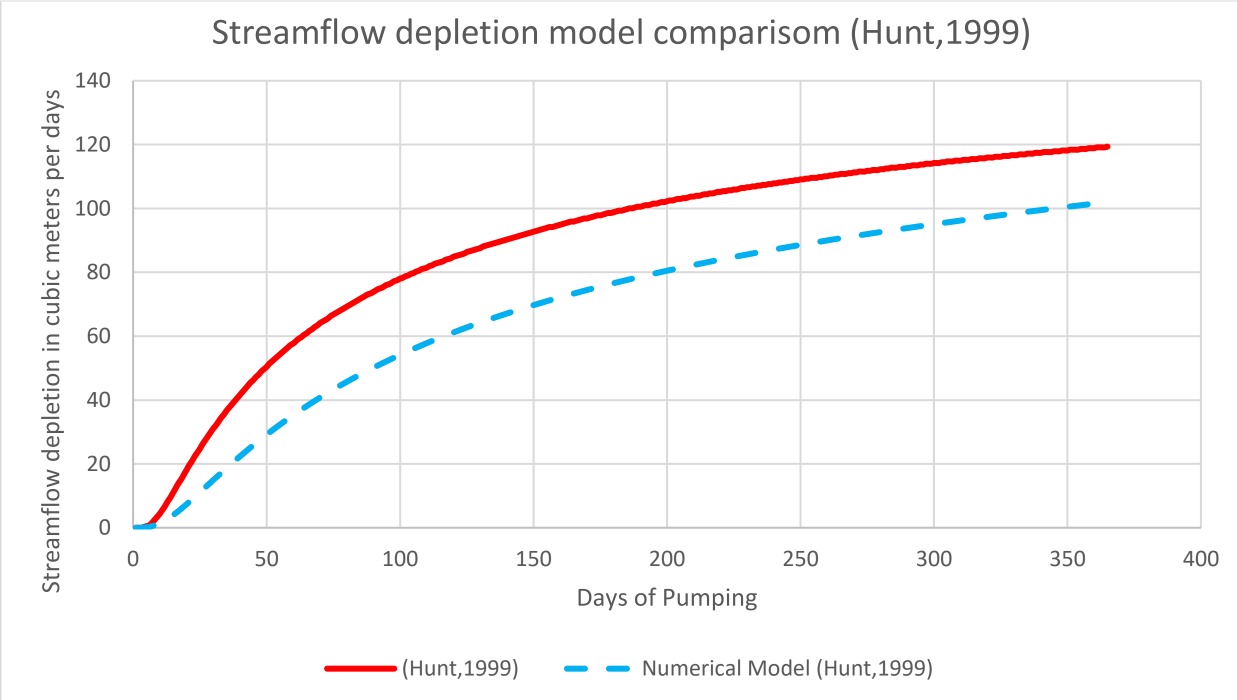


Figure 6-7: Modelled streamflow depletion from continuous pumping for both analytical and numerical model (Hunt, 1999)

6.1.4 Model Comparison for Hunt (2003)

Numerical model setup

The numerical model is designed to represent the hypothetical conceptual model for the Hunt (2003) analytical model (Figure 4-1). The setup of the model is the same as the previous model in Section 6.1.3, except that the upper layer is conceptualised as an aquitard to mimic the assumptions made for the Hunt (2003) analytical solution.

Simulation results

The simulated head at the borehole is 47 m lower than 50 m head of the river stage (constant head boundary condition (Figure 6-8). The simulated gradient is from the partially penetrating river within the first layer towards the direction of the pumping borehole. The water budget was used to quantify the volume of water induced from the boundary (river league) into the model to measure streamflow depletion rates.

