ESTIMATION OF THE PRELIMINARY GROUNDWATER RESERVE USING NUMERICAL MODELS

BY

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DECLARATION

I declare that the dissertation hereby submitted by me for the Master of Science degree at the University of the Free State is my own independent work and has not previously been submitted by me at another university/faculty. I furthermore cede copyright of the dissertation in favour of the University of the Orange Free State.

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1 Introduction

1.1 Importance of Groundwater in South Africa

Water, particularly fresh and potable water, is indispensable for human's activities on earth. Approximately 75% of the total volume of fresh water on earth is frozen in glaciers, while 0.33% is held in rivers and lakes. The remaining 24.7% is groundwater (Chorley, 1969; Zumberge and Nelson, 1984). Since the water in glaciers is not available for general consumption, groundwater forms the largest source of fresh water available to human.

Although groundwater contributes only about 15% to the national water budget and the surface water is still the most important source of drinking water, the supply potential of groundwater should not be ignored. A rough estimation using the Electronic Guide (DWAF, 1999) shows that over 80% of the South African aquifers have a supply potential of over 22800 liters/km^2 per day. About 60% of the land even has a supply potential of over 37200 liters/km^2. Considering the current population density (36 persons/km^2), the supply potential of groundwater is on average approximately 630 liters/person per day in over 80% of the country.

Groundwater is the sole or main source of potable water for some urban and many rural communities in SA. The fact that more than 280 cities or towns, five million cattle, 26 million small stock and 24% of the irrigated land are dependent on groundwater (Van Tonder, 1999), clearly explains the importance of the groundwater resource in SA.

1.2 The National Water Act and the Reserve

Many countries realize the importance of the water resources and try to protect scarce water resources through new techniques or in a lawful way. The National Water Act (Republic of South Africa, Act No. 36 of 1998, hereafter the Water Act) is introduced not only to regulate the use of water but also to protect the water resources and environment from pollution and overexploitation.

According to Water Act (Part 8), only the water, which is still available after the requirements of the Reserve, international obligations and corrective action have been met, can be allocated to users. "Reserve" means the quantity and quality of water required in order to:

(a) Satisfy basic human needs by securing a basic water supply for people who now or who will, in the reasonably near future, rely upon; take water from; or be supplied from the relevant water resource.
(b) Protect aquatic ecosystems in order to secure ecologically sustainable development and use of the relevant water resource.

As groundwater is a relevant part of the water resource, its use and allocation are subject to the Reserve. That is, before a responsible authority may authorize the use of water, the groundwater component of the Reserve (hereafter the Groundwater Reserve) must be determined. The Water Act (Section 16(2)) requires that the determination of the Reserve must be in accordance with a prescribed water resource classification system. As setting up such a classification system will take a considerably long time, the Water Act (Part 3) allows a preliminary determination of the Reserve before a final determination of the Reserve within a classification system can be made.

1.3 Preliminary Estimation of the Groundwater Reserve

The Water Research Commission of SA (WRC) initiated four research projects which started in January 1999 in order to address the issues related to the Reserve of the Water Act. A methodology of determining the Groundwater Reserve is being addressed by the research project titled "A Modeling Decision-Support System for the Groundwater Reserve". The present work is the first part of this project and focuses on developing a methodology for preliminary estimation of the Groundwater Reserve by using existing computer programs. The methodology consists of two main components:

- Estimation of the Groundwater Reserve by means of water balance calculations: The objective is to estimate the Groundwater Reserve by treating a catchment area as a water balance model (sometimes it is called a box model). Groundwater fluxes, recharge, abstraction and leakage to/from surface water bodies are balanced over the whole catchment. Due to the simplicity of the box model, evaluation of field data is only concerned with fluxes between the aquifer and adjacent aquifers and /or surface water bodies. This model gives an overview of the accessible water resources and an estimation of the Groundwater Reserve within the catchment.
- Estimation of the Groundwater Reserve by considering impacts due to abstraction: The objective is to predict the impacts due to the proposed use (abstraction) of groundwater. A numerical modelling technique is used at this stage because of its prediction ability and ability to tie together data and physical principles into a coherent and useful picture of a site. Furthermore, numerical models can give an allowed abstraction rate by comparing various constrains with the estimated local impacts. Therefore they are suitable for responsible water licensing authorities.

Another advantage of numerical models is their flexibility related to the availability of field data. A well-conceptualized numerical model (and in some cases, equivalent analytical solutions) can still produce meaningful conclusions, even if field data is scarce. A numerical model can "grow" with the field data. It means the more data is available the better the model will be. Despite the necessity of acquiring skills in the application of using numerical modeling techniques, they represent the most widely used method in today's world.

1.4 Arrangement of the Text

The present work consists of six chapters:

Chapter 1 describes the objectives of this work.

Chapter 2 briefly introduces some popular numerical groundwater models and the mathematical background of the three-dimensional groundwater flow model MODFLOW.

Chapter 3 describes the theoretical background of the water balance model and provides the working steps for determining the Groundwater Reserve using the water balance model.

Chapter 4 provides methods for estimating local impacts due to (proposed) groundwater abstraction. Especially, the SEAWATER program, which is used to calculate the steady-state salt-/fresh water interface, is developed.

Chapter 5 is a case study on the Pienaars River catchment.

Chapter 6 concludes the present work and provides suggestions on future researches and data gathering issues.

2 Numerical Models

2.1 Selection of a Numerical Model

Although the applications of numerical groundwater models to solve groundwater related problems are quite a new field in the industry, many popular models were already published in the 1980's. To date, many public domain codes are results of research works consulted by the U.S. Geological Survey, the U.S. EPA and other public sources. One of the most important criteria for selecting a model is that its source code must be available and can be modified and distributed freely. In order to select suitable models, we need to examine some of the most popular and widely used codes, which are briefly introduced in the following sections.

2.1.1 ASMWIN

ASMWIN (Aquifer Simulation Model for Windows) is a complete two-dimensional groundwater flow and transport model developed by Chiang et al. (1998). ASMWIN includes a professional graphical user-interface, a finite-difference flow model, a tool for the automatic calibration of flow models, a particle tracking model, a random walk transport model based on Ito-Fokker-Planck theory, a finite-difference transport model and several other useful modelling tools. The flow model simulates the effects of wells, rivers, drains, head-dependent boundaries, recharge and evapotranspiration. The particle-tracking model ASMPATH offers several velocity interpolation methods and uses Euler-Integration to compute flow paths and travel times.

2.1.2 MODFLOW

MODFLOW is a modular three-dimensional finite-difference groundwater flow model developed by U. S. Geological Survey. The applications of MODFLOW to the description and prediction of the behavior of groundwater systems have increased significantly over the last few years. There are two versions of MODFLOW. The first version is known as MODFLOW-88 (McDonald and Harbaugh, 1988) and the second version MODFLOW-96 (Harbaugh and McDonald, M.G., 1996a, 1996b). The "original" version of MODFLOW-88 (McDonald and Harbaugh, 1988) or MODFLOW-96 (Harbaugh and McDonald, 1996a, 1996b) simulate the effects of wells, rivers, drains, head-dependent boundaries, recharge and evapotranspiration. Since the publication of MODFLOW, numerous investigators have developed various codes for specific purposes. These codes are called *packages*, *models* or sometimes simply *programs*. *Packages* are integrated with MODFLOW, each package deals with a specific feature of the hydrologic system to be simulated, such as reservoir, wetland or lake. *Models* or *programs* can be stand-alone codes or can be integrated with MODFLOW. A stand-alone model or program communicates with MODFLOW through data files. The advective transport model PMPATH (Chiang and Kinzelbach, 1994, 1998), the solute transport model MT3D (Zheng, 1990), MT3DMS (Zheng and Wang, 1998) and the parameter estimation programs PEST (Doherty et al., 1994) and UCODE (Poeter and Hill, 1998) use this approach. The solute transport model MOC3D (Konikow et al., 1996) and the inverse model MODFLOWP (Hill, 1992) are integrated with MODFLOW. Both codes use MODFLOW as a function for calculating flow fields.

Because of rapid development of computer techniques, many professional graphical user interfaces are already available, for example Processing Modflow (Chiang and Kinzelbach, in press) or commercial software Visual MODFLOW (WHI) and Groundwater Modelling System (Brigham Young University).

2.1.3 HST3D

The heat- and solute-transport model HST3D (Kipp Jr., 1986, 1997) is a finite-difference model to simulate the groundwater flow and associated heat and solute transport based in saturated, three-dimensional flow system with variable density and viscosity. It couples the saturated groundwater flow equation, the heat-transport equation and the solute transport equation through the dependence of advective transport on interstitial fluid-velocity field, the dependence of the fluid viscosity on temperature and solute concentration, and the dependence of fluid density and on pressure, temperature, and solute concentration. The HST3D code is applicable to the study of waste injection into fresh or saline aquifers, contaminant plume movement, saltwater intrusion in coastal regions, brine disposal, freshwater storage in saline aquifers, heat storage in aquifers, liquid-phase geothermal system, and similar transport situation. The numerical solutions of HST3D are pressure, temperature, and solute concentration at each finite difference cell.

A text-based preprocessor for HST3D has been developed by Chiang (1990).

2.1.4 SUTRA

SUTRA (Voss, 1984) is a two-dimensional finite element model, which simulates saturated and unsaturated, fluid-density-dependent groundwater flow with energy transport and chemically reactive single-species solute transport. SUTRA flow simulation may be used for areal and cross-sectional modeling of saturated groundwater flow systems, and for cross-sectional modelling of unsaturated zone flow. Solute transport simulation with SUTRA may be used for modelling of variable density leachate movement, and for cross-sectional modelling of salt-water intrusion in aquifers. Energy transport simulation may be used to model subsurface heat conduction, geothermal reservoirs or thermal pollution of aquifers. Graphical user-interfaces for preparing SUTRA input data are available, for example Chiang (1989) or SUTRA-GUI (Voss, 1997). A model for inverse modelling with SUTRA has been released by Piggott and Bobba (1993). This code performs automatic model calibration by using the measured hydraulic heads or concentration values. A three-dimensional version of SUTRA is being developed in the U.S.

2.1.5 Conclusion

In recent years, many researchers have concentrated on the development of easy-to-use graphical user interface (GUI's), three-dimensional visualization systems, reactive (multi-species) transport models, inverse models for parameter estimation, flow and transport models for fractured aquifers or even methodology for uncertainty determination by using the stochastic modelling approach.

As many of these works are developing computer codes for use with MODFLOW or MODFLOW related software, MODFLOW appears to be the most suitable code for the present work, the following section describes MODFLOW in details.

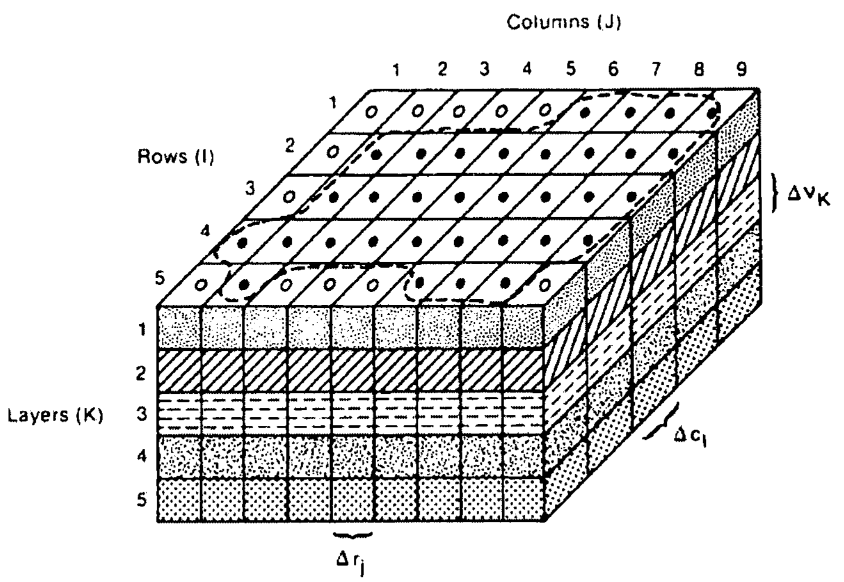
2.2 Introduction to MODFLOW

2.2.1 Basic Theory of MODFLOW

The MODFLOW model uses the finite-difference method to describe the groundwater flow. In MODFLOW, the simulation time is divided into stress periods – i.e. time intervals during which all external excitations or stresses are constant. The stress periods are in turn divided into time steps. For the spatial discretization, an aquifer system is replaced by a discretized domain consisting of an array of nodes and associated finite difference blocks or cells (Figure 2.1) in which spatial dependent variables are assumed uniform. The nodal grid forms the framework of a numerical model. Hydrostratigraphic units can be represented by one or more model layers. The thickness of model cells and the widths of columns and rows may be variable. MODFLOW uses an index notation [i, j, k] for locating the cells. For example, the cell located in the sixth row, second column, and the first layer is denoted by [6, 2, 1]. The partial differential equation describing the three-dimensional groundwater flow is replaced by a finite number of algebraic equations, written in terms of the values of the dependent variables at each cell. The algebraic equation of a cell can be obtained by applying the continuity equation, that is the sum of all flows into and out of the cell must be equal to the rate of exchange in storage within the cell. Under the assumption that the density of groundwater is constant, the continuity equation expressing the balance of flow for the cell [i, j, k] (Figure 2.2) is

$$q_{i,j-1/2,k} + q_{i,j+1/2,k} + q_{i-1/2,j,k} + q_{i+1/2,j,k} + q_{i,j,k-1/2} + q_{i,j,k+1/2} = S_{i,j,k} \Delta V_{i,j,k} \Delta h_{i,j,k} \Delta t \qquad (2.1)$$

where $q_{i,j-1/2,k}$ $[L^3T^{-1}]$ is the water exchange between cell [i, j, k] and cell [i, j-1, k]; $SS_{i,j,k}$ is the specific storage in cell [i, j, k]; $\Delta V_{i,j,k}$ $[L^3]$ is the volume of cell [i, j, k]; $\Delta h_{i,j,k}$ [L] is the change of the hydraulic head over a time interval of Δt [T].



Explanation

-- Aquifer Boundary

• Active Cell

0 Inactive Cell

r_J Dimension of Cell Along the Row Direction. Subscript (J) Indicates the Number of the Column

c_I Dimension of Cell Along the Column Direction. Subscript (I) Indicates the Number of the Row

v_K Dimension of the Cell Along the Vertical Direction. Subscript (K) Indicates the Number of the Layer

Figure 2.1 A discretized hypothetical aquifer system
(After McDonald and Harbaugh, 1988)

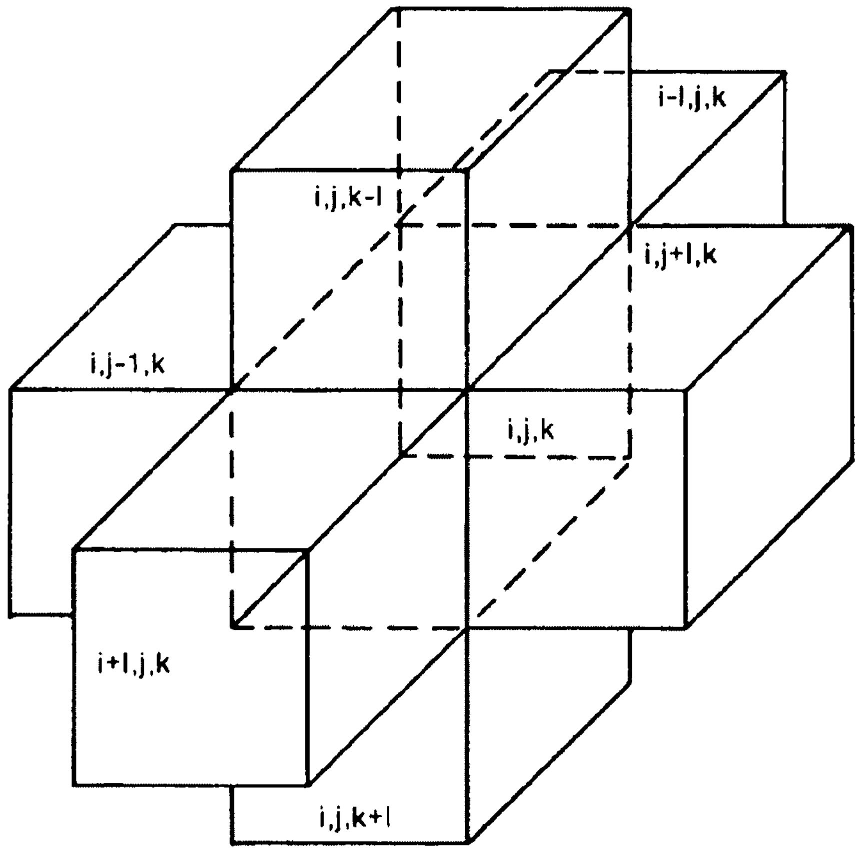


Figure 2.2 Cell [i, j, k] and its six adjacent cells

(After McDonald and Harbaugh, 1988)

If sink or source terms exist, e.g. pumping well, river leakage or recharge, etc., the sum of these terms $Q_{i,j,k}$ needs to be added into the equations (2.1). A negative value of $Q_{i,j,k}$ means that the cell loses water. The Equation (2.1) becomes

$$q_{i,j-1/2,k} + q_{i,j+1/2,k} + q_{i-1/2,j,k} + q_{i+1/2,j,k} + q_{i,j,k-1/2} + q_{i,j,k+1/2} + Q_{i,j,k} =$$
$$S_{i,j,k} \Delta V_{i,j,k} \Delta h_{i,j,k} \Delta t \qquad (2.2)$$

Using the Darcy's law, the equation (2.2) can be rewritten as:

$$\begin{aligned} & CR_{i,j-1/2,k} \cdot (h^m_{i,j-1,k} - h^m_{i,j,k}) + CR_{i,j+1/2,k} \cdot (h^m_{i,j+1,k} - h^m_{i,j,k}) + \\ & CC_{i-1/2,j,k} \cdot (h^m_{i-1,j,k} - h^m_{i,j,k}) + CC_{i+1/2,j,k} \cdot (h^m_{i+1,j,k} - h^m_{i,j,k}) + \\ & CV_{i,j,k-1/2} \cdot (h^m_{i,j,k-1} - h^m_{i,j,k}) + CV_{i,j,k+1/2} \cdot (h^m_{i,j,k+1} - h^m_{i,j,k}) + Q_{i,j,k} = \\ & S_{i,j,k} \cdot (\Delta r_j \cdot \Delta c_j \cdot \Delta v_k) \frac{(h^m_{i,j,k} - h^{m-1}_{i,j,k})}{t_m - t_{m-1}} \end{aligned} \qquad (2.3)$$

The values of h^m [L] are the hydraulic heads at the end of the m-th time step (time t_m). The definitions of the variables *CR*, *CC* and *CV* are given below:

- $CR_{i,\, j-1/2,\, k}$ [L^2T^{-1}] is the conductance in row i and layer k between cells [i, j-1, k] and [i, j, k].

  $CR_{i,j-1/2,k} = 2 \cdot KR_{i,j-1/2,k} \cdot \Delta c_i \cdot \Delta v_K / \Delta r_{j-1}$

  where

  $KR_{i,j-1/2,k}$ [LT^{-1}] is the hydraulic conductivity along the row between cells [i, j, k] and [i, j-1, k] and c_i, v_k and r_j are the dimensions of the cell [i, j, k] along the column, row and vertical directions

- $CC_{i-1/2,\, j,\, k}$ [L^2T^{-1}] is the hydraulic conductance in column j and layer k between cells [i-1, j, k] and [i, j, k].

  $CC_{i-1/2,j,k} = 2 \cdot KC_{i-1/2,j,k} \cdot \Delta r_j \cdot \Delta v_K / \Delta c_{i-1}$

  where

  $KC_{i-1/2,j,k}$ [LT^{-1}] is the hydraulic conductivity along the column between cells [i, j, k] and [i-1, j, k];

- $CV_{i,j,k-1/2}$,[L^2T^{-1}] is the conductance in row i and column j between cells [i, j, k-1] and [i, j, k]

  $$CV_{i,j,k-1/2} = 2 \cdot KV_{i,j,k-1/2} \cdot \Delta r_j \cdot \Delta c_i / \Delta v_{k-1}$$

  where

  $KV_{i,j,k-1}$ [LT^{-1}] is the hydraulic conductivity along the vertical direction between cells [i, j, k] and [i, j, k-1].

To calculate heads in each cell in the finite-difference grid, MODFLOW prepares one finite difference equation for each cell, expressing the relationship between the head at a cell and the heads at each of the six adjacent cells at the end of a time step (time t_m). Because each equation may involve up to seven unknown values of head, and because the set of unknown head values changes from one equation to the next through the grid, the equations for the entire grid must be solved simultaneously at each time step.

2.2.2 Packages for MODFLOW

To date, more than 20 packages for specific features have been integrated into MODFLOW. Some of them deal with the solution of the system of equations, e.g. PCG2 (Hill, 1990a, 1990b) or DE45 (Harbaugh, 1995). Some other packages are designed to simulate geotechnical features, for example the Interbed-Storage package (Leake and Prudic, 1991) simulates subsidence of land surface due to groundwater abstraction and the Horizontal-Flow Barrier package (Hsieh and Freckleton, 1993) simulates thin, vertical low permeability geological features such as cut-off walls. Most of the integrated packages, however, are designed for simulating the interaction between surface water and groundwater. Chiang and Kinzelbach (1991-1999) provide an overview of these packages and documentations on a CD-ROM. The following sections describe the packages, which deal with groundwater and surface water interaction. These packages are Drain, River, General-Head Boundary, Streamflow-Routing, Lake, Reservoir and Wetland.

2.2.2.1 Drain Package

The Drain Package (McDonald and Harbaugh, 1988) is designed to simulate the process of removing water from the aquifer at a rate proportional to the difference between the head in the aquifer and the elevation of a drain. When the hydraulic head is greater than the drain elevation, water flows into the drain and is removed from the groundwater system. Discharge to the drain is zero if the hydraulic head is lower than or equal to the median drain elevation. Recharge from the drain is always zero, regardless of the hydraulic head in the aquifer. For each model cell containing a drain, discharge rate to the drain (*Qd*) is calculated by

$$Qd = CD \cdot (h - d) \qquad h > d \qquad (2.4)$$

$$Qd = 0 \qquad h \leq d \qquad (2.5)$$

where h [L] is the hydraulic head in the cell containing a drain, d [L] is the median drain elevation and CD [L^2T^{-1}] is the equivalent conductance describing all of the head loss between the drain and the region of a cell. Similar to the value of *CRIV* in the River Package, the value of *CD* depends on the characteristics of the convergence flow pattern toward the drain and drain characteristics, it is usually unknown, and must be adjusted during model calibration.