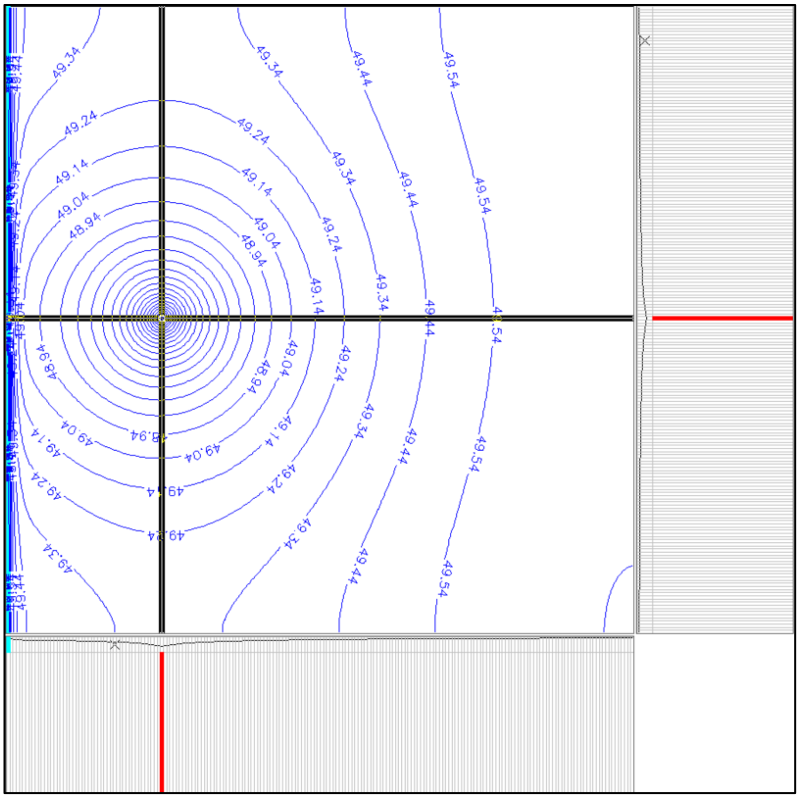


Figure 6-8: PMPATH cross section of simulated water table (model has a flat-water table that is only affected by the abstraction at the well, forming a drawdown cone).

Comparison with analytical model

Figure 6-9 shows that the analytical solution simulates slightly higher streamflow depletion rates at an early time, after that they begin to diverge, with the numerical model simulating larger values of streamflow depletion than the analytical solution. The largest difference, at 365 days (1 year), is about 5 m^3/day, with the numerical model value being about 8% more than the analytically calculated value. The following might be the contributing factor of significant variations in comparison between numerical model and analytical solution:

- The effect of pumping borehole at the lower layer and inclusion of the aquitard layer can cause a delay repose of streamflow depletion, thus less streamflow depletion compared to the one simulated by analytical solution (Hunt,2003).
- The two-layer numerical model considers the vertical movement of groundwater (3D model), while the analytical solution only considers the horizontal (2D) flow of groundwater within the aquifer.

- Thickness of the aquitard layer (first layer) is approximated based on the elevation (top of the layer) of the model as compared to the analytical solution.
- The numerical model uses the confined aquifer layer type, which was originally developed for the Theis equation, however, analytical solutions are applied to unconfined aquifers in practical application, but the model is built upon the Theis mathematical equations which assume a confined aquifer.
- Difference in parameters (river package within the numerical model is largely dependent on the length of the river within the cell of the model).

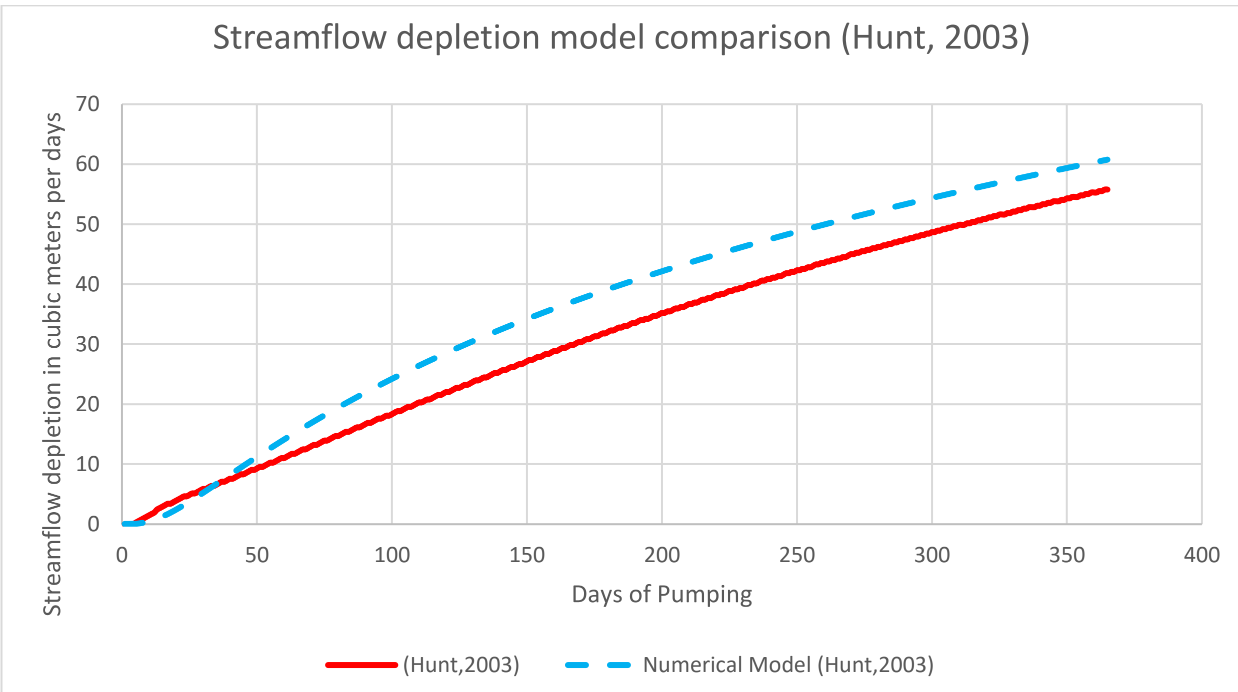


Figure 6-9: Modelled streamflow depletion from continuous pumping for both analytical and numerical model (Hunt, 2003)