2.2.2.2 River Package

River Package (McDonald and Harbaugh, 1988) is used to simulate the effect of flow between groundwater system and a surface-water feature, like river or open channel, etc. In the river package, a river or stream is divided into reaches, and each reach corresponds to a single cell of a MODFLOW model. The flow to or from a surface water body is calculated for each cell, and added into the term Q in equation (2.3).

The River package offers a simplified estimation of the flow between surface water and groundwater under the following assumptions:

- A discrete low-permeability riverbed layer is present

- Measurable head losses between the river and the aquifer are limited to those across the riverbed itself.
- The underlying model cell remains fully saturated.
- The seepage to or from a river is uniformly distributed over the cell area.

Based on the above-mentioned assumptions and applying Darcy's Law, the rate of leakage from a river to a groundwater system is calculated by:

$$Q_{riv} = CRIV(H_{riv} - h) \qquad h > RBOT \qquad (2.6)$$

$$Q_{riv} = CRIV(H_{riv} - RBOT) \qquad h \leq RBOT \qquad (2.7)$$

where:

Q_{riv} [L^3T^{-1}] is the rate of leakage from a river to a groundwater system,

CRIV [L^2T^{-1}] is the hydraulic conductance of the riverbed,

H_{riv} [L] is the stage of the river,

h [L] is the hydraulic head in the cell underlying the river reach,

RBOT [L] is the elevation of the riverbed bottom.

In the case where h is greater than H_{riv}, Q_{riv} is negative. This means that water flows from the aquifer into the river and is removed from the groundwater system. The River Package can be used to replace the Drain Package by setting H_{riv} equal to *RBOT*.

The value of *CRIV* is usually estimated by

$$CRIV = (K \cdot L \cdot W)/M \qquad (2.8)$$

Where *K* [LT^{-1}] is the hydraulic conductivity of the riverbed material, *L* [L] is the length of the river reach within a cell, *W* [L] is the width of the river and *M* [L] is the thickness of the riverbed. In most cases, however, the value of *CRIV* cannot be determined in this way, because the assumptions about the river package are often not satisfied. Rather, the *CRIV* values represent equivalent conductance describing the resistance to the water flow between a river and a groundwater system. These values must often be determined during the course of a model calibration.

2.2.2.3 General-Head Boundary Package

The General-Head Boundary Package (McDonald and Harbaugh, 1988) is used to simulate head-dependent flow boundaries (Cauchy boundary conditions) and is mathematically similar to those of the River or Drain packages. For each model cell with the general-head boundary condition, flow Qb [L^3T^{-1}] from or to an external source (surface water body) is assumed to be in proportion to the head difference between groundwater and surface water:

$$Qb = C_b \cdot (h_b - h) \qquad (2.9)$$

where h_b [L] is the hydraulic head (stage) of the external source, h [L] is the hydraulic head in the aquifer and C_b is the hydraulic conductance of the general-head boundary. If a very large value of C_b is used, a general-boundary head cell is equivalent to a fixed-head cell.

2.2.2.4 Streamflow-Routing Package

The Streamflow-Routing Package (Prudic, 1989) is designed to account for flow in streams and simulate the water interaction between streams and groundwater. In this package, streams are divided into segments and reaches. A segment consists of a group of reaches connected in downstream order (Figure 2.3). Each reach corresponds to an individual cell in the finite-difference grid. Streamflow is accounted for by specifying flow for the first reach in each segment, and then computing streamflow to adjacent downstream reaches as inflow in the upstream reach plus or minus leakage from or to the aquifer in the upstream reach. The accounting scheme used in this package assumes that streamflow entering the modelled reach is instantly available to downstream reaches. This assumption is generally reasonable because of the relatively slow rates of groundwater flow.

The streamflow into a segment that is from tributary streams is computed by adding the outflows from the last reach in each of the specified tributary segments. If a segment is a diversion, then the specified flow into the first reach of the segment is subtracted from flow in the main stream. However, if the specified flow of the diversion is greater than the flow out of the segment, then no flow is diverted from that segment.

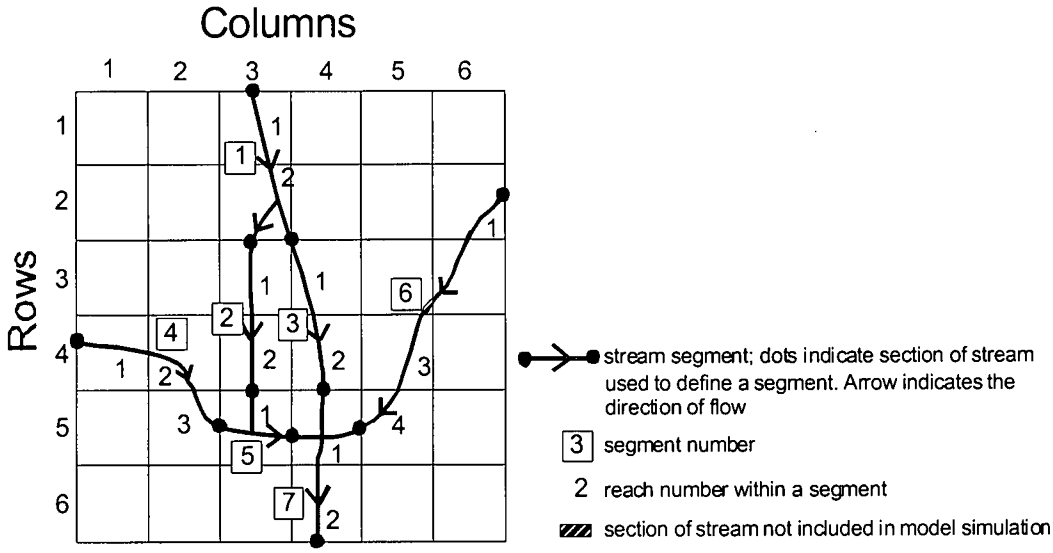


Figure 2.3 Numbering system of the Streamflow-Routing Package (After Prudic, 1989)

Although the Streamflow-Routing Package allows two or more diversions from main streams, the flow rate to the diversions need to be known. This package does not track the amount of flow in the rivers, does not include a time function for streamflow, nor does it permit rivers to go dry during a given period of simulation. Therefore, the Streamflow-Routing Package is not a true surface water model, but rather an accounting program that tracks the flow in one or more streams which interact with groundwater.

Similar to the River package, the leakage (Q_{str}) to or from the aquifer through the streambed for each model cell containing a stream reach can be computed by:

$$Q_{str} = CSTR \cdot (H_{str} - h) \qquad h > SBOT \qquad (2.10)$$

$$Q_{str} = CSTR \cdot (H_{str} - SBOT) \qquad h \leq SBOT \qquad (2.11)$$

where H_{str} [L] is the head in a stream reach, h [L] is the head in the model cell, $SBOT$ [L] is the elevation of the bottom of the reach, $CSTR$ [L^2T^{-1}] is hydraulic conductance of the streambed. $CSTR$ can be calculated in the same way as the River Package.

2.2.2.5 Reservoir Package

The Reservoir Package (Fenske et al., 1996) simulates the water interaction between reservoir and aquifer. The Reservoir package is ideally suited for cases where leakage from or to reservoirs may be a significant component of flow in a groundwater system. However, this package assumes that the reservoir stages over time are known. If it is unknown, then a more complex conceptualization, e.g. the Lake Package below, would be needed in which the reservoir stage would be computed as part of the simulation rather than having the stage specified as model input.

For each model cell within a reservoir, water flow Q_{res} [L^3T^{-1}] between reservoir and underlying groundwater system is computed in a manner identical to the River package (see above):

$$Q_{res} = CRES(H_{res} - h) \qquad h > RESBOT \qquad (2.12)$$

$$Q_{res} = CRES(H_{res} - RESBOT) \qquad h \leq RESBOT \qquad (2.13)$$

where H_{res} [L] is the reservoir stage and h [L] is the hydraulic head. $RESBOT$ [L] is the elevation of the base of the reservoir-bed sediment. $CRES$ is the hydraulic conductance of the reservoir bed and can be calculated by

$$CRES = (HC_{res} \cdot DELR(j) \cdot DELC(i)) / Rb \qquad (2.14)$$

where HC_{res} [LT^{-1}] is the vertical hydraulic conductivity of the reservoir-bed material, $DELC(i)$ and $DELR(j)$ are the width of the cell [i, j, k] along the column and row directions. Rb [L] is the thickness of the reservoir-bed.

Three options are available for simulating leakage between a reservoir and the underlying groundwater system. The first option simulates leakage only to layer 1; the second option simulates leakage to the uppermost active cell; and the third option simulates leakage to a specified layer for each active reservoir cell. Inherent in the simulation of reservoirs is that the reservoir only partially penetrates an active model cell. If the reservoir fully penetrates a cell, the reservoir leakage will be simulated in a lower cell. Thus, water exchange between the groundwater system and the reservoir takes place across the bottom of the reservoir and the top of the model cells.

2.2.2.6 Lake Package

The Lake Package (Version 1: Cheng and Anderson, 1993; Version 2: Council, 1998) is designed to simulate lake effects to the groundwater system. While the Reservoir Package requires input of measured stages, the Lake Package calculates groundwater fluxes and lake level fluctuations over time by the simulating dynamic water exchange between lake and groundwater system.

For each model cell within a lake, the leakage (Q_{lake}) to or from the aquifer through the lakebed is computed by:

$$Q_{lake} = CLAKE \cdot (H_{lake} - h) \qquad h > LAKBOT \qquad (2.15)$$

$$Q_{lake} = CLAKE \cdot (H_{lake} - LAKBOT) \qquad h \leq LAKBOT \qquad (2.16)$$

where H_{lake} [L] is lake stage, h [L] is the hydraulic head in the model cell; $LAKBOT$ is the bottom elevation of the lakebed; $CLAKE$ [L^2T^{-1}] is the hydraulic conductance of lakebed, its definition is the same as the hydraulic conductance of a riverbed.

The Lake Package includes all features of the Reservoir Package and handles lake-stream interaction with the streamflow-routing package. The lake stage is calculated as a transient response to evaporation, precipitation, surface runoff, streamflow and groundwater flux. The first version of the Lake Package uses Manning's equation to calculate the outflow from the lake into a stream. The second version requires the input of three empirical parameters to describe the outflow from a reservoir. The capability of simulating dynamic flow conditions in stream/lake/reservoir systems is still limited by the Streamflow-Routing package. To solve this problem, a surface water flow model needs to be coupled with MODFLOW.

2.2.2.7 Wetland Package

The Wetland Package (Restrepo et al., 1998) considers surface flow, wetting and drying, evapotranspiration and the vertical and horizontal flux of the wetland-aquifer interaction. In this package, surface water flow in wetlands is represented by two components: (a) sheet flows through dense vegetation and (b) water movement through wetland slough channel. The overall water movement in wetlands is often dominated by slough channels, because they have less vegetation and the flow velocity in the slough channel is much higher than the velocity of sheet flow.

The Wetland Package enables the top layer of a grid system to contain the above-mentioned two components. To simulate the sheet flow, horizontal hydraulic conductivities in the groundwater flow model are modified by considering the Manning's roughness coefficient, vegetation density, water depth as well as thickness and hydraulic conductivity of the muck.

The interaction between aquifer and wetland is calculated in the same way as the River Package. However, the hydraulic conductance between the wetland and groundwater system is determined by the product of the cell area and the harmonic mean of the vertical hydraulic conductivities of the muck and aquifer.

3 Preliminary Estimation of the Groundwater Reserve and the Allocatable Groundwater Resource

As described in section 1.3, the procedure to estimate the Groundwater Reserve is divided into two steps. The first step is to estimate the Groundwater Reserve by means of the regional groundwater balance. The second step is to predict and evaluate the environmental impacts caused by proposed groundwater abstractions. The former is discussed in this chapter.

3.1 Water Balance Model and the Groundwater Reserve

The water balance model, also referred to as water budget, mass balance or black box model, was developed in the 1940's by Thornthwaite (1948) and revised by Thornthwaite and Mather (1955). A water balance model is a quantitative evaluation of water gained or lost from the groundwater system under investigation during a specific period of time. It is assumed that the change of groundwater stored in the system relates directly to the change of groundwater level within the system, and it can be expressed as:

$$\textit{Change in storage} = \textit{Inflow} - \textit{Outflow} \qquad (3.1)$$

For a groundwater system shown in Figure 3.1, equation 3.1 may be written in the form of water balance terms:

$$\frac{\Delta Q_S}{\Delta t} = \pm Q_N \pm Q_R \pm Q_B \pm Q_L \pm Q_W \qquad (3.2)$$

where

ΔQ_S [L^3]	=	change in storage within the time increment Δt
$\pm Q_N$ [L^3/T]	=	natural groundwater recharge or discharge
$\pm Q_R$ [L^3/T]	=	leakage rate from/to surface water bodies

$\pm Q_B$ [L^3/T] = subsurface flow rate from/to horizontal adjacent aquifers

$\pm Q_L$ [L^3/T] = leakage rate from/to vertical adjacent aquifers

$\pm Q_W$ [L^3/T] = local infiltration or abstraction rate, e.g. wells

Δt [T] = time increment

Under the steady-state groundwater flow condition, the change in storage is zero, and the equation can be rewritten as:

$$\pm Q_N \pm Q_R \pm Q_B \pm Q_L \pm Q_W = 0 \qquad (3.3)$$

Estimation of the Groundwater Reserve discussed in the following section is based on the assumption that the groundwater flow system is under the steady-state condition.

3.2 Calculation of the Water Balance Terms

3.2.1 Estimation of Natural Groundwater Recharge and Discharge

Groundwater recharge due to precipitation or discharge due to evaporation and transpiration is generally the most prominent element in the water balance calculation. They are influenced by many factors, such as precipitation intensity, the depth of groundwater level, soil type, climate condition, vegetation, slope of ground surface, etc. Because of the unidentified nature of these factors, an accurate estimation of natural groundwater recharge and discharge is very difficult.

The EXCEL-Spreadsheet (Van Tonder and Xu, 1999) is used to determine the net groundwater recharge. RECHARGE features a user-friendly interface. It is capable of using different methods including the Chloride method, Saturated Volume Fluctuation method (Van Tonder, 1989), Cumulative Rainfall Departure method (Wenzel, 1936; Temperley, 1980), isotopes method, and a series of qualified guess methods. Bredenkamp et al. (1995) provide a detailed discussion on these methods. The data required by each method are listed in Table 3.1.

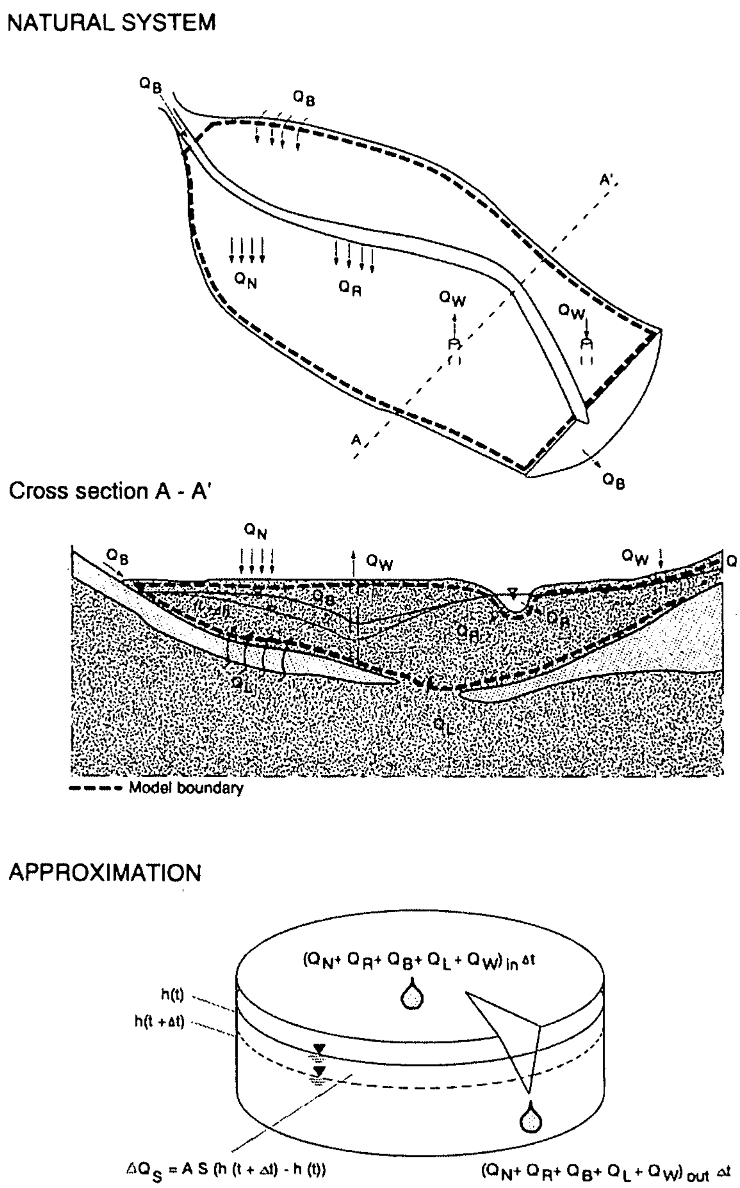


Figure 3.1 The water balance model for a groundwater flow system (After Spitz and Moreno, 1996)

Table 3.1 Data required by the RECHARGE program

Methods		Data
Chloride method		• Average chloride concentration in groundwater and rainfall • Average annual rainfall
Saturated Volume Fluctuation (SVF) method		• Monthly abstraction from aquifer • Lateral inflow and outflow • Monthly water level
Cumulative Rainfall Departure (CRD) method		• Lateral inflow and outflow • Monthly abstraction from aquifer • Monthly water level • Monthly precipitation
Qualified Guess	Soil and vegetation information	• Percentage coverage of area with soil material (% clay content) • Percentage coverage of area of woods/trees, grass lands and bare soil
	Geology	Type of soil covering the ground surface
	Vegter	Vegter's map (WRC and DWAF, 1995)
	Acru	ACRU map (Roland Schulze) (WRC and DWAF, 1995)
	Harvest potential	Harvest Potential Map (DWAF, 1996)
	Base flow	Groundwater Component of River Base Flow Map (WRC and DWAF, 1995)
Isotopes (Residence time)		• $^{2}\delta$ values in (un)saturated zone • ^{18}O values

3.2.2 Estimation of the Leakage Rate from/to Surface Water Bodies

Surface water bodies (rivers, lakes, channels, streams, …etc.) recharge groundwater or discharge groundwater. The exchange rate of water is usually controlled by the difference in the hydraulic heads (water levels) and the resistance of the media between the groundwater and surface water bodies. According to the water levels, the surface water body can be classified into three groups (Figure 3.2):

- Influent: The groundwater level is lower than the surface water
- Effluent: The groundwater level is higher than surface water level, groundwater always recharges surface water.
- Intermittent: The groundwater level is higher than the bed of the surface water body, but depending on the elevation of the water level, groundwater may recharge surface water body or surface water recharges groundwater.

The leakage principle discussed in section 2.2.2.2 is normally used to calculate the leakage rate from or to surface water bodies. This approach is, however, not always practical for the preliminary estimation of the Groundwater Reserve, because all factors in those equations of section 2.2.2.2 vary with time and position and their use often requires considerable effort during the parameter estimation. Another approach, which uses the numerical model PMWIN (Chiang and Kinzelbach, in press) as a computational tool, is suggested below:

1. Discretize the study area:
   The study area is subdivided into finite-difference blocks, the block sizes depend on the size of the study area and the desired resolution.
2. Assign the measured groundwater level into each block:
   Generally, the observed water levels are only available at some discrete points, and an interpolation method must be used to obtain the water levels for all finite-difference blocks within the study area. If the groundwater levels are well correlated with the topography, the interpolation program TRIPOL (Van Tonder et al., 1996) is a better choice. Otherwise, interpolation methods such as Inverse Distance (Shepard, 1968) or Kriging can be used.