6.2 Approach 2 – Complex model conditions

To further understand the applicability of the analytical models and their limitations, a complex system which closer represents a real-world groundwater system is considered for comparison with a numerical model. The model addresses a partially penetrating river, variable boundary conditions, variable river stage, variable elevations and a heterogeneous subsurface.

6.2.1 Conceptual model

The model is based on a tutorial model provided with the PMWIN software. The model area is a river valley bounded by impermeable outcrops to the north and south, and groundwater flow from west to east. The unconfined model has three hydrostratigraphic units, where each layer is homogeneous and anisotropic. There is an upper aquifer, underlain by an aquitard, and a lower aquifer. There is a single borehole pumping from the bottom of the aquifer at a rate of 500 m^3 /d (Figure 6-10).

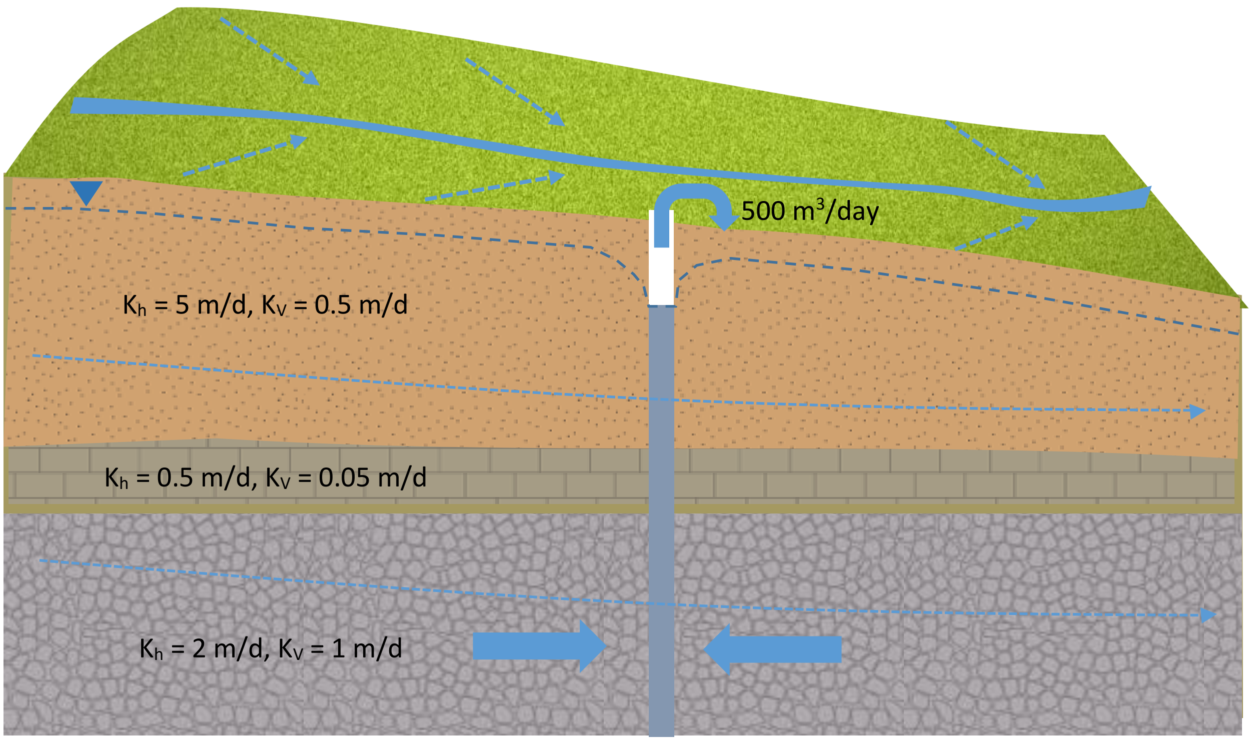


Figure 6-10: Hypothetical model depicting a complex aquifer - stream conceptualisation for stream flow depletion.