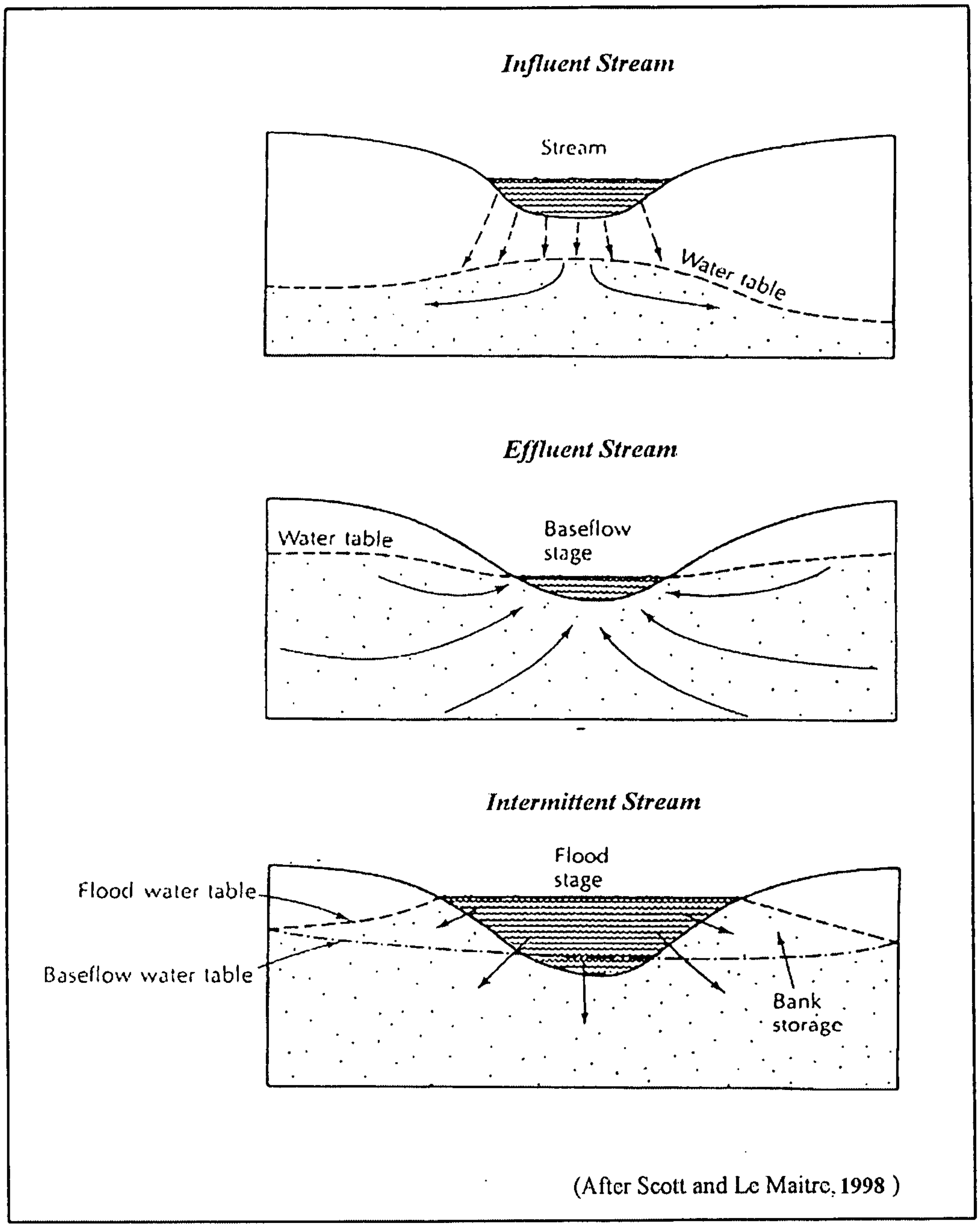


Figure 3.2 An illustration of the different relationship between river and groundwater

3. Assign transmissivity values or hydraulic conductivity values and the aquifer thickness to each block.
4. Assign a high storage value to each block:
   When a high storage value, say 1.0, is used, the water level of each block will only change very slightly during a flow "simulation". This ensures that the measured water level will be used for the flux calculation (see point 6 below).
5. Perform the transient flow simulation for one second:
   PMWIN calculates and saves the subsurface fluxes between each finite-difference block.
6. Calculate the flux interaction between groundwater and surface water
   The groundwater in- and outflows from or to surface water bodies are calculated by using the water budget calculator of PMWIN. This calculator requires that different subregions be assigned to the surface water body and the rest groundwater model domain.

Strictly speaking, this approach is not a usual way to build and run a groundwater flow model. Here, the computer program is used as a tool to calculate groundwater flow for each block under the given transmissivity and groundwater level values. The governing equation behind this approach is Darcy's law. The short 'simulation' duration of one second and high storage value is used so that the given groundwater level remains practically unchanged when groundwater flow is calculated.

3.2.3 Estimation of the Leakage Rate from/to Vertically Adjacent Aquifers

If an aquifer system includes several aquifers in the vertical direction, the water level difference between neighbour aquifers will force the exchange of water, which can be estimated by:

$$q_L = C(h_1 - h_2) \qquad (3.4)$$

where

q_L [L/T] = leakage flux between two aquifers per unit square meter

h_1 [L] = head in aquifer 1

h_2 [L] = head in aquifer 2

C $[T^{-1}]$ = resistance factor between aquifer 1 and aquifer 2

The resistance factor C between two aquifers is normally calculated by:

$$C = \frac{2K_1 \cdot K_2}{d_1 \cdot K_2 + d_2 \cdot K_1} \qquad (3.5)$$

where

K_1 [L/T] = vertical hydraulic conductivity of aquifer 1

K_2 [L/T] = vertical hydraulic conductivity of aquifer 2

d_1 [L] = thickness of aquifer 1

d_2 [L] = thickness of aquifer 1

Similar to section 3.2.2, the computer program PMWIN can be used to calculate the leakage rate between two aquifers:

1. Discretize the study area
2. Assign top and bottom elevations of aquifers to each block
3. Assign the measured groundwater level to each block
4. Assign transmissivity or horizontal hydraulic conductivity values to each block
5. Assign vertical hydraulic conductivity values to each block
6. Assign a high storage values to each block
7. Perform a transient flow simulation for one second
8. Calculate the flux between two aquifers
   Assign different subregions (say 1 and 2) to different aquifers, and use the water budget calculator to compute the flux.

3.2.4 Estimation of the Leakage Rate from/to Horizontally Adjacent Aquifers

The calculation of over boundary flow rates can be omitted if natural no-flow boundaries are used as the boundaries of the study area, for example, groundwater divides, impermeable fault or rock. If it is not possible, the over boundary flow can be estimated by using the computer program PMWIN. The steps 1 to 5 are same as those in section 3.2.2. The last step is:

6. Calculation of over boundaries flow:

   A strip along the boundary of the study area is assigned to a subregion, and the water budget calculator is used to compute the in- and outflows to the subregion. The value of the inflow is equivalent to the groundwater outflow over the boundary (water leaves the system). The value of outflow is equivalent to the groundwater inflow over the boundary (water enters the system).

In addition, chemical and isotopic analyses of groundwater can also be used to evaluate the inflow rate (Desaullniers et al., 1981), and the water balance calculation of neighbouring areas can also provide useful information.

3.2.5 Estimation of the Existing Groundwater Abstraction

Existing groundwater abstractions can only be obtained through investigation of the abstraction boreholes within the study area. Existing databases, e.g. the National Groundwater Data Base (NGDB), should be used. One should be aware of the possibility that many boreholes are not registered in the database. If a useful database does not exist, information, such as land use maps (for estimating irrigation) and population (for estimating drinking and industrial uses) can be used to estimate the existing abstraction rate.

3.2.6 The Basic Human Needs Reserve and the Ecological Reserve

The basic human needs reserve is obtained by the product of the groundwater-dependent population and 25 l/day per person (Water Service Act: Act No. 108 of 1997). The change of groundwater-dependent population in future must be considered.

The Ecological Reserve is defined in order to protect aquatic ecosystems. One of the easiest ways (but not always the best) to protect aquatic ecosystems is to prevent these systems from changing. This means that the estimated groundwater discharge to aquatic ecosystems or surface water bodies (i.e. rivers, wetland, springs) should not be allocated for further uses. It is therefore suggested to define this groundwater discharge as the required Ecological Reserve. Doing so, we obtain a conservative, but safe, solution.

3.3 Determine the Groundwater Reserve and the Allocatable Groundwater Resource

All terms of the water balance model and the calculation scheme are summarized below:

Inflow to the aquifer (groundwater):

Net groundwater recharge (Mm^3/a)

Leakage rate from surface water (Mm^3/a)

Leakage rate from horizontally adjacent aquifers (Mm^3/a)

Leakage rate from vertically adjacent aquifers (Mm^3/a)

Total inflow (Mm^3/a)

Outflow from the aquifer (groundwater):

Existing abstraction (Mm^3/a)

Leakage rate to surface water (Mm^3/a)

Leakage rate to horizontally adjacent aquifers (Mm^3/a)

Leakage rate to vertically adjacent aquifers (Mm^3/a)

Total outflow (Mm^3/a)

Groundwater Reserve:

Basic human needs (BHN) (Mm^3/a)

Ecological reserve (Mm^3/a)

Allocatable Groundwater Resource:

Allocatable Groundwater Resource= Total Inflow - Existing Abstraction (excluding BHN) - BHN - Ecological Reserve

3.4 Summary of the Data Sources

The data and sources used in this chapter are summarized in Table 3.2. Additional data, which can help to understand geological conditions of a study area, are listed in Table 3.3.

Table 3.2 Summary of the data sources

Data Type	**Sources**
Geology • Borehole logs	National Groundwater Data Base (NGDB), Council for Geoscience
Geo-hydrological information • Borehole location and observed groundwater level • Aquifer property • Existing abstraction • Groundwater Recharge (Table 3.1)	NGDB, Field Data, National Groundwater Maps (Vegter, 1995)
Ecological information • Ecological constrains	Ecologists /Surface water hydrologists (DWAF)
Other data • Population depending on groundwater	Surveyor General, Field data

Table 3.3 Additional data sources

Data Type	**Sources**
Geology • Lithological characteristics • Structures	National Groundwater Data Base (NGDB), Field data, Council for Geoscience, National groundwater maps (Vegter, 1995)
Geo-hydrological information • Existing groundwater models • Groundwater quality • Geo-hydrological maps	NGDB, QualDB, Field data, National groundwater maps (Vegter, 1995)
Hydrological information • Surface water bodies • Surface water levels • Gauging station records • Evaporation	Surface water hydrologists (DWAF), Field data, WR90 – rainfall (Midgley et al, 1994), National groundwater maps – baseflow (Vegter, 1995)
Other data • Land cover/Land use • Topography	Field data, CSIR - ARC, Surveyor General

4 Environmental Impacts due to Groundwater Abstraction

As discussed in section 1.2, the "Reserve" or the "Groundwater Reserve" means the quantity and quality of water required to satisfy the *basic human needs* and to protect aquatic ecosystems. In other words, the groundwater quality and the environmental impacts due to the groundwater abstraction must be estimated. Numerical models are intended to be used for solving these problems, as there is obviously no other easy way to solve complex problems involving pumping wells, recharge, rivers, pollution transport, etc.

The design steps and application of numerical models are not discussed here, as many excellent scientific books are available. For example, Anderson and Woessner (1991) introduce the basic concepts and application of groundwater models; Spitz and Moreno (1996) provide a step-by-step approach for constructing groundwater models; and Chiang and Kinzelbach (in press) provide a user-friendly computer program and numerous examples for solving groundwater flow and pollution transport problems.

As far as pure fresh groundwater aquifers are concerned, negative environmental impacts due to groundwater abstractions are mainly a result of excessive drawdown. For example, the change in flow rates of springs is proportional to the change in the hydraulic heads (groundwater levels) in aquifers. When the groundwater level drops, the flow rate in a spring drops proportionally. When the groundwater level drops below the ground surface of the spring location, the spring dries out. Another example, the change in leakage rates between groundwater and surface water bodies (rivers, wetlands, etc.) are subject to the change in hydraulic heads in aquifers and the change in water levels in surface water bodies (refer to section 2.2.2). The drawdown of hydraulic heads also causes a number of major environmental impacts, such as land surface subsidence, degradation of groundwater dependent vegetation, reduction of flow rates and water levels in rivers or saltwater intrusion. All these impacts, except the last three, can easily be estimated by using numerical models constructed by using

PMWIN (Chiang and Kinzelbach, in press). If the estimated impacts are higher than allowed, the groundwater abstraction rate must be reduced accordantly.

Problems associated with groundwater sensitive vegetation are beyond the scope of this work. The problem associated with the reduction of flow rates and water levels in rivers can be solved by using a coupled groundwater / surface water flow model (Chiang et al., 2000), which is being developed. The following sections discuss the saltwater intrusion problems and describe a computer program for solving the sharp salt/fresh water interface under the steady state condition.

Saltwater intrusion becomes a problem in coastal areas where fresh water aquifers are hydraulically connected with the seawater. When a large amount of fresh water is withdrawn from these aquifers, saltwater flows towards the pumping well due to the hydraulic gradient. As a result, the aquifer is salinized and the water quality in the pumping wells deteriorates. In serious cases, the water obtained from those pumping wells may not be used for supplying basic human needs (Groundwater Reserve). Therefore, it is necessary to estimate the salt- and freshwater interface. The saltwater intrusion problem has been intensively discussed in literature and analytical solutions have been derived. For example, Custodio and Bruggeman (1987) gives an excellent reference to global saltwater intrusion problems. Strack (1976) derives an analytical technique for solving three-dimensional interface problems in a coastal aquifer with a fullly penetrated well. However, it is restricted to steady state flow conditions in homogeneous isotropic porous media where the vertical flow component can be neglected by using the Dupuit-Forchheimer assumption (Dupuit, 1863; Forchheimer, 1886). Bruggeman (1999) gives a series of analytical solutions or semi-analytical solutions for the flow of fresh water along a sharp interface that completely separates fresh water bodies from underlying saltwater. The analytical solutions in Bruggeman (1999) are grouped into one-dimensional horizontal interface flow, two-dimensional radial-symmetric horizontal interface flow, and general two-dimensional horizontal interface flow.

Although analytical solutions are easy to apply, they are restricted to simplified cases, idealised boundary conditions and horizontal flow conditions (Dupuit- Forchheimer assumption). These restrictions can be overcome by using numerical models. Guo and

Bennett (1998) and Oude Essink (1998) use numerical solute transport models, which consider the density-driven groundwater flow and use a linear relationship between chloride concentration and the water density. Essaid (1990a, 1990b) developed the numerical model SHARP for determining the location of the sharp interface between salt- and fresh water. As SHARP does not solve the steady-state sharp interface directly, a considerable computation time is required until the unsteady-state solutions of SHARP do not change with time.

The program SEAWATER is developed for calculating the location of the steady-state sharp interface, because it is of interest for solving management problems, such as pumping rate limitations. This program is based on MODFLOW and the Ghyben-Herzberg approximation (Badon-Ghyben, 1888; Herzberg, 1901). Existing MODFLOW graphical user-interfaces, such as PMWIN (Chiang and Kinzelbach, in press) can be used to prepare data for SEAWATER. The following sections briefly describe the Ghyben-Herzberg approximation, and the SEAWATER program.

4.1 Ghyben-Herzberg Approximation

The Ghyben-Herzberg approximation was derived independently by Badon-Ghyben (1888) and Herzberg (1901). Figure 4.1 shows the idealized Ghyben-Herzberg model. Ghyben and Herzberg assume static equilibrium and a hydrostatic pressure distribution in the fresh water region with stationary seawater. Equation 4.1 describes the static equilibrium between the fresh water pressure (P_f) and the saltwater pressure (P_s).

$$P_f = \gamma_f g(h_s + h_f) = P_s = \gamma_s g h_s \qquad (4.1)$$

where γ_s [ML^{-3}] is the saltwater density, γ_f [ML^{-3}] is the fresh water density, g [LT^{-2}] is the gravity acceleration. The depth of the interface (h_s) from the sea level is

$$h_s = \frac{\gamma_f}{(\gamma_s - \gamma_f)} h_f = \delta h_f \qquad (4.2)$$

Although the Ghyben-Herzberg approximation was derived for phreatic aquifers, the approximation also applies to confined aquifers. Bear (1972) points out that the Ghyben-Herzberg approximation is a conservative estimation of the fresh/ saltwater interface. The validity of the Ghyben-Herzberg approximation is investigated in Bear and Dagan (1962)

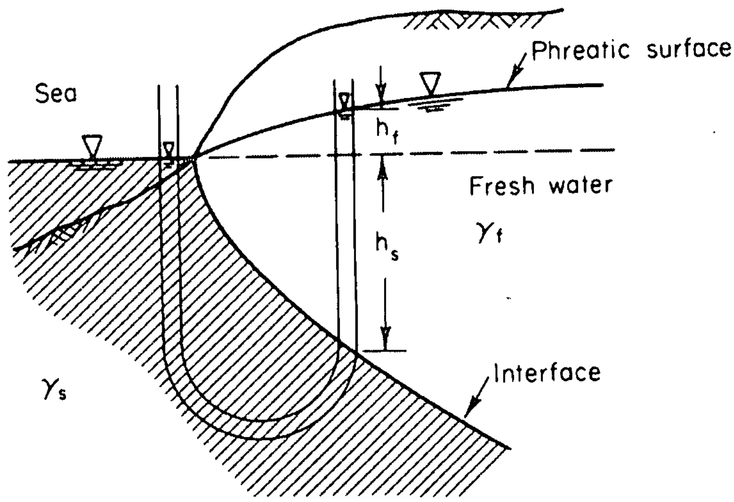


Figure 4.1 Ghyben-Herzberg interface model (Bear, 1972)

4.2 The SEAWATER Program

The SEAWATER program uses MODFLOW and the Ghyben-Herzberg approximation (Badon-Ghyben, 1888; Herzberg, 1901) to calculate the steady-state sharp interface between saltwater and freshwater by means of an iterative solution. The iterative steps are:

1. A MODFLOW flow model is constructed and in the first model layer, the model cells along the coast are specified as fixed-head cells with the head equal to the sea level.
2. A flow simulation is performed and a distribution of freshwater heads is obtained.
3. Using the Ghyben-Herzberg approximation, a new location of the interface is calculated. The model cells, which lie below the interface, are set to no-flow

(inactive) model cells, which lie above the interface, are set to variable-head (active) model cells.

4. Repeat 2-4 until the location of the interface does not change.

The iterative solution can be used together with almost all packages of MODFLOW. Note that the resolution of the result is defined by the vertical discretisation of the model. The calculated interface location can only be as precise as the layer thickness.

4.3 Compare SEAWATER with Analytical Solutions

The results of the SEAWATER program are compared with analytical solutions given in Bear (1972), Bruggeman (1999) and Strack (1976). All these solutions assume that the vertical flow rates can be neglected in relation to the horizontal ones. This is where the numerical solution is superior to the analytical solutions. The following sections discuss the solutions in detail.

4.3.1 Confined Aquifers with Lateral Freshwater Inflow

Figure 4.2 shows a vertical cross-section of a confined aquifer and the boundary conditions. The aquifer is assumed to have an infinite extent along the coast. Fresh water is flowing at a constant rate of $q = 6.6\times10^{-6}$ $m^3/s/m^2$ from the left-hand side of the aquifer and towards the sea on the right-hand side. The (horizontal) hydraulic conductivity value is K=0.001 m/s. The location of the toe of the steady-state sharp salt-/freshwater interface is (Bear, 1972)

$$L = \frac{KD^2}{2(\frac{\gamma_f}{\gamma_s - \gamma_f}) \cdot Q_0} \qquad (4.3)$$

where γ_s $[ML^{-3}]$ is the saltwater density and γ_f $[ML^{-3}]$ is the fresh water density. Q_0 $[L^2T^{-1}]$ represents the fresh water flowing toward the sea per unit length of coast. Refer to Figure 4.2 for the definitions of L and D.

The location of the interface below the sea level is

$$z^2(x) = D^2 - \frac{2Q_0 \cdot x \cdot (\frac{\gamma_f}{\gamma_s - \gamma_f})}{K} \qquad (4.4)$$

The set-up of these conditions are often referred to as the Henry's problem (1964). Refer to Pinder and Cooper (1970), Lee and Cheng (1974), Segol et al. (1975), Frind (1982), Huyakorn et al. (1987), Voss and Souza (1987), Croucher and O'Sullivan (1995) for the numerical solutions to the transition zone between freshwater and saltwater.

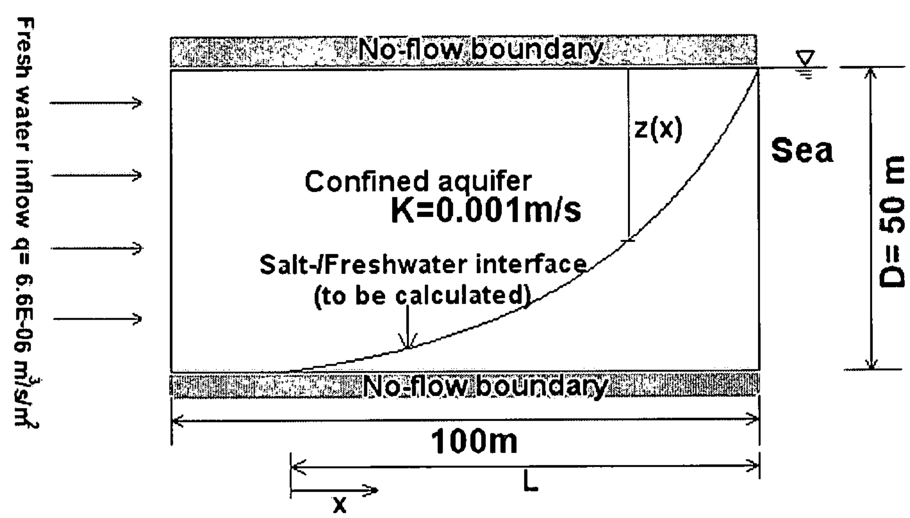


Figure 4.2 Cross-sectional view of a confined coastal aquifer

PMWIN (Chiang and Kinzelbach, in press) is used to construct a numerical flow model. Refer to Figure 4.3, the model is divided into 50 layers in the vertical direction, 100 columns and 1 row. The size of each model blocks (cells) is 1m × 1m × 1m. The cells of the first column are fixed-flux boundary cells with q = 6.6E-6 m3/s (each). The cell on the upper-right corner of the model is a fixed-head boundary cell with the head h = 0 m (sea level).

After the model is constructed, SEAWATER is used to calculate the steady-state sharp interface. Figure 4.4 shows the analytical result and the numerical results obtained by SEAWATER and the boundary-element method (Volker and Rushton, 1982). The results are in agreement. Note that as MODFLOW is a 3D-flow model, the vertical hydraulic conductivity values can be considered. In this model, the vertical hydraulic conductivity is set to a high value of 0.01 m/s. This minimizes the hydraulic gradient in the vertical direction, and thus the numerical result can be compared with the analytical result. Figure 4.5 shows the result by using a vertical hydraulic conductivity value of 0.001 m/s. In comparison with the analytical solution, which does not consider the vertical flow terms, the interface is closer to the sea because freshwater does not flow out at the same rate and builds a higher freshwater head at the left-hand side. As a result of applying the Ghyben-Herzberg approximation, the location of the interface is deeper and closer to the sea.