6.2.2 Setup

The cell size of the model set-up is 250 m, thus the model domain is represented by 27 columns (27 x 250 m = 6 750 m) and 20 rows (20 x 250 m = 5 000 m). The Constant head boundary is assigned to the west and east and the no-flow boundary is to the north and south boundary of the model. Three layers have been assigned to the model with the first layer assigned as an unconfined aquifer, and both second and third layers have been assigned as confined/unconfined (transmissivity varies) which then allows the model to decide whether the layers are confined or unconfined based on the simulated heads as well as the top layer.

A 3rd type boundary condition (river package) was set up to account for the hydraulic conductance of the riverbed, head in the river and elevation of the riverbed bottom. For a 365-day simulation period, the transient flow model was run with 500 m^3/day of pumping using the initial hydraulic heads from the steady-state flow model.

6.2.3 Simulation results

Figure 6-11 shows the simulated head distribution of the model with the drawdown pronounced at the pumping borehole. The simulated hydraulic head distribution illustrates groundwater flowing from west to east within the river valley as well towards the river, which is altered by pumping. On the western side, groundwater originally flows towards the river (gaining stream), but the influence of the abstracting borehole causes the water to be induced from the river into the aquifer by groundwater abstraction. The water budget is used to quantify streamflow depletion using the rates of river leakage into the model within each time step of the simulation.

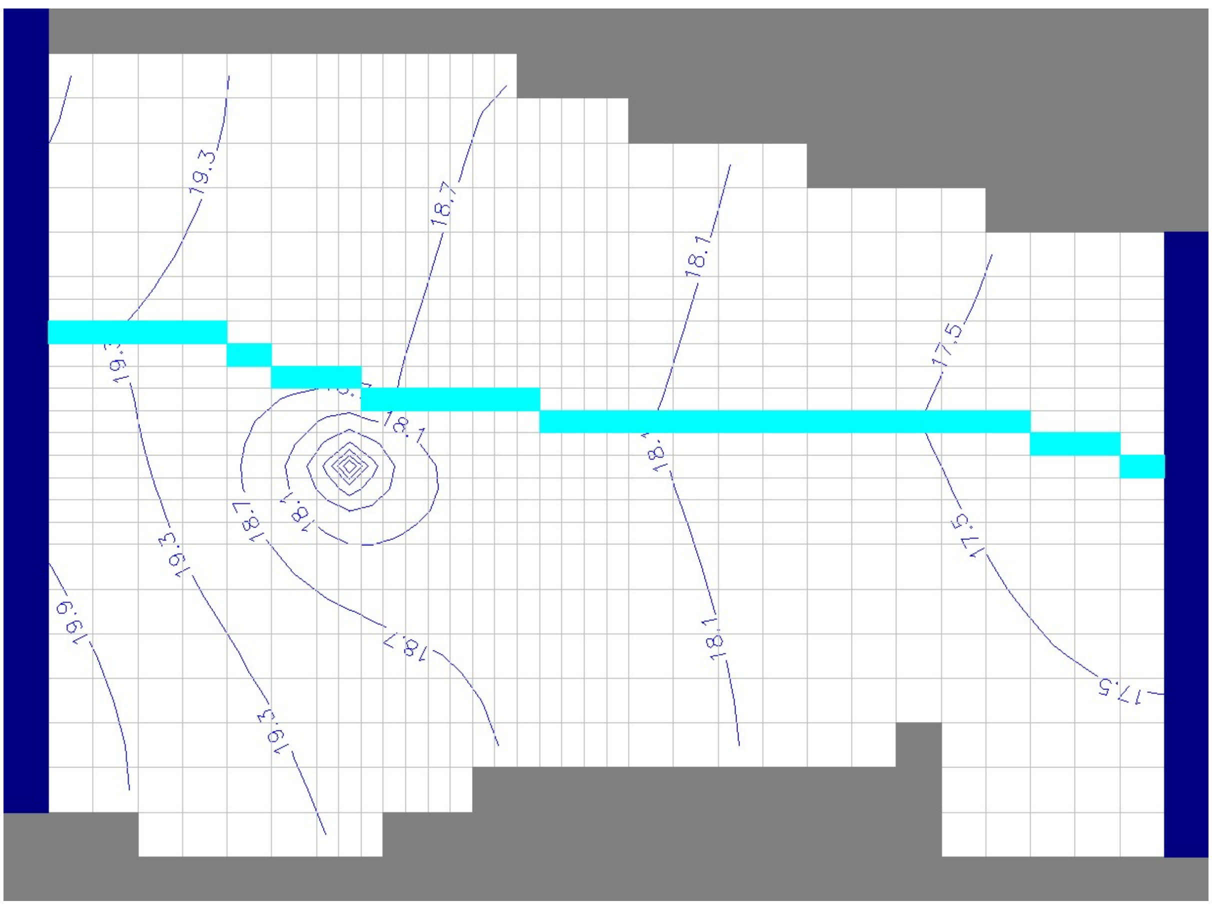


Figure 6-11: Drawdown resulting from pumping borehole one (1) at the end of the simulation period (non-flow boundary condition to south and north, constant head boundary condition to east and west, 3rd type boundary condition assigned at the top layer of the model and assigned abstraction borehole).