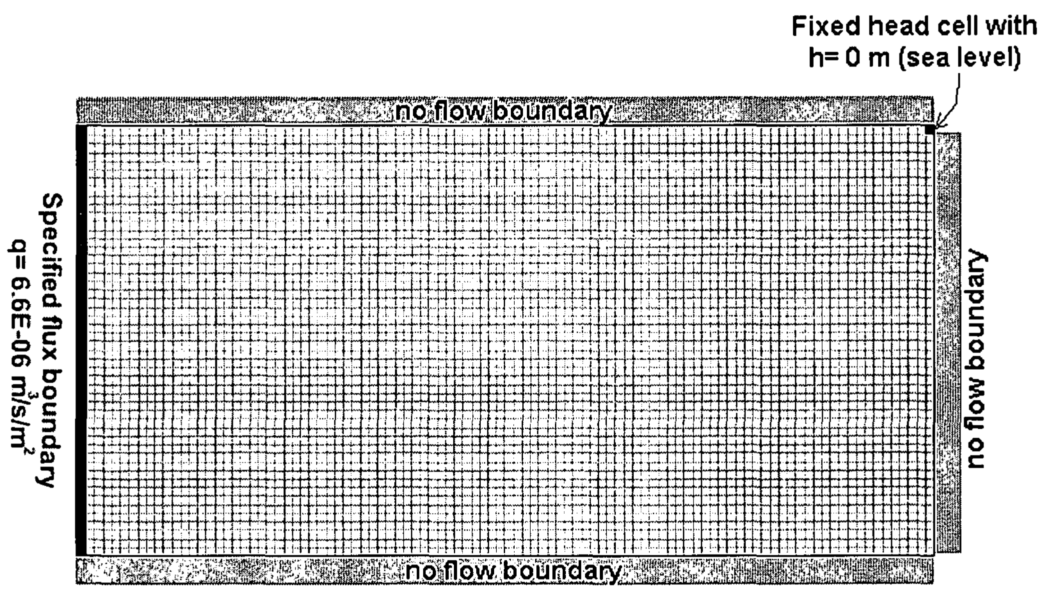


Figure 4.3 Configuration of the flow model

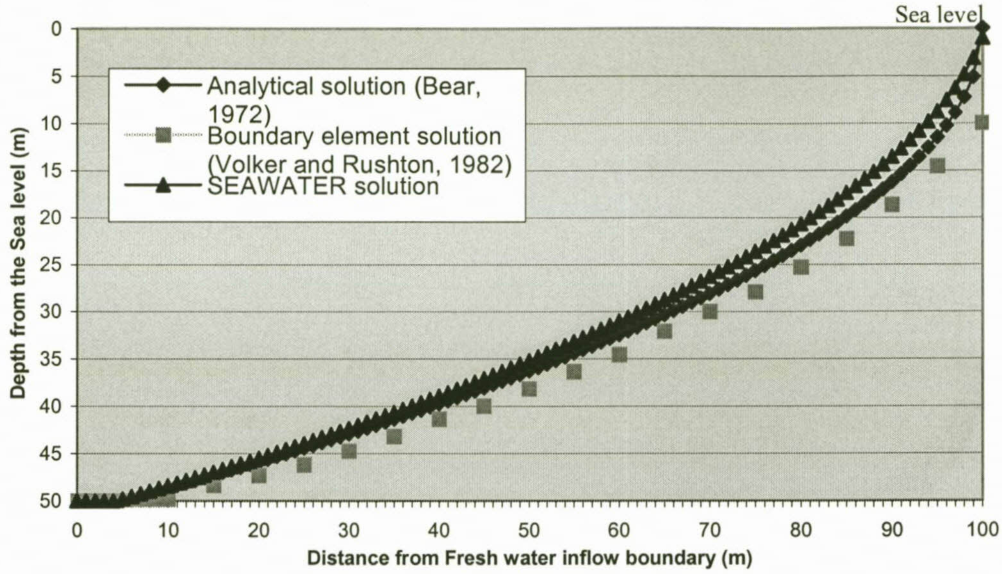


Figure 4.4 Comparison of the interface locations obtained by different methods

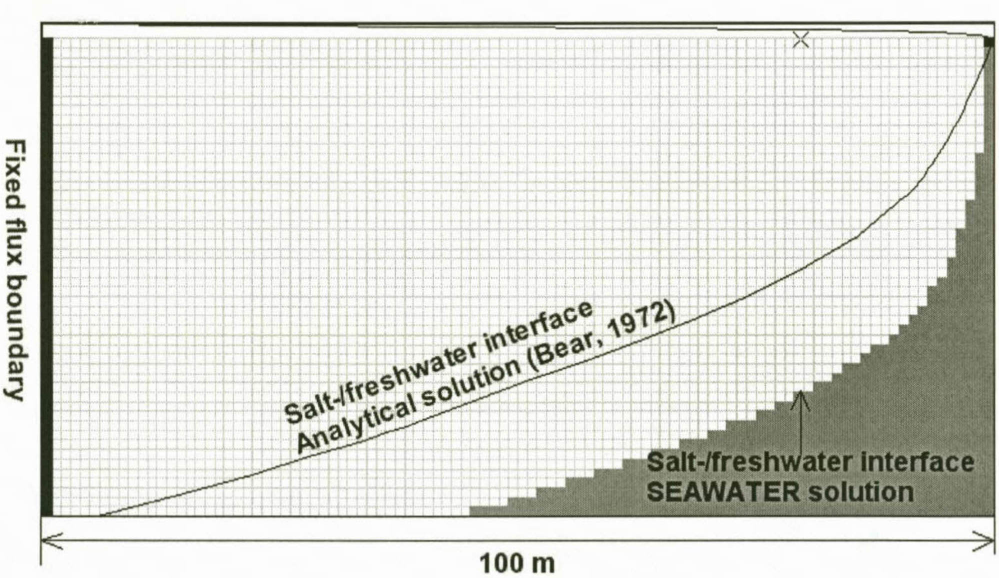


Figure 4.5 Result of SEAWATER by using a vertical hydraulic conductivity value of 0.001m/s

4.3.2 Unconfined Aquifers with Infiltration from Precipitation

Figure 4.6 shows a vertical cross-section of a costal aquifer between two open parallel seawater boundaries with infiltration. The aquifer is assumed to have an infinite extent along the coast and infinite thickness. The distance between the parallel seawater boundaries is 2×B. The location of the steady-state sharp salt-/freshwater interface is (Bruggeman, 1999)

$$z^2(x) = \frac{P}{(\frac{\gamma_s - \gamma_f}{\gamma_f}) \cdot K}(B^2 - x^2) \qquad (4.5)$$

where γ_s [ML^{-3}] is the saltwater density and γ_f [ML^{-3}] is the fresh water density and p [LT^{-1}] is the infiltration rate.

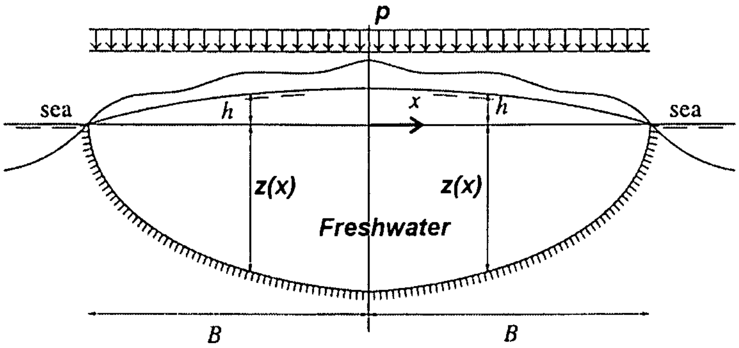


Figure 4.6 Vertical cross-section of a coastal aquifer between two open parallel sea water boundaries (After Bruggeman, 1999)

PMWIN is used to construct a numerical flow model. Refer to Figure 4.7, the model is divided into 50 layers in the vertical direction, 101 columns and 1 row. The size of each model blocks (cell) is 1m × 1m × 0.2m. The first and last cells of the first layer is fixed-head boundary cell with the head h = 0 m (sea level). The infiltration is simulated by using the Recharge package of MODFLOW (McDonald and Harbaugh, 1988). The infiltration rate is 1×10^{-8} m/s. The hydraulic conductivity value is K=0.001 m/s.

After constructing the model, SEAWATER is used to calculate the steady-state sharp interface. Figure 4.8 shows the analytical result as well as the result obtained by SEAWATER. Note that although the top elevation of the first model layer is set to 2 m for display purposes, the actual saturated thickness is used for the calculation.

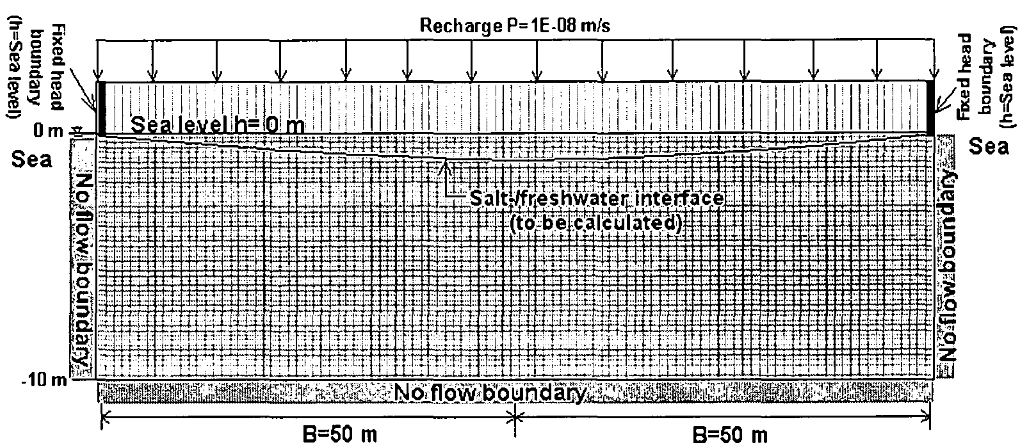


Figure 4.7 Configuration of the flow model

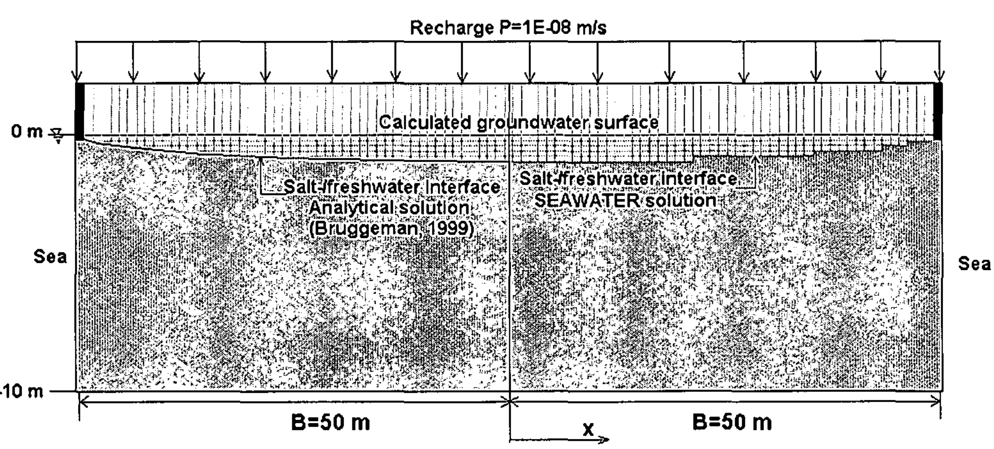


Figure 4.8 Comparison of the interface locations obtained by an analytical solution and SEAWATER

4.3.3 Upconing of the Salt-/Freshwater Interface

Figure 4.9 shows a cross-sectional view of a confined costal aquifer and its boundary conditions. The aquifer is assumed to have an infinite extent along the coast. Fresh water flows at a constant rate of $q = 6.6\times10^{-6}$ m^3/s/m^2 from the fixed-flux boundary at left-hand side of the aquifer and toward the sea in the right-hand side. The (horizontal) hydraulic conductivity value is K=0.001 m/s. A fully penetrating pumping well is located at a distance of x_w = 37.5 m from the coastal line.

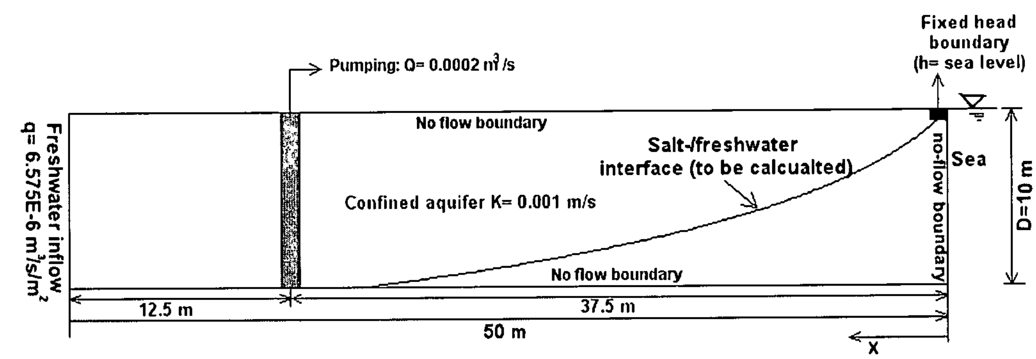


Figure 4.9 Vertical cross-section of a confined coastal aquifer with a pumping well

The location of the toe of the steady-state sharp salt-/freshwater interface (L_c) can be obtained by using Figure 4.10 (Strack, 1976). First, the values λ and μ are calculated by using Equations 4.6 and 4.7.

$$\mu = \frac{Q}{Q_0 \cdot x_w} \qquad (4.6)$$

$$\lambda = \frac{K \cdot D^2}{Q \cdot x_w} \cdot \frac{\gamma_s - \gamma_f}{\gamma_f} \qquad (4.7)$$

Where D [L] is the aquifer thickness, γ_s [ML^{-3}] is the saltwater density and γ_f [ML^{-3}] is the fresh water density, K [LT^{-1}] is the hydraulic conductivity, Q_0 [L^{2}T^{-1}] represents the discharge flowing into the sea per unit coast length. The ratio L_c/x_w can be read from the type curves shown in Figure 4.10. For the given conditions, $L_c/x_w = 0.52$, i.e. L_c = 19.5 m.

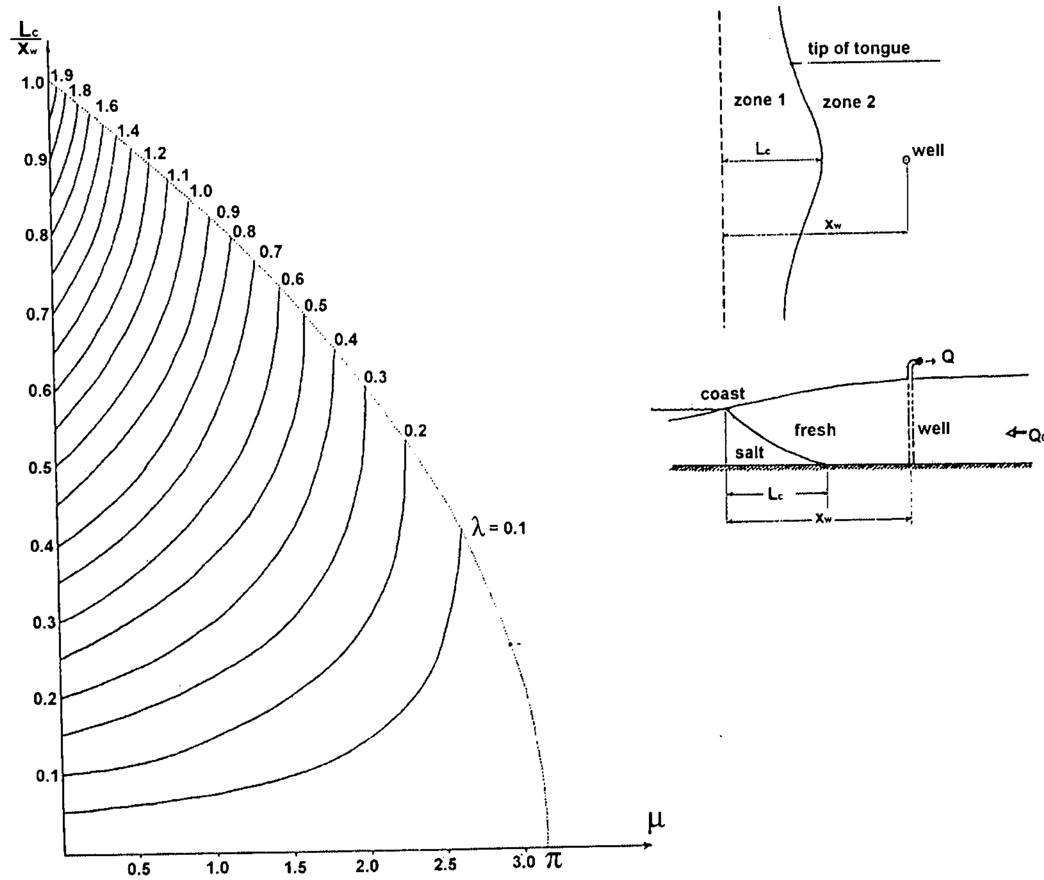


Figure 4.10 Strack's (1976) type curves for determining the location of the interface toe

A numerical flow model is constructed by using PMWIN (Chiang and Kinzelbach, in press). Refer to Figure 4.11, the model is divided into 10 layers in the vertical direction, 50 columns and 21 row. The no-flow boundaries are far enough so that the cone of depression caused by the pumping well will not reach these boundaries. The size of each model blocks (cell) is 1m × 1m × 1m. The cells of the first column are fixed-flux boundary cells with $q = 6.6\times10^{-6}$ m^3/s (each). The cells on the upper-right corner of the model are fixed-head boundary cells with the head h = 0 m (sea level).

After the model is constructed, SEAWATER is used to calculate the steady-state sharp interface. Figure 4.12 shows the numerical result (shaded cells) obtained by

SEAWATER. The dash-line representing the interface is calculated from the freshwater head using the Ghyben-Herzberg approximation. The location of the interface toe is 19 m from the coast, which is close to Strack's solution. However, the latter does not give the location of the interface and cannot be used to solve the saltwater intrusion problem, when partially penetrating wells or more than one well exist. Bear (1979) attempts to solve the upconing problems for a single pumping well with analytical methods. However, the approaches are too complicated to be considered as a practical tool. As discussed before, the SEAWATER program can be used with almost all MODFLOW packages, including drain, well, river, recharge, evaporation, etc. Figure 4.13 shows an application example. The boundary conditions are the same as shown in Figure 4.11. The pumping well is only penetrated to 5 m below the sea level with a pumping rate of 0.0006 m^3/s. The upconing effect of the saltwater is clearly visible. Figure 4.14 shows a three-dimensional image of the interface. The image was created with the visualization software named Groundwater Explorer (Bekker and Chiang, in prep.).

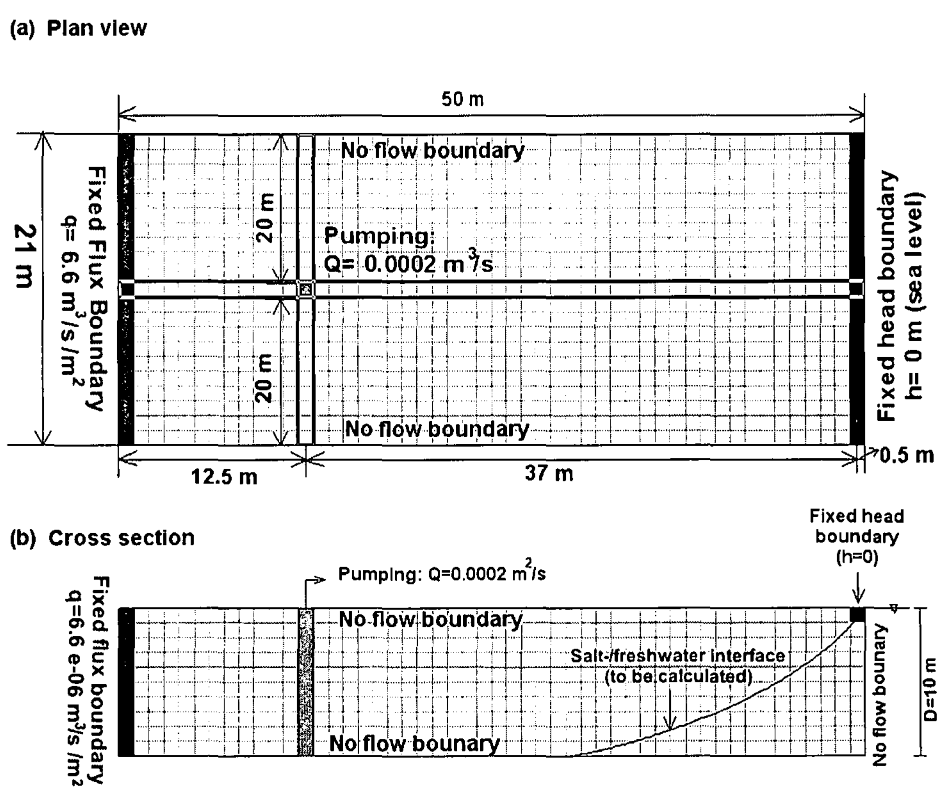


Figure 4.11 Configuration of the flow model

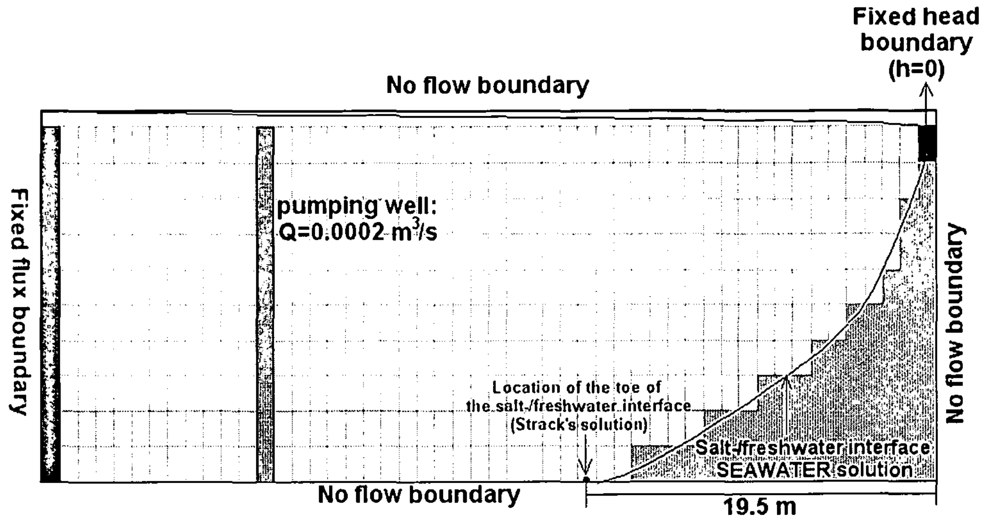
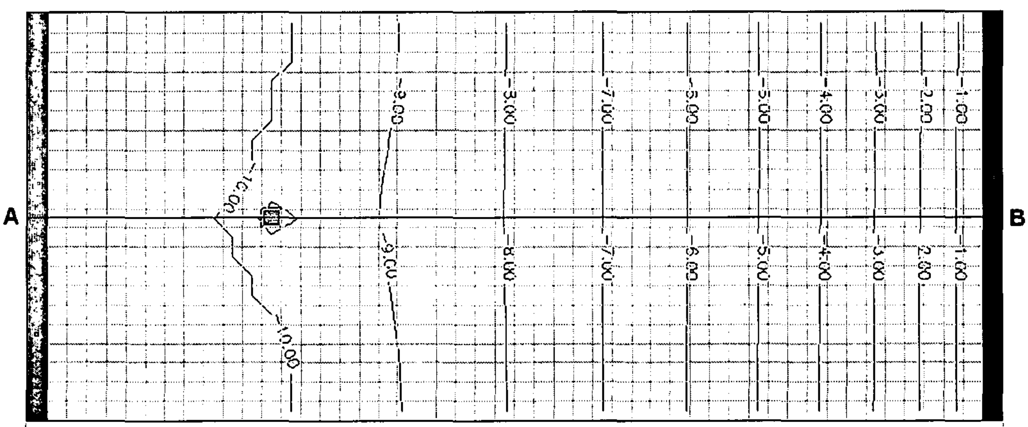
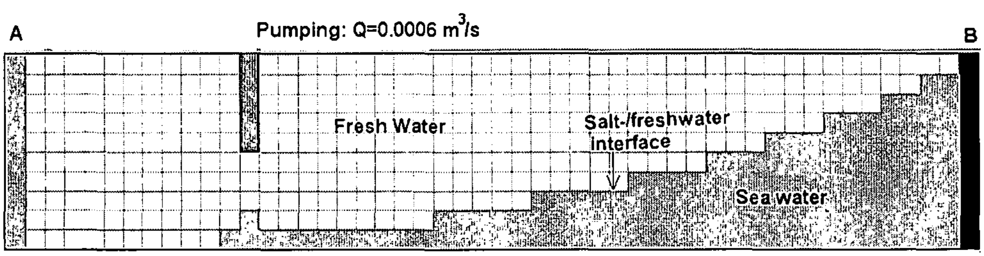