6.2.4 Comparison with analytical model

Figure 6-12 illustrates model estimates of streamflow depletion from continuous groundwater abstraction. Because the pumping borehole is a further distance away from the river (1000 m), the analytical model predicts a delayed response in streamflow depletion reaching 50% or more of the pumping rate within 50 days. In comparison to the numerical model, Jenkins (1965), Hantush (1965) and Hunt (1999) overestimate streamflow depletion at late time. The following factors might be contributing:

- Jenkins (1968) assumes a fully penetrating stream with the aquifer from which abstraction is taking place.
- Jenkins (1968), Hantush (1965), and Hunt (1999) consider a single layer with homogenous aquifer materials.
- Analytical solutions cannot account for irregular boundary conditions.
- Analytical models are 2D flow and neglect vertical flow.

Hunt (2003) simulates less streamflow depletion than the numerical model due to an approximation in the analytical solution to approximate the complicated numerical model.

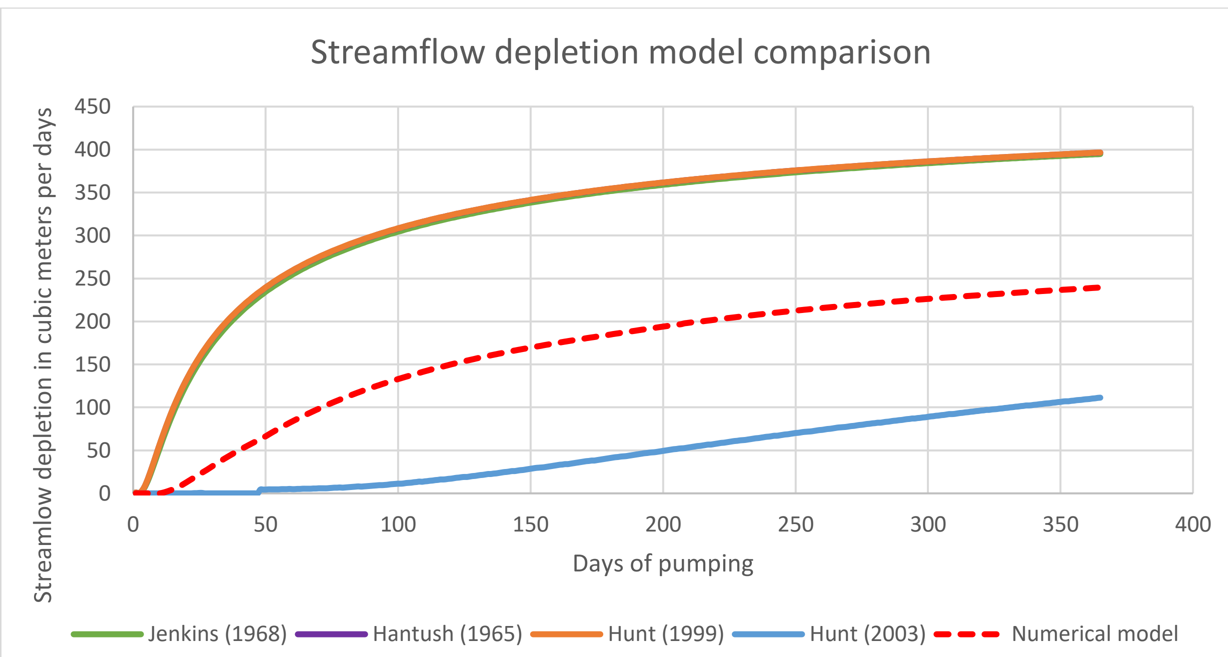


Figure 6-12: Modelled streamflow depletion from continuous pumping for both analytical and numerical model.

6.3 Summary

The comparison highlights the following points regarding the use of the analytical model:

The comparison between analytical and numerical models for the first (analytical conditions) approach serves to demonstrate that there is a correlation between these methods as long as the numerical model is designed to have the same simplifying assumptions as the respective analytical solution. The discrepancies then amounted to the mathematical differences between the solution methods.

The comparison between the analytical and numerical models for the second (complex model) approach demonstrates the limitations of these analytical models when applied to complex boundary conditions in heterogeneous, unconfined aquifers. The Jenkins (1968), Hantush (1965) and Hunt (1999) analytical models estimate more streamflow depletion than the numerical model, while the Hunt (2003) analytical model estimates less streamflow depletion. The overestimated streamflow depletion is accounted for by the limiting simplifying assumptions that cannot account for regional groundwater flow, irregular boundaries, vertical flow or unconfined conditions. The underestimate by Hunt (2003) is accounted for the over-restriction/compensation of applying an aquitard at the river itself. This approach also highlights that the discrepancy caused by these limiting assumptions is much larger than the difference between the various analytical solution formulations.

Chapter 7 SUMMARY AND DISCUSSION

The dissertation has addressed the following objectives:

- Development of hypothetical conceptual models with various aquifer-stream conceptualisations, using available data from a real-world system, to characterised GW-SW interactions with an emphasis on streamflow depletion.
- Evaluating model approaches to assess streamflow depletion from groundwater pumping on the developed hypothetical models, particularly using analytical solutions that are simple to implement.
- Developing numerical models to understand the implication of analytical model assumptions and how analytical models can simulate streamflow depletion.

7.1 Assessment of streamflow depletion

To evaluate streamflow depletion two approaches were utilised, namely analytical and numerical models.

7.1.1 Analytical models

Four analytical solutions are simulated to quantify streamflow depletion based on the developed hypothetical conceptual models. Jenkins (1968) based on its fewest possible parameters and ease to implement, the maximum streamflow depletion after 365 days of simulation period is 130 m^3/day. The Jenkins (1968) model overestimates streamflow depletion due to its assumptions. The extended solution, which is the Hantush (1965) model, introduced the impacts of a semi-impervious streambed layer, which reduced the simulated maximum streamflow depletion rate to 126 m^3/day. As the solution moves to an improved stream-aquifer conceptualisation, the Hunt (1999) model, due to its introduction of streambed impedance and partially penetrating stream further reduces streamflow depletion to 119 m^3/day. The introduction of an aquitard layer increases the delay response at early pumping time. After 365 days of continuous pumping, the volume of water induced through the streambed into the aquifer is 56 m^3/day.

Sensitivity analysis is performed to determine how changing parameters will impact the simulated streamflow depletion rate; this will help decision-makers to focus on these parameters that require more accuracy and thus focused data collection efforts. To further understand the relative sensitivity of each parameter, a statistical approach is considered using the ANOVA (Analysis of Variance). The abstraction rate is considered to be the most significant sensitive parameter because of its probability value of 4.54E-07. Distance separation is also sensitive with a probability value of 1.32E-06. Transmissivity and storativity are less sensitive to streamflow depletion with their p-values of 0.021, and 0.023, respectively. The pumping period has no significant difference where the p-value is 0.051, and this is because only the time simulation is extended but not the actual parameter.

7.1.2 Numerical models

Numerical models are applied and designed to represent the analytical solutions to assess how the solution would be implemented in a numerical model. Two approaches have been used within this dissertation namely:

- Analytical model condition, which aims at creating a numerical model to simulate analytical condition as closely as possible to understand the implication of the analytical model assumptions.
- Complex numerical model condition, considering a real-world complex numerical model and assessing how analytical solutions can represent streamflow depletion under these conditions.

The comparison between analytical and numerical models for the analytical conditions approach demonstrates a strong correlation between these models when the numerical models are designed to represent the same simplifying assumptions as analytical solutions. The discrepancies then amounted to the mathematical differences between the solution methods. Comparison between analytical and numerical models for the complex model approach demonstrates the limitations of comparisons due to the following contributing factors:

- The overestimated streamflow depletion is accounted for by the limiting simplifying assumptions that cannot account for regional groundwater flow, irregular boundaries, vertical flow or unconfined conditions.
- Restriction/compensation of applying an aquitard at the river itself for Hunt's (2003) analytical solutions.
- Discrepancy caused by these limiting assumptions is much larger than the difference between the various analytical solution formulations.

7.2 Limitations

Streamflow depletion within groundwater-surface water systems occurs in many different geological and hydrogeological conditions. However, this study predominantly operated within a simplified conceptual framework, neglecting certain intricate real-world elements. These simplifications might not fully capture phenomena such as irregular boundaries, unconfined conditions, or intricate regional groundwater flow dynamics. The discrepancy observed between analytical and numerical models partly stems from these oversights, limiting the accurate representation of actual hydrogeological scenarios.

Analytical models, although sensible for their simplicity and ease of implementation, inherently rely on a set of assumptions that might not hold in all scenarios. For instance, assumptions of infinite aquifer extent or fully penetrating streams could result in significant deviations from reality. These assumptions overlook the vertical flow components and fail to account for localised variations, leading to discrepancies when compared to more complex numerical models.