Figure 4.12 Comparison of the results obtained by Strack's solution (1976) and SEAWATER



(a)



(b)

Figure 4.13 An application example. (a) Interface elevation contours in a plan view; (b) Interface in cross-section AB

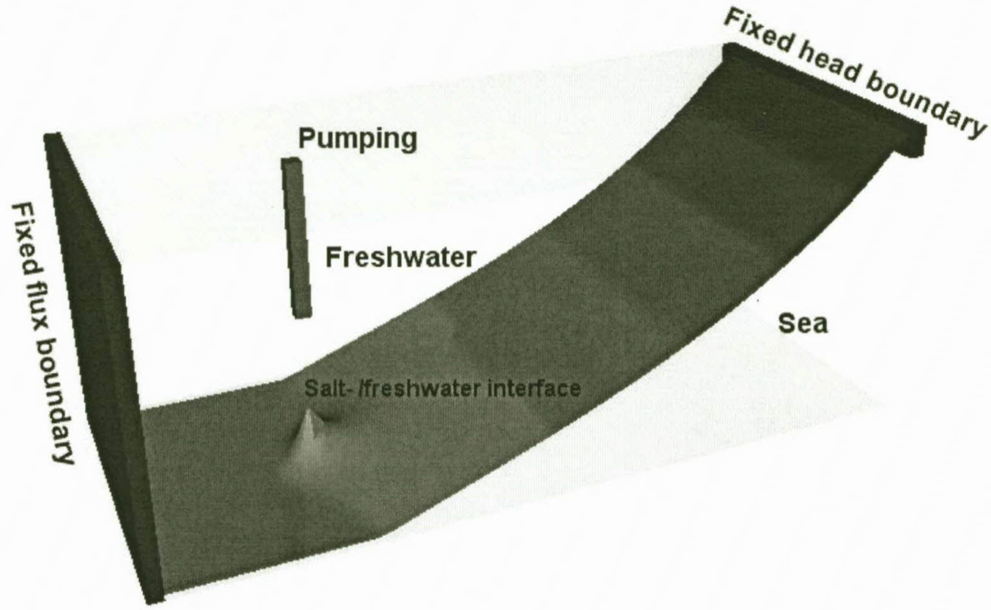


Figure 4.14 A three-dimensional image of the sharp salt-/freshwater interface calculated by SEAWATER

5 Case Study: Preliminary Estimation of the Groundwater Reserve in the Pienaars River Catchment

In this Chapter, the methodology described in Chapter 3 is used to estimate the Groundwater Reserve in the Pienaars River Catchment. The estimated groundwater reserve represents the maximum amount of groundwater resources which can theoretically be allocated to water users. To secure the ecological reserve, however, it is necessary to evaluate possible environmental impacts due to groundwater abstraction prior to licensing any water use. Some methods for evaluating local impacts are described in Chapter 4.

As required by DWAF, the estimation of the preliminary Groundwater Reserve should be based on the current available data and should be finished within a period of 60 days. The required time for completion of each of the steps are given below. It is emphasized that the calculation only takes 10 days. Most of the time is spent on the data gathering, as it is the basis of the estimation.

The following steps and times are required by the water balance model. Except for the last step, every step is discussed in separate sections in this Chapter.

1. Gathering data and understand the study area (40 days)
2. Evaluation of the current groundwater abstraction (2 day)
3. Estimation of Groundwater Recharge (2 day)
4. Estimation of fluxes between groundwater and surface water (2 days)
5. Estimation of subsurface flow (2 days)
6. Estimation of the basic human needs and ecological reserve (5 day)
7. Balance of all estimations (1 day)
8. Report writing (6 days)

5.1 Gathering Data and Understanding the Study Area

Geology

The study area is located to the north of Pretoria along the Pienaars River and covers 4800 square kilometres. The boundaries of the study area coincide with the boundaries of the surface catchments (watersheds). Groundwater resources in this area were extensively developed in the water supply programs of part of former homelands and self-governing states (Hobbs 1996; Parsons, 1999). The area includes a part of the Springbok Flats aquifer which has been studied by Tredoux (1993). The geological conditions in this area is very complex. The area consists of sedimentary rocks, including sandstone, shale, grit, conglomerate, dolomite, etc., and igneous rocks, including gabbro, andesite, basalt, carbonatite, granite, ..., etc. (Figure 5.1).

As extensive information about the geological setup of this area is not available, the underground is treated as a single aquifer.

Surface water

The Pienaars River is the largest river in the study area. The river initially flows from south to north, then changes its direction to the west in the centre of the study area, and flows into the Klipvoor Dam. A wetland with an extent of about 20 km x 3 km exists in the centre of the study area. There are tributaries from each of the Quaternary catchments to the Pienaars River (Figure 5.2).

River fluxes for some gauging stations are available and their positions of the stations are shown in Figure 5.3. However, the distribution of gauging stations and the available data are not sufficient for calculating the water balances (evaporation, inflow, outflow, and interaction with groundwater) for the different river segments.

Topography

The topographical map of the study area shows that elevations of Pienaars River and its tributaries are lower than their environment (Figure 5.4). From a geomorphological viewpoint, this is a very common situation, as river valleys are formed by the erosion of the ground surface through water movement.

Groundwater

According to the National Groundwater Date Base, there are more than 467 registered boreholes in the study area (Figure 5.5). The data is fragmented and very little historical groundwater level information is available. Groundwater level hydrographs at some selected boreholes (Figure 5.6) indicate that the groundwater level varies temporally.

As the quality of the historical data is not good enough for calculating the very long-term average groundwater level, it was decided to use the average groundwater level after 1990. A groundwater level contour map (Figure 5.7) is generated by the Bayesian interpolation method (Van Tonder et al., 1996). According to Figure 5.7, groundwater follows the topography and flows towards the lower parts of the study area and recharges the wetland and rivers. The flow rate from groundwater into the wetland and rivers needs to be calculated in this case study (see section 5.4).

The Baysian method is used because the groundwater level is well correlated to the groundwater surface and therefore the topography can be used as additional information for interpolating the groundwater surface. Figure 5.8 shows the correlation between the groundwater level and the ground surface. In this figure, the water levels were plotted against the ground surface elevations. A linear trend was fitted. The correlation between ground surface and groundwater level is calculated by Equation (5.1). In this case, the correlation coefficient is 0.98 (98%).

$$\rho_{X,Y} = \frac{Cov(X,Y)}{\sigma_X \cdot \sigma_Y} \qquad (5.1)$$

where

$\rho_{X,Y}$	correlation coefficient of groundwater level and ground surface
$Cov\ (X,Y)$	covariance of groundwater level and ground surface
σ_X	standard variance of groundwater levels

σ_Y standard variance of ground surface

X groundwater level data set

Y ground surface data set

The fact that groundwater surface is correlated with the topography implies that the boundaries at the study area coincide with the groundwater divides.

According to the available information, no groundwater contamination is observed, except the Springbok Flats, where higher nitrate values have been measured (Tredoux, 1993). Over-fertilizing or pit latrines could be the cause.

5.2 Estimation of the Current Groundwater Abstraction

Detailed abstraction information in this area is not available. According to the National Groundwater Data Base (NGDB), there are 467 registered boreholes within the study area, and the total abstraction is 165.73 l/s, equal to 5.2 Mm^3/a. It is possible that this amount is less than the real abstraction. Parsons (1999) estimated the existing abstraction (mainly for the irrigation and human needs purpose) as 20 Mm^3/a, however this value is not reasonably explained. According to the land use map of the study area (Figure 5.9), about 10% of area (480 km^2) is cultivated land. There is no data to indicate how much cultivated land is irrigated by groundwater, it is assumed that 10% of cultivated land depends on groundwater. The estimated demand of irrigated water for vine, orchard and cereal per annum is about 600mm (Zietsman, et al., 1996). Therefore, about 28.8 Mm^3/a of groundwater is abstracted for irrigation purposes. At present, about 800000 people depend on groundwater for their human needs. There is no data to exactly indicate how much groundwater is pumped out for human needs, so the 25 l/day/person (National Water Act) is used. This implies that 7.31 Mm^3/a of groundwater is used and must be reserved to secure the basic human needs. In total, an amount of about 36.1 Mm^3/a of the groundwater abstraction is estimated.

5.3 Estimation of Groundwater Recharge

The computer program RECHARGE (Van Tonder and Xu, 1999) is used to estimate the average annual net groundwater recharge. The available methods are described below. The input data of each method are given in Table 5.1.

5.3.1 Chloride Method

The Chloride method assumes that the chloride is a conservative tracer, and enters the soil as part of the infiltrating rainwater and is subsequently concentrated by transpiration from plant and by direct evaporation from soil. If the principal source of chloride in groundwater is from atmosphere, then the recharge can be expressed as (Houston, 1987):

$$GR = P \cdot \frac{Cl_r}{Cl_{gw}} \tag{5.2}$$

where R [LT^{-1}] is groundwater recharge; P [LT^{-1}] is the mean annual precipitation; Cl_r [ML^{-3}] is chloride concentrate in rain; Cl_{gw} [ML^{-3}] is chloride concentration of groundwater. Cl_{gw} [ML^{-3}] should be the harmonic mean of the chloride concentration in the case of many boreholes.

Figure 5.10 shows the chloride concentration contours of groundwater in the study area. The harmonic mean of chloride concentration in the groundwater is 50 mg/l. The chloride concentration in the rainwater is 1.2mg/l. The average annual precipitation is 687 mm. Insert these values into Equation (5.2), the groundwater recharge can be obtain: GR=16.5 mm/a.

5.3.2 Saturated Volume Fluctuation (SVF) Method

The SVF method was introduced by Van Tonder (1989). It is based on the saturated water balance, which entails the solution of the following equation:

$$R + (I - O) - Q = S \cdot \Delta V \tag{5.3}$$

where R $[L^3T^{-1}]$ is groundwater recharge; I $[L^3T^{-1}]$ is the inflow into aquifer; O $[L^3T^{-1}]$ is the outflow from aquifer; Q $[L^3T^{-1}]$ is the withdrawal from aquifer; S [dimensionless] is the specific yield and ΔV $[L^3T^{-1}]$ is the change in saturated volume.

If the inflow terms are balanced by the outflow terms, the change in groundwater-storage is zero ($\Delta V=0$), then Equation (5.3) can be rewritten as:

$$R = Q - (I - O) \qquad (5.4)$$

Because no continuous observation of groundwater level is available in the study area, the monthly average groundwater level is used. Figure 5.11 shows the average drawdown and rainfall during June 1984 to Nov. 1994. Although the groundwater level is varying with time, but the general tendency is stable, and the variation of groundwater level is correlated with the rainfall, Equation (5.4) can be used to approximately estimate the groundwater recharge. The withdrawal from the aquifer Q is equal to 28.8 Mm^3/a (see section 5.2), the inflow into aquifer over boundaries I is equal to 4.62 Mm^3/a (see section 5.4), the outflow from the aquifer over the boundaries O is equal to 30.05 Mm^3/a (see section 5.4 and 5.5). The study covers an area of 4800 km^2. Insert these values into Equation (5.4), groundwater recharge can be obtained: R=11.5 mm/a.

5.3.3 Cumulative Rainfall Departure (CRD) Method

It is not certain who first used the cumulative rainfall departure (CRD) method, but it can be traced back to 1936. Wenzel (1936) demonstrated that the CRD series corresponds remarkably well with the fluctuation of the groundwater level. Many people have used this approach to derive the piezometric response to rainfall (Smit, 1977; Kok, 1981; Taylor, 1983; Temperley, 1980). They clearly showed that the natural groundwater level fluctuation depends on the rainfall departure. Bredenkamp et al. (1995) gives a detail introduction about theory and application of CRD method. Gerrit (1999) gives an equation to calculate the groundwater level by using CRD. According to him, the cumulative rainfall departure can be written as:

$$CRDi = CRD_{i-1} + (P_i - C) \qquad (5.5)$$

where CRD_i [L] is the cumulative rainfall departure for month i; CRD_{i-1} [L] is the cumulative rainfall departure for month i-1; Pi [LT^{-1}] is the precipitation. C [LT^{-1}] is the cut off of the rainfall.

The groundwater level can be written as:

$$h_i = h_{i-1} + F(CRD_i - CRD_{i-1})/S + (I_i - Q_i)/(S * A) - Q_i/(S * A) \qquad (5.6)$$

where h_i [L] is the head (m) in month i; h_{i-1} [L] is the head in month i-1; CRD is the cumulative rainfall departure. I [L^3/T] is the inflow into aquifer; O [L^3/T] is the outflow from aquifer; Q [L^3/T] is the withdrawal from aquifer; S [dimensionless] is the specific yield; A [L^2] is the area of study area; F is a fraction, and can be expressed by using following equation:

$$F = \frac{R_e}{P - C} \qquad (5.7)$$

where the R_e [LT^{-1}] is the effective recharge. P and C have the same mean as above.

Insert Equation (5.5) into Equation (5.6), then:

$$h_i = h_{1-1} + F(P_i - C)/S + (I_i - Q_i)/(S * A) - Q_i/(S * A) \qquad (5.8)$$

Three parameters are unknown in Equation (5.8): F, C and S. The RECHARGE program calculates these parameters by fitting the observed water level and calculated water level.

For this case study, input same data with last section into CRD sheet of RECHARGE, and arrange the three unknown parameters until a reasonable or acceptable fit between observed groundwater level and calculated groundwater level is reached. Figure 5.12

shows the fit. The cut off of rainfall is 32 mm/month, the specific yield is 9.5E-3, and the fraction is 2.6%. The groundwater recharge is then 11.5 mm/a.

5.3.4 Qualified Guess

The qualified guess of recharge is performed by using existing data, expert opinion and interpolation of known values. In this case study, seven qualified guess methods are used: soil and vegetation information, geological information, Vegter'map, ACRU map, Harvest Potential map, Base flow map, and expert's guesses. The data used for these qualified guess methods are shown in Table 5.1.

The groundwater recharge values obtained by using different methods of the computer program RECHARGE are listed in Table 5.2. The weighted average recharge is 14.1 mm/a. It means that the net groundwater recharge rate due to precipitation in the study area is 67.78 Mm^3/a.

Table 5.1 Data used by the RECHARGE program

Methods		Data
Chloride method		Average chloride concentration in groundwater (50 mg/l) Average chloride concentration in rainfall (1.2 mg/l) Average annual rainfall (687 mm)
Saturated Volume Fluctuation (SVF) method		Monthly abstraction from aquifer (2.4E+06 m^3/month) Lateral outflow and inflow (outflow: 82328 m^3/d, inflow: 12658 m^3/d) Monthly water level (average monthly water level from June 1984 to Nov. 1994)
Cumulative Rainfall Departure (CRD) method		Lateral outflow and inflow (outflow: 82328 m^3/d, inflow: 12658 m^3/d) Monthly abstraction from aquifer (2.4E+06m^3/month) Monthly water level (average monthly water level from June 1984 to Nov. 1994) Monthly precipitation (monthly precipitration from June 1984 to Nov. 1994)
Qualified Guess	Soil and vegetation information	Percentage coverage of area with soil material (loam: 80%, clayey: 20%) Percentage coverage of area of woods/trees, grass lands and bare soil (tree: 30%, grass: 35%)
	Geology	Geology (Sandstone, mudstone Sand 60%; hard rock 40%) Soil cover (>0.5m 90%, <0.5m 10%)
	Vegter	Vegter's map (Recharge from the map: 16 mm/a)
	Acru	ACRU map (recharge from the map: 14 mm/a)
	Harvest potential	Harvest Potential Map (Harvest potential: 16000 $m^3/km^2/a$)
	Base flow	Groundwater Component of River Base Flow Map (Groundwater component of river base flow: 10 mm/a)
	Expert's guess	Dave Bredenkamp: 16.49 mm/a Gerrit Van Tonder: 13.74 mm/a Dziembowski: 14.46 mm/a

Table 5.2 Estimated recharge values for the Pienaars River study area

Summary of Recharge		MAIN	
Pienaars River			
Method	mm/a	% of rainfall	Certainty (Very High=5 ; Low=1)
Cl	16.5	2.4	4
SVF: Equal Volume	11.5	1.7	4
CRD	11.5	1.7	4
Qualified Guesses :			
Soil	13.8	2.0	3
Geology	16.1	2.3	3
Vegter	16.0	2.3	3
Acru	14.0	2.0	3
Harvest Potential	16.0	2.3	2
Expert's guesses	14.9	2.2	2
Base Flow (minimum Re)	10.0	1.5	1
^2H displacement method			
Groundwater Flow Model			
Average recharge	14.1	2.1	

Estimated Recharge Value for the study area = 67.78 Mm^3/a

Area (Km^2) =	4800.0
Annual Rainfall (mm) =	687.0

5.4 Estimation of the Leakage Rate between Groundwater and Surface Water

To estimate the flow between groundwater and surface water, the computer program PMWIN5 (Chiang and Kinzelbach, in press) is used (refer to section 3.2.2) and the following steps are performed:

1. The study area is subdivided into 500 m x 500 m blocks (Figure 5.13), the aquifer is conceptualized as one model layer.
2. The Baysian interpolated groundwater level is assigned to each block.
3. Assign transmissivity values to each block.
   As no transmissivity value is available, borehole yields and empirical equations (Kirchner and Van Tonder, 1995) are used to estimate the Transmissvity values, which are shown in Figure 5.14. These values are assigned to each block.
4. Assign 1.0 as the storage value of each block.
5. Transient flow simulation is performed for 1 (one) second.
6. Calculate flux between groundwater and surface water
   The groundwater in- and outflow from or to surface water bodies are calculated by using the water budget calculator of PMWIN. A strip along the river, wetland, and the Klipvoordam is assigned as a subregion. As the hydraulic conductivity is normally larger and the hydraulic gradient is smaller in the alluvia basin of a river, different strip-width of 0.5, 1.5, 3.5 and 4.5 km have been used in order to get an average estimate of the groundwater flow rate. The groundwater flow to or from the subregion with different strip-width is shown in Table 5.3. It shows that the flow rates remain nearly constant for a width above 1.5 km. Using the average of these flow rates, the groundwater inflow rate into the subregion (rivers, wetland, dam) is estimated at 26.2 Mm^3/a and outflow rate (from the river, wetland, dam to aquifer) is 4.62 Mm^3/a.

5.5 Estimation of the Leakage Rate from/to Adjacent Aquifers

Subsurface flow includes the groundwater flow horizontally over boundaries of the study area and leakages vertically to/from other aquifers. As the subsurface was treated as a single aquifer, only the estimation of groundwater flow horizontally over the boundaries are necessary. Similar to section 5.4, the computer program PMWIN5 is used to estimate this quantity (refer to section 3.2.4). The steps 1 to 5 are the same as those of section 5.4. The last step is:

6. Calculate groundwater flow over boundaries
   The water budget calculator of PMWIN5 is used. A strip along the catchment boundary is assigned as a subregion. The width of the strip is varied from 0.5, 1 to 1.5 kilometers. The groundwater flow with different strip-widths is shown in Table 5.4. The inflow rates are set to zero, because the boundaries are groundwater divides. Using the average of these flow rates, the groundwater outflow rate is 3.85Mm^3/a.

Table 5.3 Calculated leakage rates between groundwater and surface water for different strip-widths

Strip-width (km)	Inflow (Mm^3/a)	Outflow (Mm^3/a)
0.5	20.45	9.27
1.5	26.01	5.53
2.5	26.01	4.52
3.5	26.61	3.78

Table 5.4 Calculated leakage rates over boundaries for different strip-widths

Strip-width (km)	Outflow (Mm^3/a) (Flow out of the study area)
0.5	3.33
1.0	4.17
1.5	3.85
Average	3.79

5.6 Estimation of Basic Human Needs and Ecological Reserve

The basic human need (25 l/day per person) for 800000 people is 7.31 Mm^3/a.

Groundwater, surface water (lake, dam, etc.), run-off, direct rainfall and even municipal discharge contribute to the in stream-flow of a river. Rainfall is occasional but many rivers are perennial, fed by groundwater. It is suggested to treat the portion of groundwater which discharges into the surface water as a required ecological reserve. In this case study, the ecological reserve for groundwater is 26.2 Mm^3/a. This amount may be reduced if it is proved that a reduced amount will not cause problems to the ecosystems or requirement of down-stream areas. A study regarding this point is beyond the scope of present study and being processed by other researcher.

5.7 Determine the Groundwater Reserve and the Allocatable Groundwater Resource

All elements of the water balance model for the study area are summarized below:

Inflow to the aquifer (groundwater):

Net groundwater recharge:	67.78 Mm^3/a
Leakage rate from surface water:	4.62 Mm^3/a
Leakage rate from horizontally adjacent aquifers:	0 Mm^3/a
Leakage rate from vertically adjacent aquifers:	0 Mm^3/a
Total inflow:	72.4 Mm^3/a

Outflow from the aquifer (groundwater):

Existing abstraction rate:	36.11 Mm^3/a
Leakage to surface water:	26.2 Mm^3/a
Leakage rate to horizontally adjacent aquifers:	3.85 Mm^3/a
Leakage rate to vertically adjacent aquifers:	0 Mm^3/a
Total outflow:	66.16 Mm^3/a

Groundwater Reserve:

Basic human needs (BHN):	7.31 Mm^3/a
Ecological reserve (=discharge into surface water)	26.2 Mm^3/a

Allocatable Groundwater Resource:

Allocatable Groundwater Resource = Total Inflow - Existing Abstraction (excluding BHN) - BHN – Ecological reserve

= 72.4-28.8-7.31-26.2 =10.09 Mm^3/a

5.8 Discussion of the Results

It is the principle of mass conservation, that the total inflow of groundwater must be equal to the total outflow. The result of this case study, however, does not follow this principle. There are some uncertainties to be considered:

1. Uncertainty in the estimation of flow between groundwater and surface water.
   As the estimation is based on Darcy's law, the value is directly proportional to the transmissivity value. Consequently, this uncertainty results from the uncertainty in the value of the aquifer transmissvity. It is possible to reduce this uncertainty if a better estimation of this value can be made by selecting a river section and setting up two surface water flow measurement stations at the upstream and downstream sides of this section. The (equivalent) transmissivity value over the section can be obtained by a water balance study. The order of this transmissivity value will at least give an idea of the actual tramsmissivity value over the study area.

   To understand the relation between groundwater and the wetland, it is suggested that flow measurement facilities be set up at the upstream (inflow) and the downstream (outflow) sites of the wetland. A rainfall and an evaporation station close to the wetland will also help to establish the water balance calculation for the wetland. The water exchange rate between the wetland and groundwater can be estimated by water balance.

2. Uncertainty in the estimation of the leakage rates from/to adjacent aquifer.
   This uncertainty results from the uncertainty in the transmissivity values, geological setup and the groundwater levels. The first one causes a similar problem to the one described in 1. By treating the study area as one single aquifer (due to the unknown complexity of the geological setup), possible deep groundwater flow is ignored. Deep groundwater flows are often important resources (e.g. hot spring) for lower areas. A better understanding of aquifer parameters, aquifer structure and the groundwater flow pattern can help to reduce this uncertainty. The required work, however, can be enormous.

3. Uncertainty in the estimation of existing groundwater abstraction.

   This is a statistical work. If registration data are incomplete or do not exist, the existing groundwater abstraction rate must be estimated by using other information, for example land use maps or irrigation requirements. Other unknown groundwater losses, for example spring flow, can also cause the uncertainty.

4. Uncertainty in the estimation of groundwater recharge.

   Although the estimation of groundwater recharge is uncertain in its nature, the values obtained by using different methods remain in a certain range. In this case study, the value of the groundwater recharge is the most reliable one.

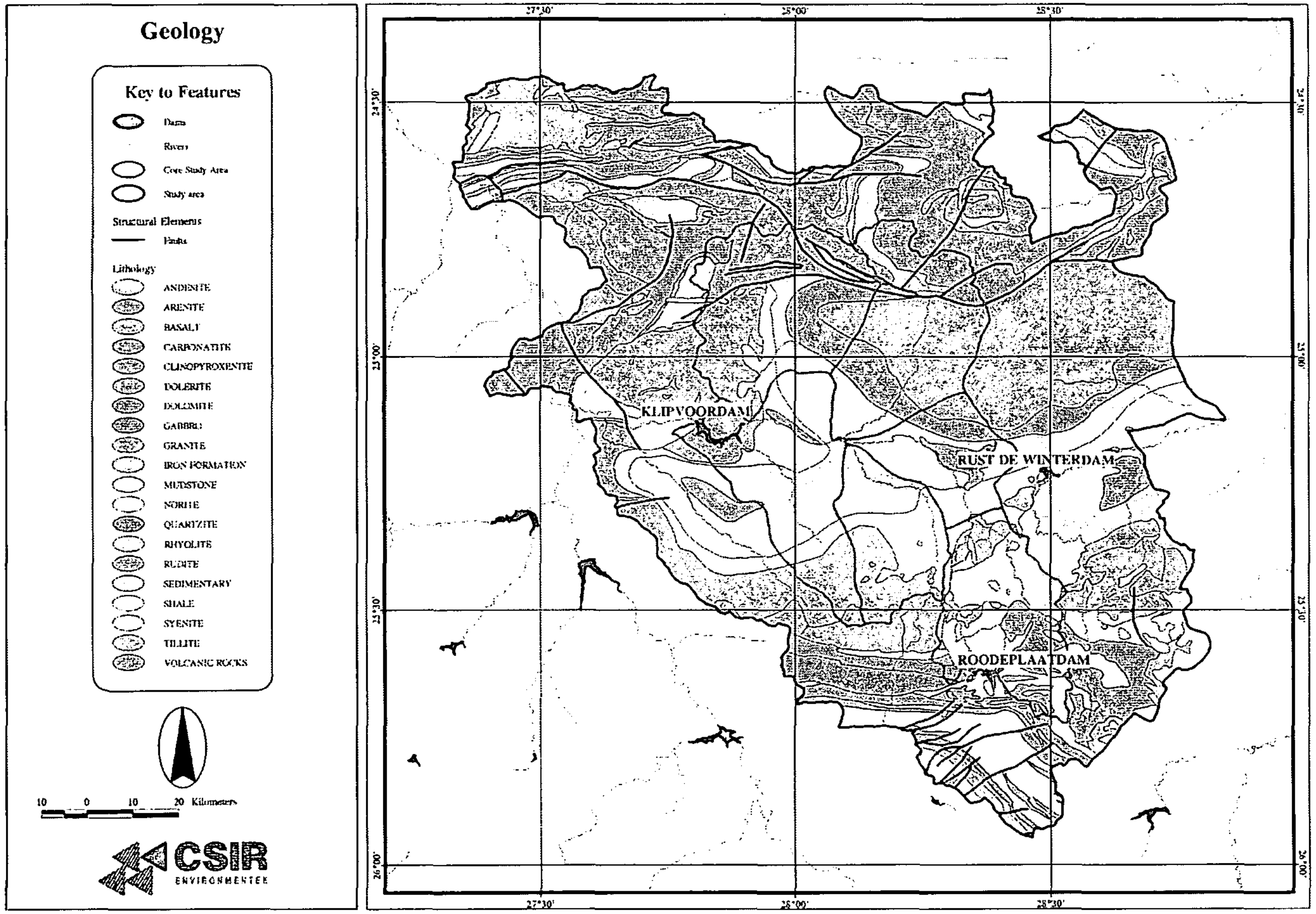


Figure 5.1 Geological map of the Pienaars River study area

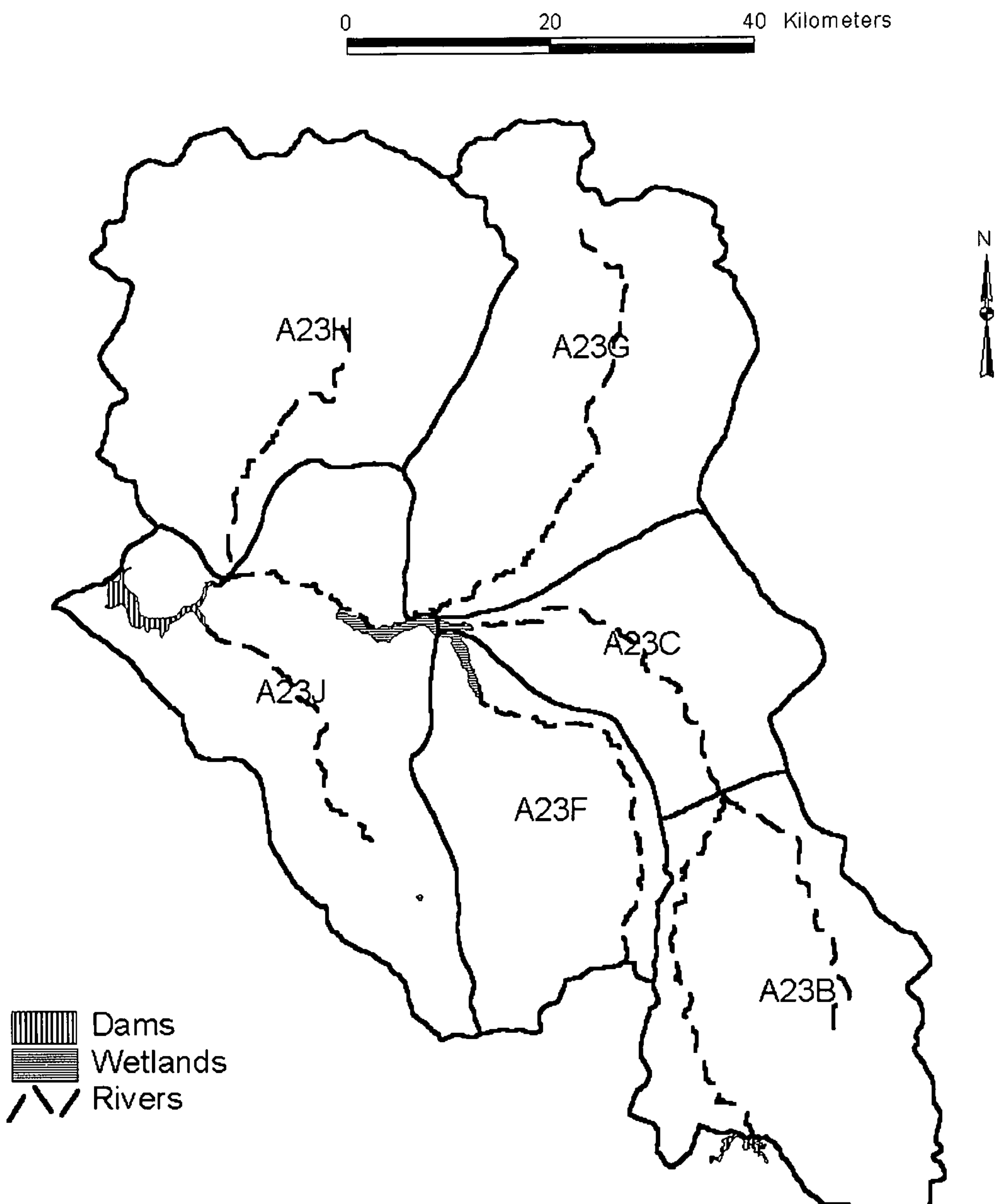


Figure 5.2 Quaternary catchments and surface water bodies within the Pienaars River study area

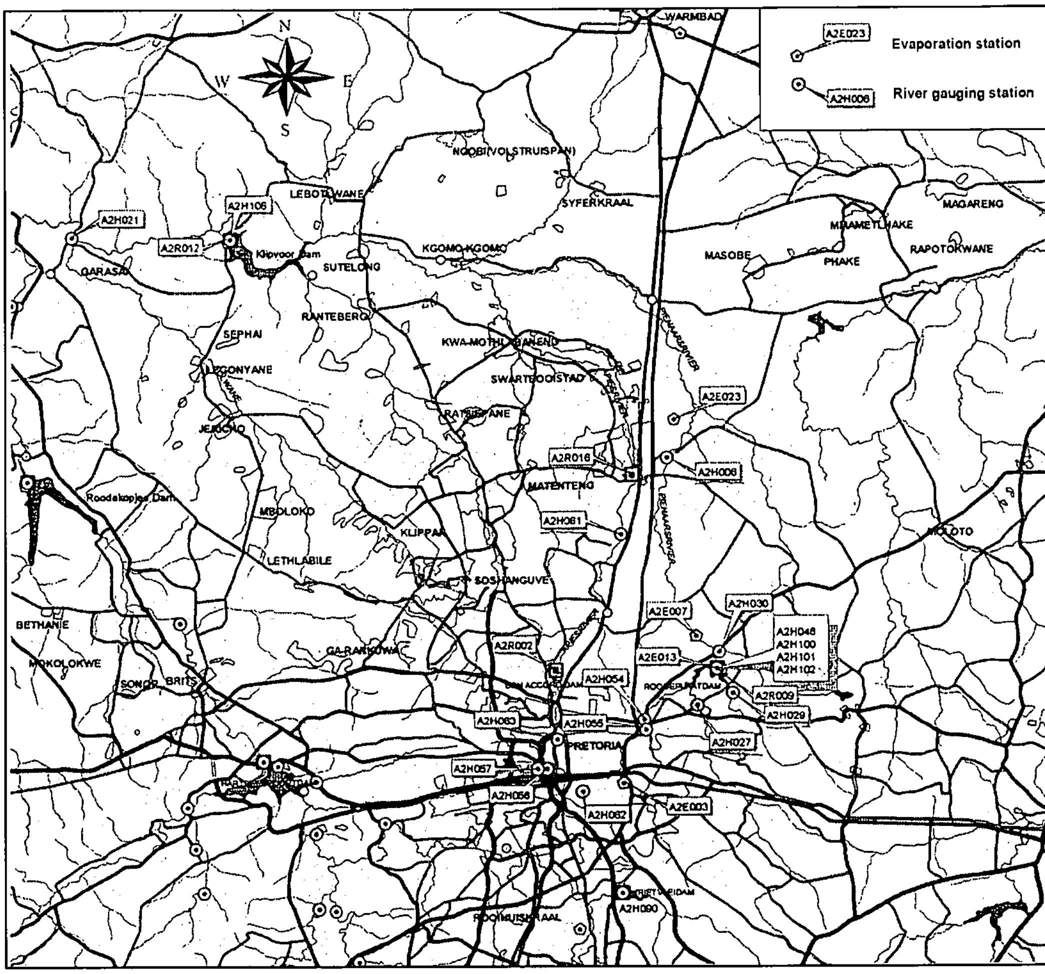


Figure 5.3 Positions of the gauging stations

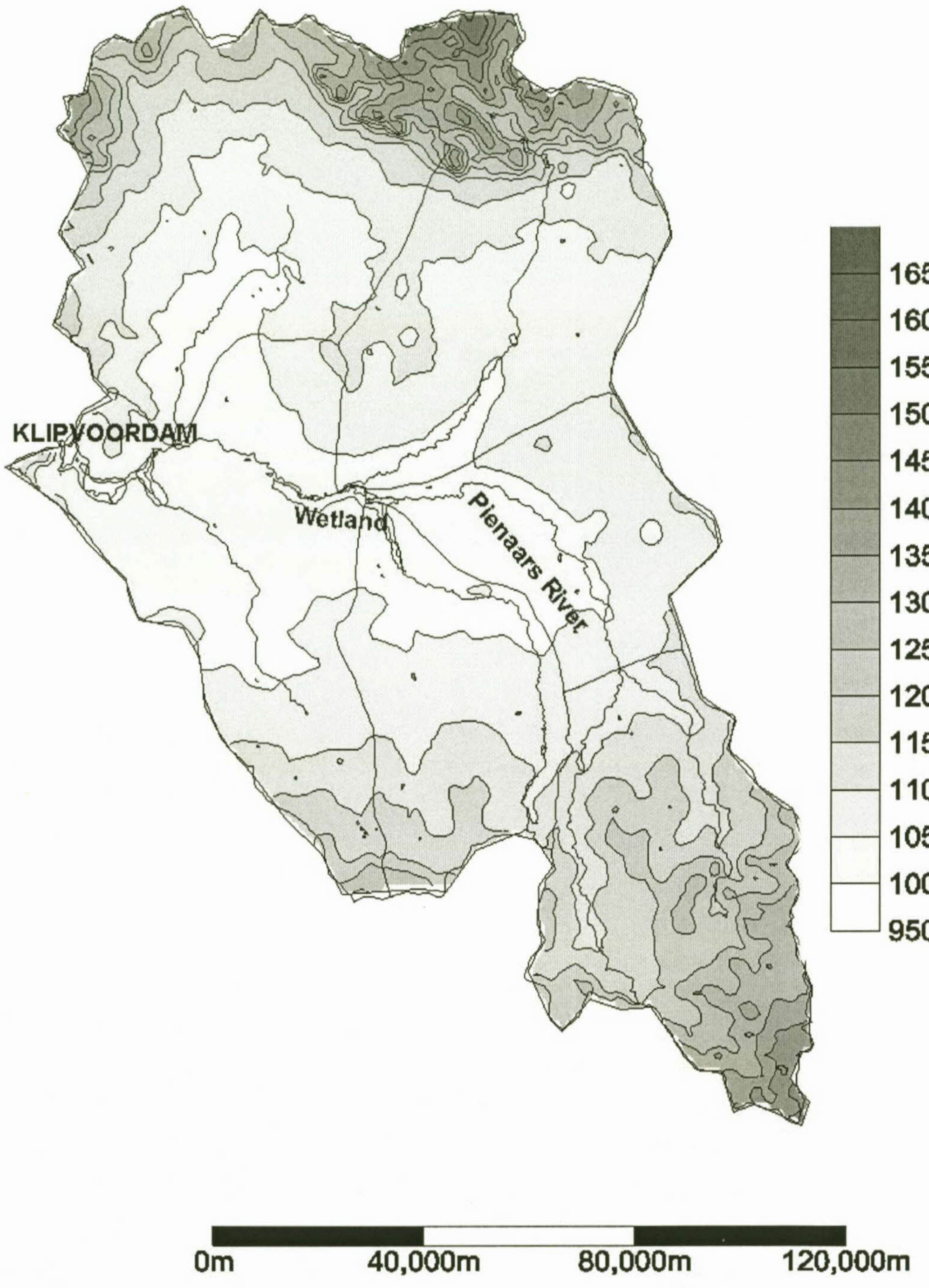


Figure 5.4 Topographical map of the Pienaars River study area

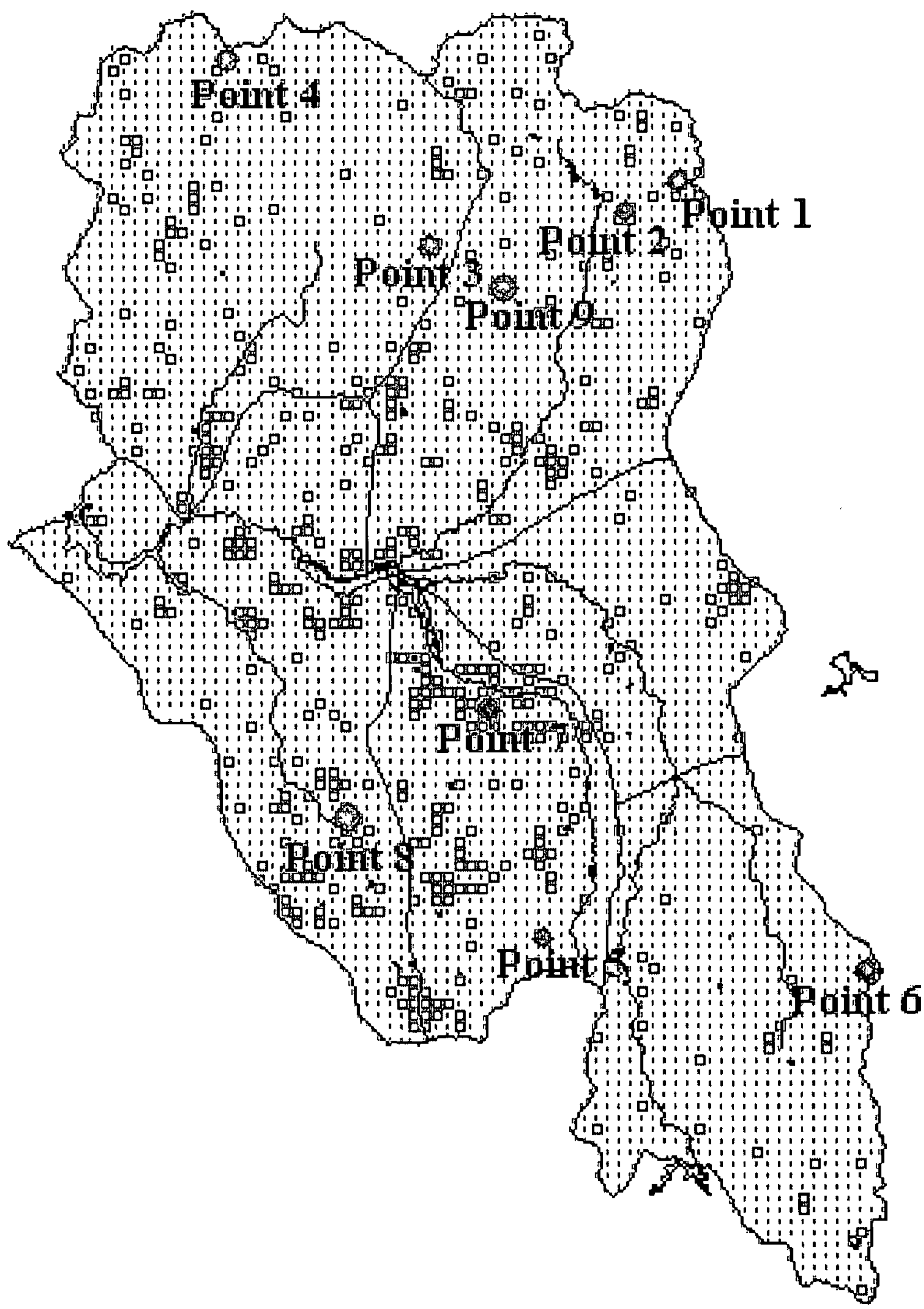
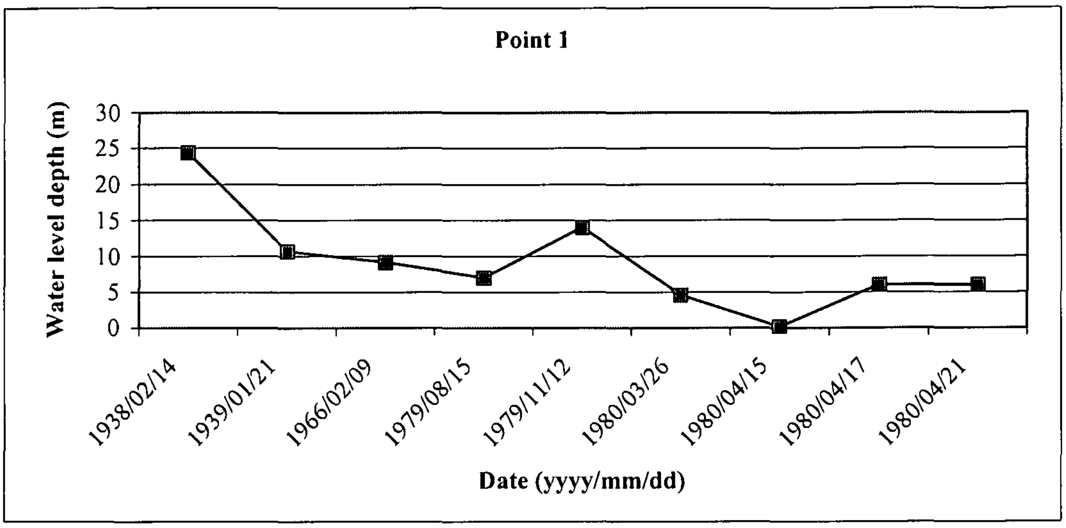


Figure 5.5 Registered boreholes in the Pienaars River study area (Groundwater level versus time graphs at point 1 to 9 are shown in Figure 5.6).