The study faced limitations due to the absence of comprehensive field data for model calibration. The hypothetical nature of the conceptual models hindered direct validation against measured observations, potentially affecting the accuracy of simulated results. In geohydrology, empirical data play a pivotal role in refining models and improving their predictive capabilities. The reliance on hypothetical scenarios without empirical calibration limits the reliability and generalisation of the findings.

Sensitivity analyses conducted in this study identified critical parameters influencing streamflow depletion rates. However, the finite ranges considered for variables and the reliance on hypothetical models might not encompass the full spectrum of variations encountered in real-world hydrogeological systems. Parameters such as transmissivity, storativity, and distance separation, while assessed for sensitivity, might exhibit varied influences across diverse geological formations, warranting a broader exploration of their ranges and interactions.

The difference observed between analytical and numerical models stems not only from underlying simplifications but also from the fundamental differences in their approaches. Analytical models, with their simplistic assumptions, may inadequately represent the complexities inherent in realistic hydrogeological systems. In contrast, numerical models, while more comprehensive, introduce discretisation and computational approximations that might lead to their own set of discrepancies and uncertainties, particularly in boundary conditions and model setup.

Understanding and quantifying the discrepancies between analytical and numerical models present some challenges, because these discrepancies often emerge due to differing conceptualisations and assumptions, making it difficult to isolate and attribute the exact sources of discrepancies observed in simulated streamflow depletion rates.

The limitations highlighted in the study significantly impact its findings in several specific ways:

- The study's reliance on a simplified conceptual framework, which does not account for complex real-world elements such as irregular boundaries, unconfined aquifers, and intricate regional groundwater flows, limits the accuracy and applicability of its conclusions. This oversimplification could result in underestimating or overestimating streamflow depletion, as the real-world complexities often play a critical role in influencing groundwater and surface water interactions. For instance, failing to consider vertical flow components might lead to inaccurate predictions of groundwater recharge and discharge dynamics.
- The assumptions inherent in analytical models, such as infinite aquifer extent or fully penetrating streams, can lead to significant deviations from real-world conditions. These assumptions neglect important factors like localised variations in aquifer properties or the presence of partial penetration, which can drastically alter the flow dynamics. As a result, the models may provide oversimplified or misleading results, particularly in heterogeneous systems where localised features strongly influence flow patterns. This can lead to incorrect assessments of water availability and streamflow depletion rates.
- The absence of comprehensive field data for model calibration means that the models were not validated against real-world observations. This lack of validation can undermine the credibility and accuracy of the simulated results, as the hypothetical scenarios used may not accurately reflect the complexities of actual hydrogeological systems. Without empirical data, there is a risk that the model parameters and predictions are not representative, leading to unreliable or generalised findings that may not be applicable in specific real-world contexts.
- The study's sensitivity analyses, while identifying critical parameters influencing streamflow depletion, were limited to a finite range of variables. This restricted exploration might not capture the full spectrum of variability found in natural systems, potentially overlooking key interactions and dependencies. For instance, the influence of parameters like transmissivity and storativity can vary greatly across different geological settings, and a narrow focus may miss significant variations that could alter the outcomes. This limitation could result in an incomplete understanding of the factors driving streamflow depletion, leading to oversimplified management recommendations.
- The fundamental differences between analytical and numerical models, particularly in how they conceptualise and handle boundary conditions, contribute to discrepancies in their outputs. Analytical models often use idealised assumptions that ignore complexities such as non-linear boundary conditions or variable aquifer properties, leading to potentially significant errors. Numerical models, despite being more detailed, can introduce errors through discretization and computational

approximations. These discrepancies make it challenging to reconcile differences in predicted streamflow depletion rates and to accurately determine the reliability of model outputs. The resultant uncertainties can complicate decision-making processes, particularly in water resource management and policy formulation.

Chapter 8 CONCLUSION AND RECOMMENDATIONS

The main findings of the study are:

- Based on the hypothetical case model, the simpler analytical models overestimate streamflow depletion more than the subsequent improved analytical solutions.
- The Hunt (2003) model has an improved conceptualisation of aquifer-stream dynamics because it accounts for the streambed influence; and therefore, a delay in the response time is seen. In cumulative volume graphs for both analytical solutions, the volume of streamflow depletion decreases as the analytical solution approaches the real-world system (i.e., improved stream-aquifer conceptualization conditions), this is due to the introduction of parameters that delay the streamflow depletion.
- The improved conceptualisation of aquifer–stream dynamics demonstrates that it requires more time to obtain steady-state streamflow depletion (i.e., Hunt, 2003 does not reach a steady-state rate of streamflow depletion than Jenkins, 1968 assuming all other factors are equal).
- For higher transmissivities, the time lag between pumping and depletions is shorter, with q/Q= 1 being approached more quickly than small transmissivities.
- As transmissivity ranges decreased depending on the geology, the more time delay of time response of streamflow depletion.
- Streambed leakage in the Hantush (1965) model and the aquitard introduced in the Hunt (2003) model have a great impact on reducing streamflow depletion as transmissivity increases.
- Numerical models simulate slightly less streamflow depletion as compared to the Jenkins (1968) model. Numerical models are approximations and the discrepancy seen could be a function of the discretisation error associated with numerical models for both space (grid size) and time (time steps). Refining the numerical model around the borehole slightly changes the results of the model simulation. The analytical model assumes that the aquifer extends infinitely long from the stream, however, a numerical model is bounded by no-flow boundary conditions (north, east, and south), thus less water available from the aquifer, and thus technically more streamflow depletion would have been simulated.
- The numerical model results correlate better with the Hantush (1965) model, than the Jenkins (1968) model. The 3rd boundary condition thus better represents the leakance factor, than a constant head boundary condition represents a fully penetrating steam; as a result, the numerical model shows a

better comparison with the Hantush (1965) model that considers the leakance factor of the streambed material.

- The Hunt (2003) model simulates a slightly higher streamflow depletion at early times, after that they begin to diverge, with the numerical model simulating larger values of streamflow depletion than the analytical solution. The differences between the two models could be due to the effect of a bounded aquifer system within the numerical model that is not reflected in the Hunt (2003) model. The two-layer numerical model considers the vertical movement of groundwater (3D model), while the analytical solution only considers the horizontal (2D) flow of groundwater within the aquifer and the difference in parameters (river package within the numerical model is largely dependent on the length of the river within the cell of the model.
- Analytical solution fails to address more complex systems as it either overestimates or significantly underestimates streamflow depletion rates due to its limitation of the simplified assumptions. However, analytical solutions serve as conservative tools to be used by water managers in giving early recommendations and mitigation for potential streamflow depletion due to groundwater pumping.
- Analytical models are often suitable for simpler, well-defined hydrogeological problems where assumptions can be adequately met. They provide quick and straightforward solutions that can give initial insights into the system's behaviour. Numerical models, on the other hand, are essential for capturing complex geometries, heterogeneities, and boundary conditions that analytical solutions may not handle accurately.

The thorough examination of streamflow depletion using a variety of analytical and numerical models has revealed the crucial dynamics influencing how groundwater pumping affects streamflow. This study emphasises the complexity of aquifer-stream interactions and the shortcomings of oversimplified analytical models by analysing several variables, including distance from boreholes, aquifer characteristics, and model complexity. The identification of time lags, the impact of geological variations, and the impact of model assumptions outline an essential future course for the management of water resources. These findings bridge a significant gap, providing water managers with valuable insights to navigate the complex balance between groundwater abstractions and maintaining streamflow. With the delineation of limitations in analytical models and the advantages offered by numerical simulations, this research serves as a beacon, advocating for a more holistic approach in coming up with sustainable strategies for mitigating streamflow depletion, and ensuring the responsible usage of groundwater resources for future generations.

Based on the findings, the following recommendations are given:

- Based on the parameters varying and sensitivity analysis, water managers should take into consideration parameters that greatly influence streamflow depletion (i.e. groundwater abstraction, distance separation). Therefore, the management and recommendation (safe yield) must be taken into consideration for water managers. Groundwater abstraction rates should be monitored, and data collection should be accurately acquired to give out the best recommendation as this can greatly impact the surface water. In addition, borehole allocation near the stream must also be taken into account as distance separation greatly impacts streamflow.
- Analytical solutions as screening tools can address how distance proximity could potentially affect streamflow depletion, therefore buffer zones need to be properly set based on site-specific circumstances, especially within a South African context.
- Newly developed analytical models (improved stream-aquifer conceptualisation) should also be looked at in the quantification of streamflow depletion and their strength in addressing complex systems.
- Based on the comparison of analytical solutions with developed numerical model, analytical solutions provide simple tools for the assessment of streamflow depletion and are appropriate for general purposes where the quick assessment of stream depletion is useful such as screening tools for water managers to use as supporting document for water use licensing application, this document provided broad capabilities for analytical model applications for varying aquifer conditions.
- Analytical solutions prove to overestimate and/or underestimate streamflow depletion when compared to complex settings. However, they prove to be conservative and can greatly assist water managers in formulating early mitigation measure approaches.
- The numerical model is potentially a comprehensive management tool for informing the development of sustainable groundwater allocation due to its strength to address complex settings and limitation assumptions of analytical solutions. Considerable work has gone into developing current models and testing them as groundwater management tools. However, more resources would be required to construct and test a sophisticated groundwater model for use in groundwater management.

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