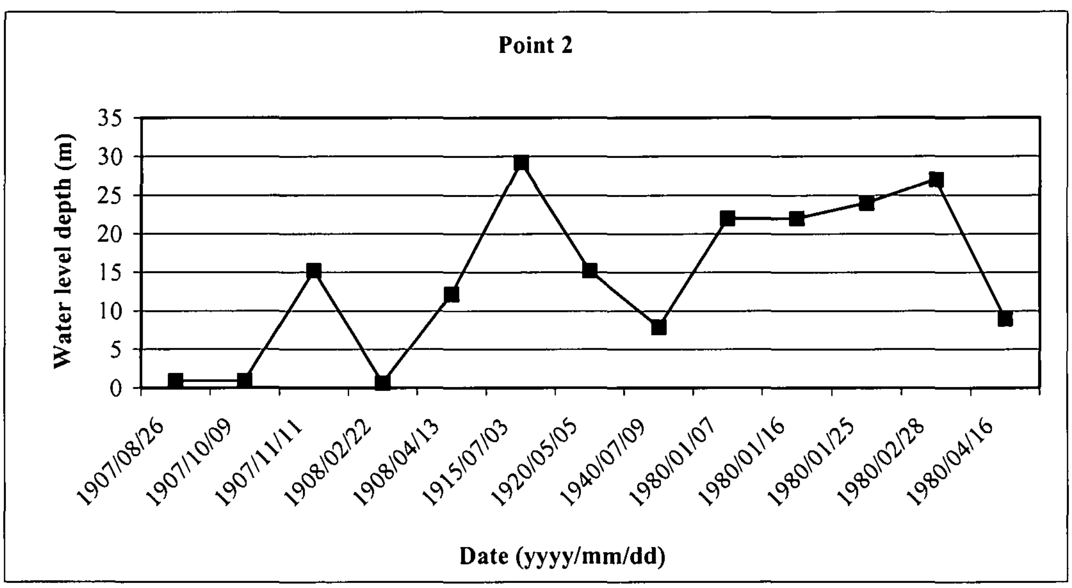
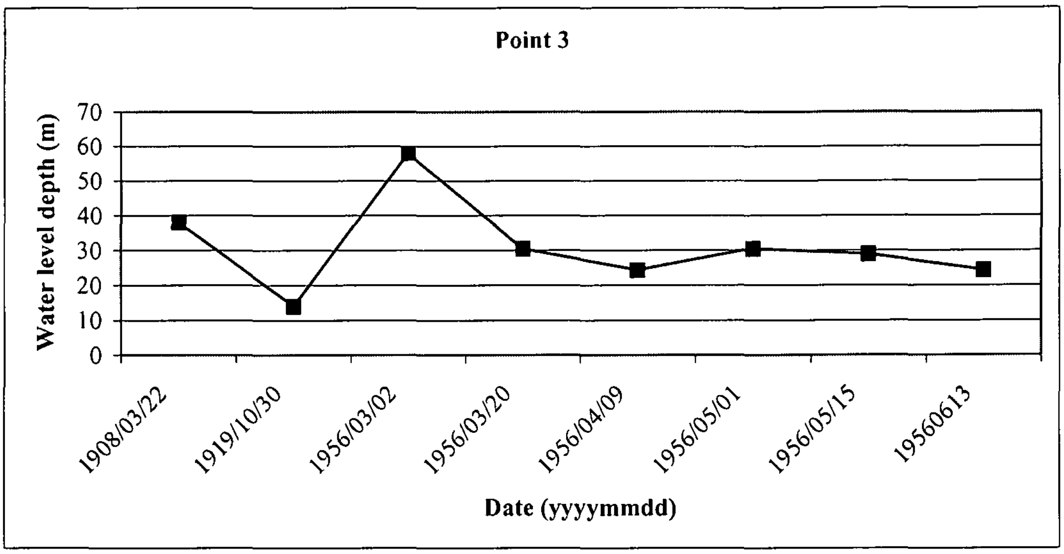


Figure 5.6 Water level depth from the ground surface against time in selected points (The position of the selected points is shown in Figure 5.6)



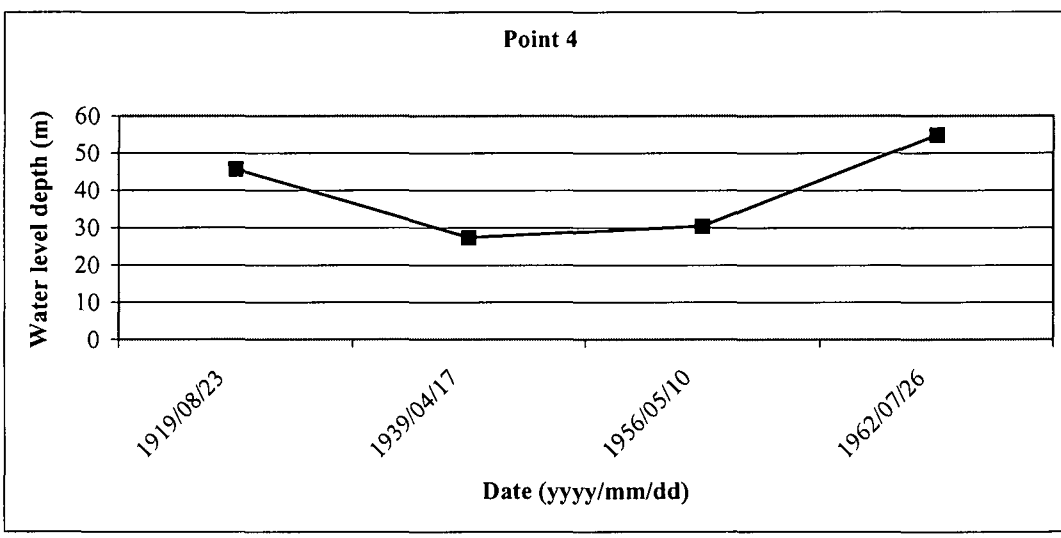
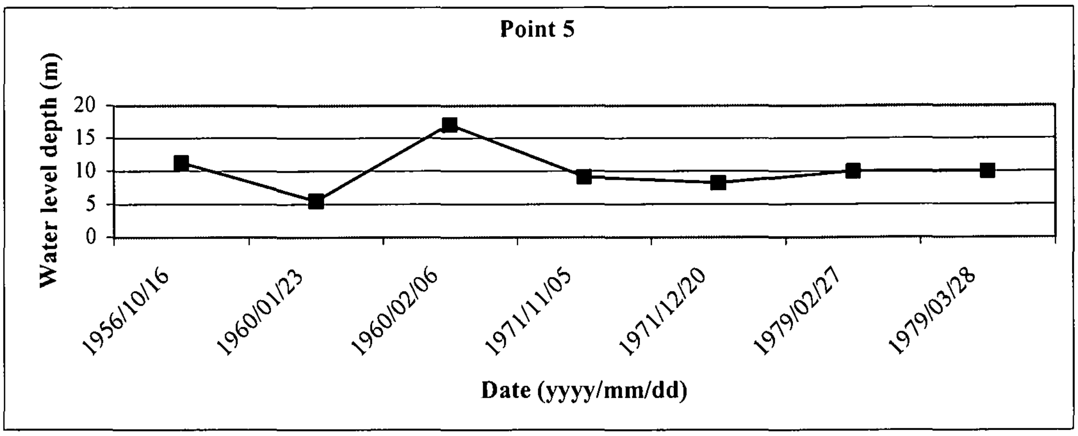


Figure 5.6 Water level depth from the ground surface against time in selected points (The position of the selected points is shown in Figure 5.6) (Continued)



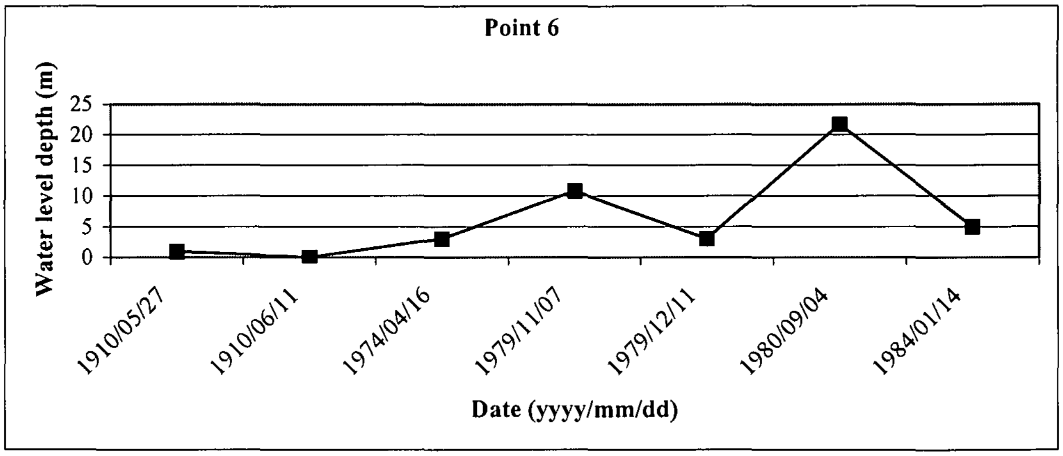
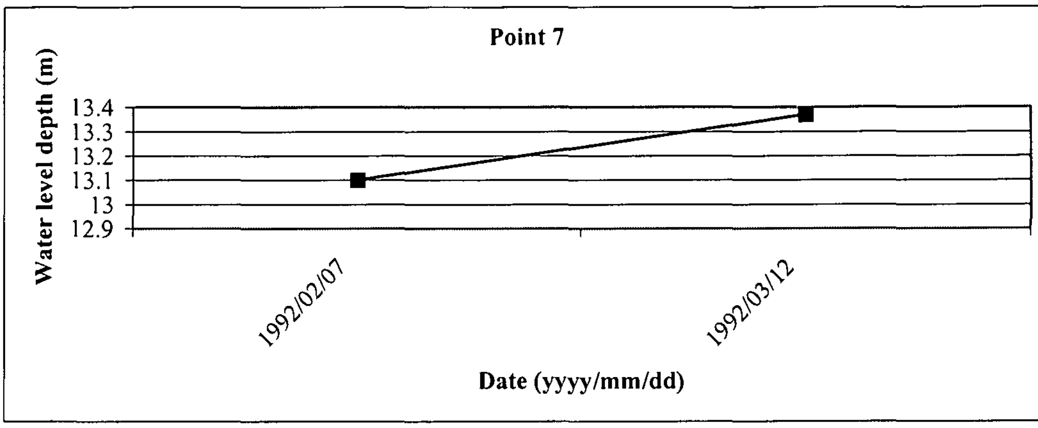


Figure 5.6 Water level depth from the ground surface against time in selected points (The position of the selected points is shown in Figure 5.6) (Continued)



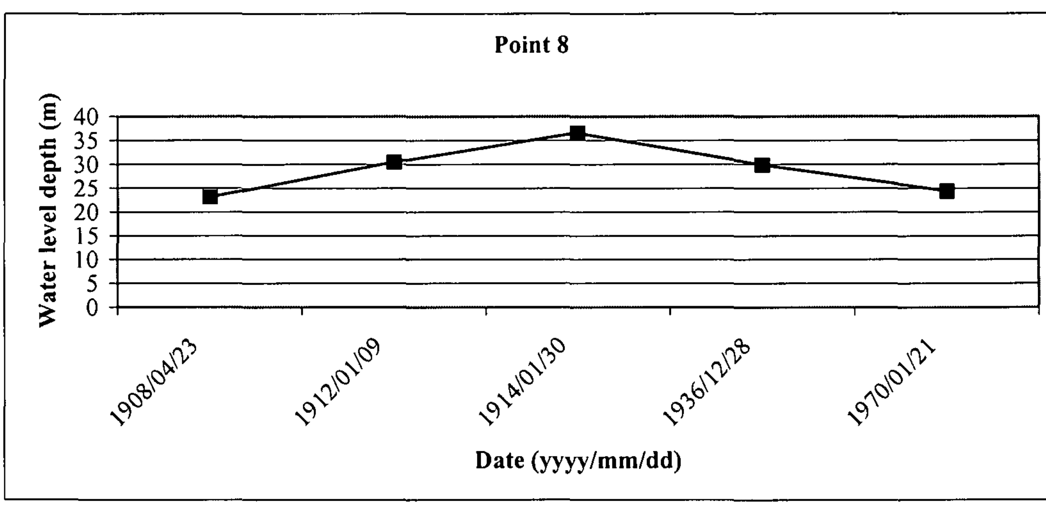


Figure 5.6Water level depth from the ground surface against time in selected points (The position of the selected points is shown in Figure 5.6) (Continued)

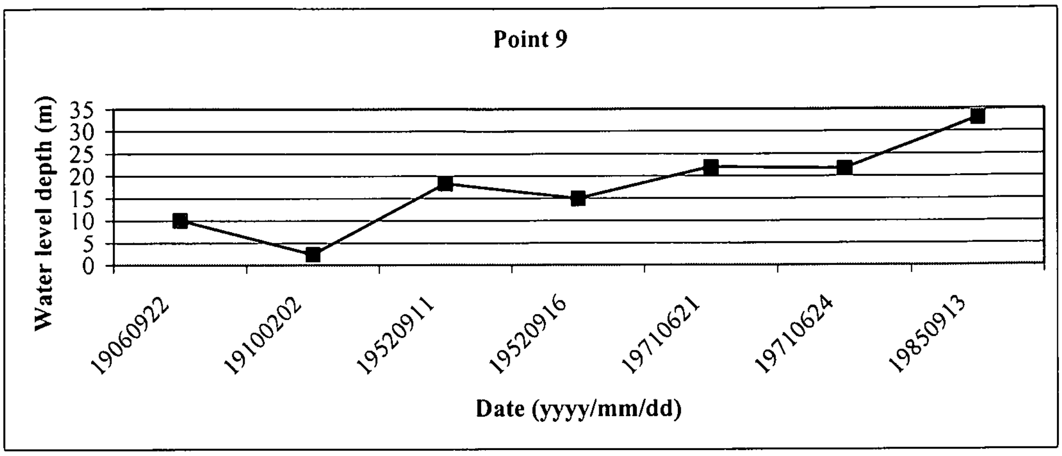


Figure 5.6 Water level depth from the ground surface against time in selected points (The position of the selected points is shown in Figure 5.6) (Continued)

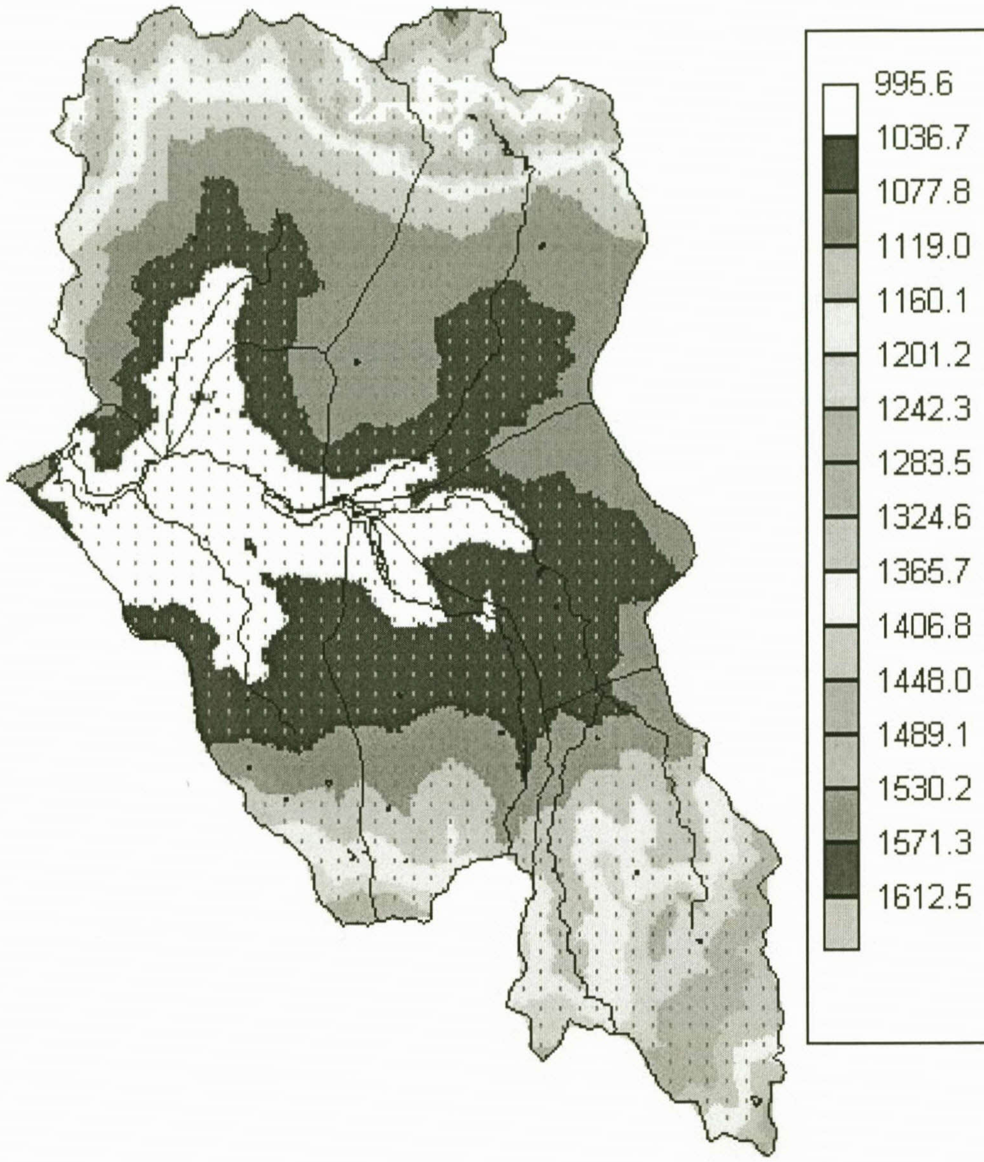


Figure 5.7 Contour map of the groundwater level in the Pienaars River study area

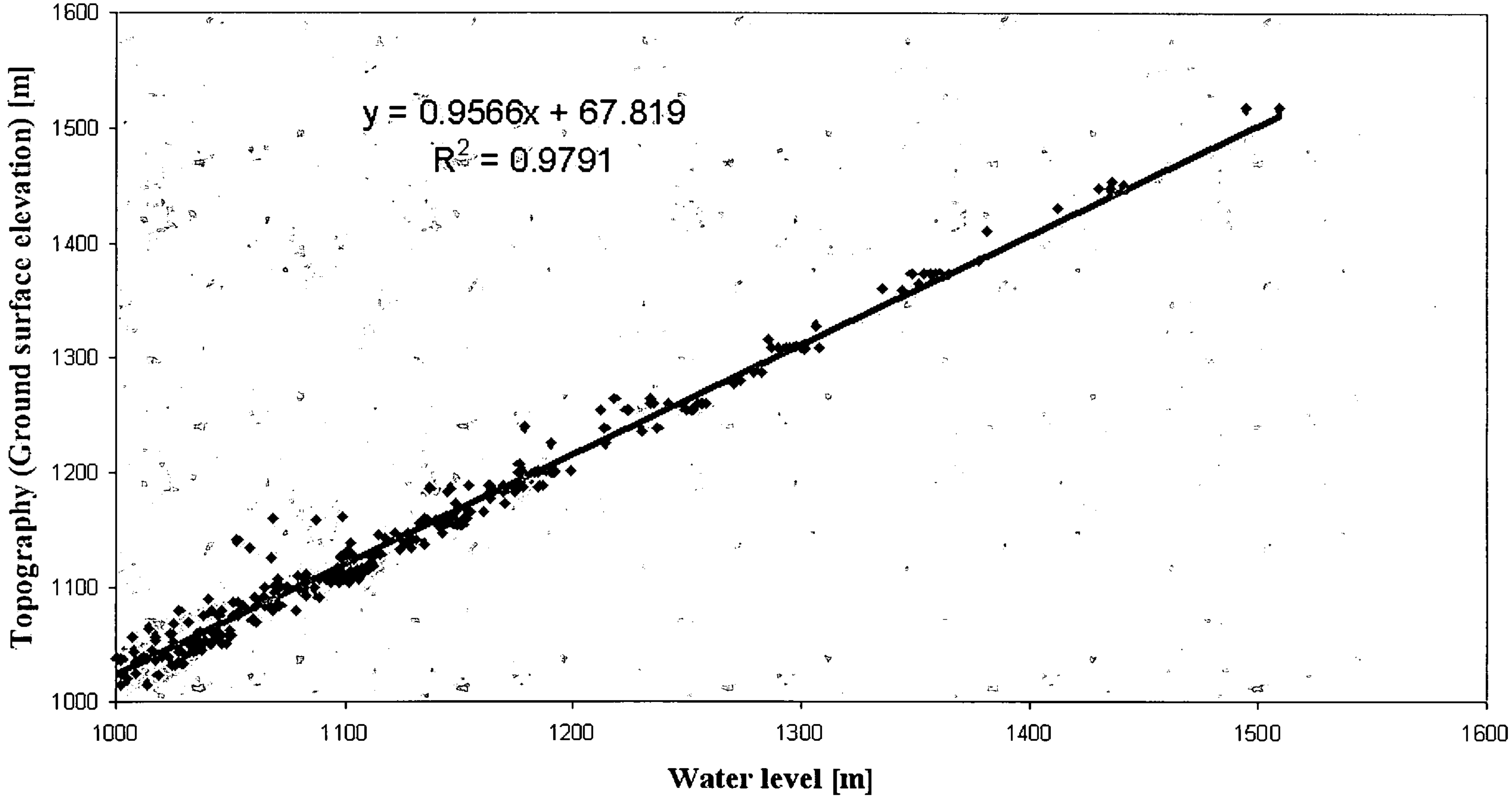


Figure 5.8 Correlation between topography and groundwater levels

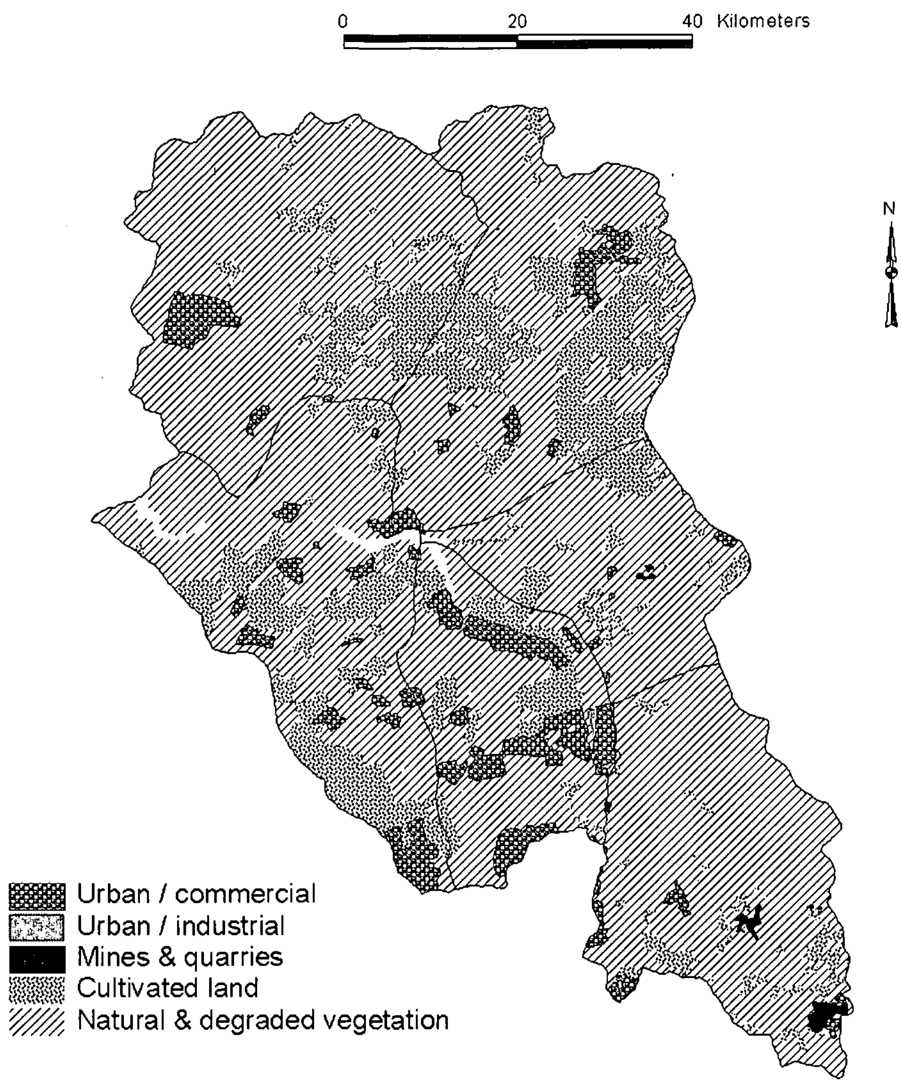


Figure 5.9 Land use map of the Pienaars river study area

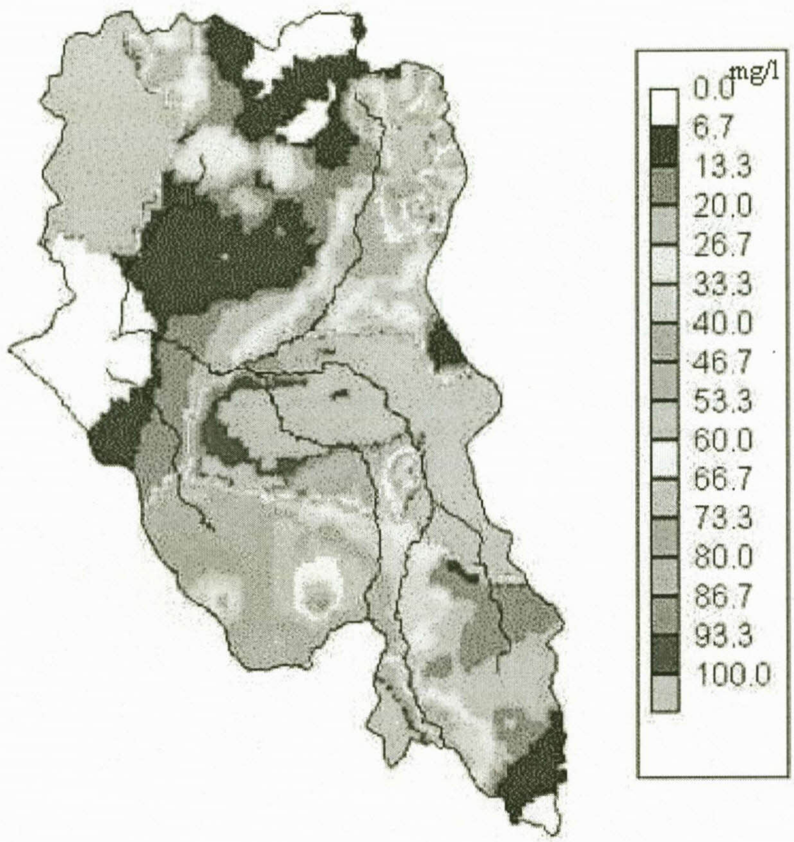


Figure 5.10 Contour map of the chloride concentration in groundwater in the Pienaars River study area

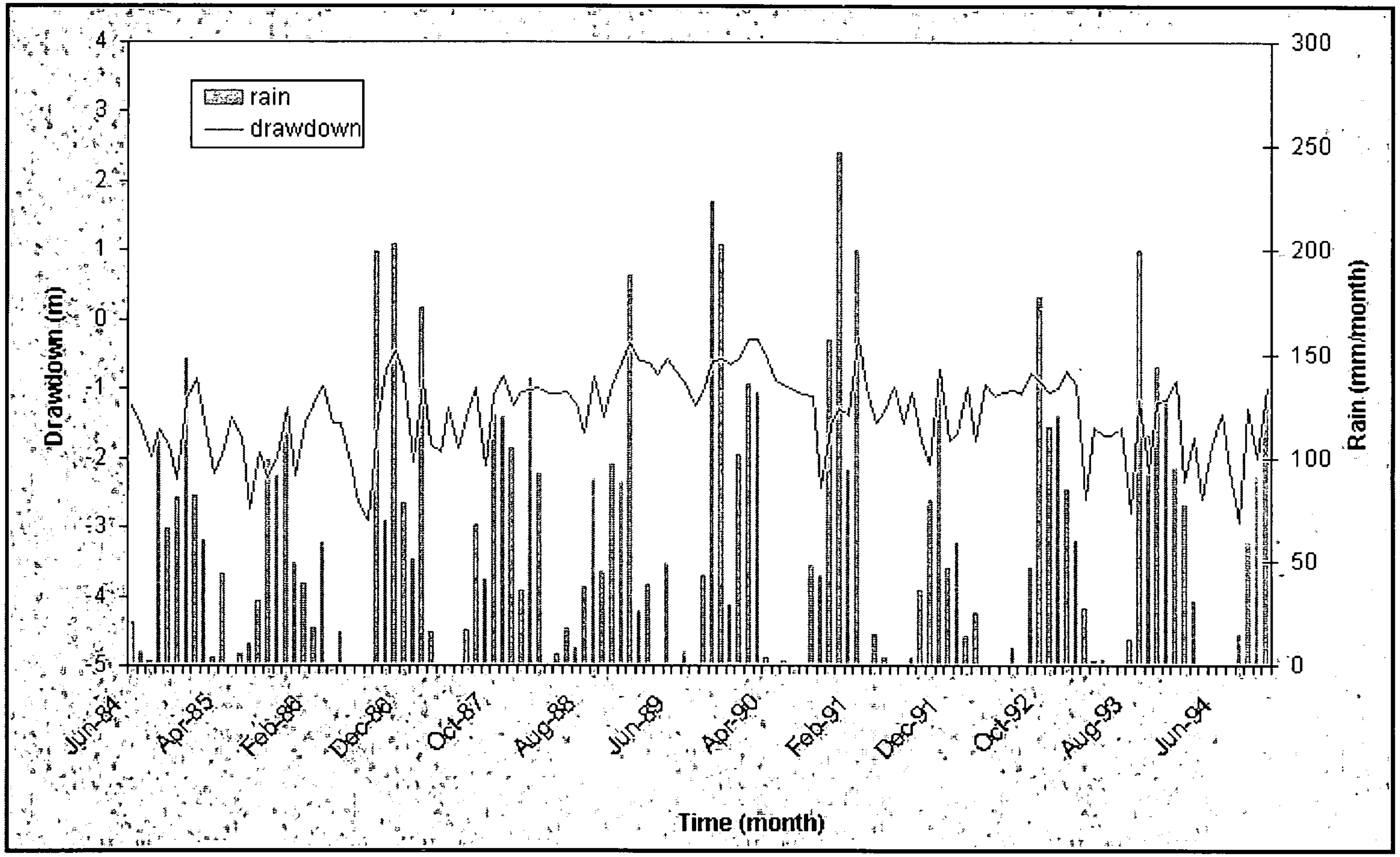


Figure 5.11 Rainfall and drawdown

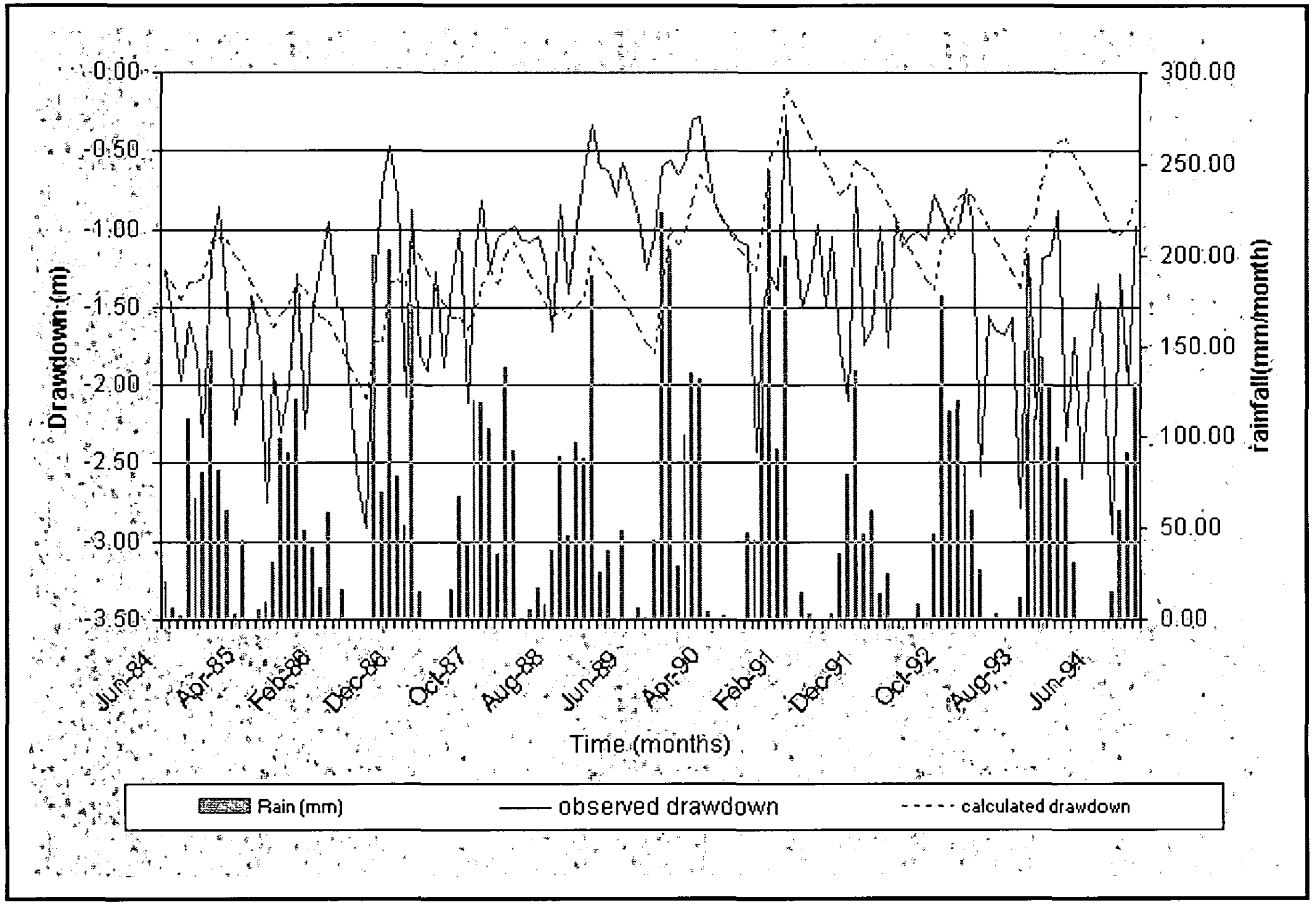


Figure 5.12 Observed drawload and calculated drawdown

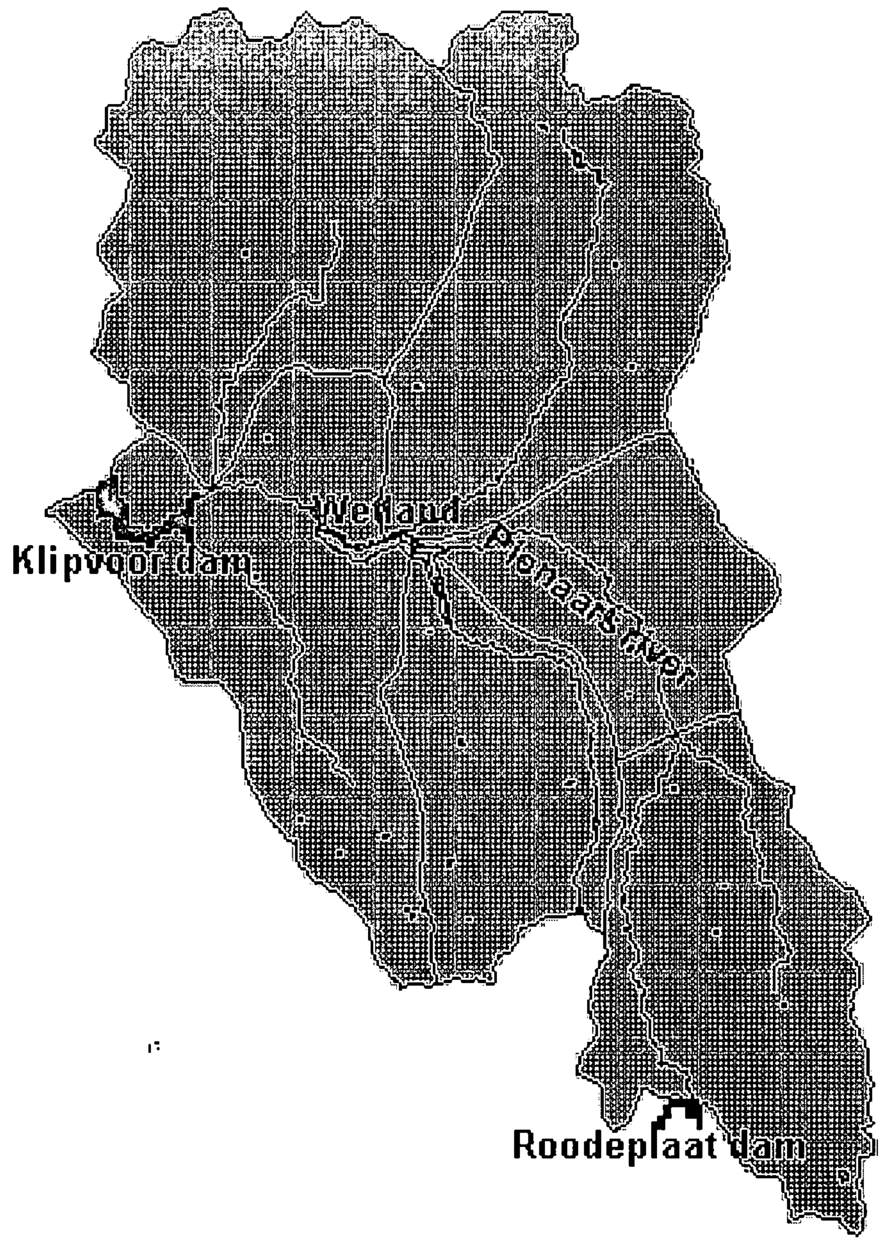


Figure 5.13 Finite-difference mesh for estimating groundwater flows

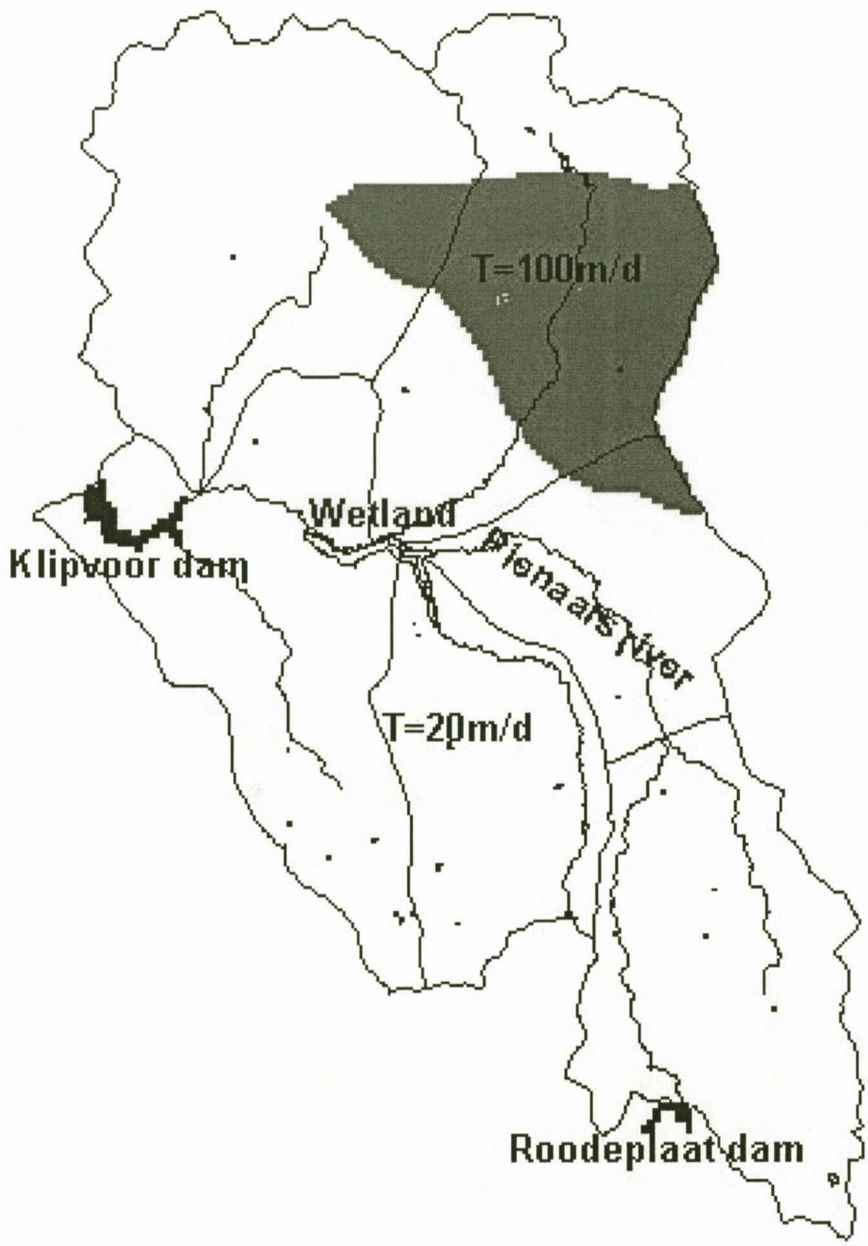


Figure 5.14 Transmissivity values used for estimating groundwater flows

6 Conclusions

The Water Act requires the determination of the Groundwater Reserve to secure the *basic human needs* and to protect the environment (aquatic ecosystems in terms of the Act). In other words, it is necessary to determine the allocatable groundwater resources and its use may not cause unacceptable negative impacts. As numerical models have been greatly developed in assessing groundwater resources in recent years, it is used to estimate the Groundwater Reserve.

An investigation into existing groundwater numerical models has been done in order to select the suitable numerical model for determining the Groundwater Reserve. The MODFLOW program is selected for the present work because of the widespread of applications and the existence of user-friendly graphical interfaces, e.g. Chiang and Kinzelbach (in press).

The procedure of estimating the allocatable groundwater resources can generally be divided into two steps:

The first step is to estimate the total accessible groundwater resources by means of the regional groundwater balance. Groundwater fluxes, recharge, abstraction, and leakage to/from surface water bodies are balanced over the whole catchment. The methodology is used for the Pienaars River catchment, where the groundwater recharge is the only inflow term to the aquifer and its determination will directly affect the result. The interaction (water exchange) between groundwater and surface water (rivers and wetland) cannot be quantified without additional data, especially river flow rate and evaporation measurements in proper positions. To date, most available computer models are unable to compute the dynamic interaction between river stages and the groundwater levels. As a result, in most MODFLOW (or other groundwater models) applications, the river stages are often assumed to be known. This is, however, not applicable in many cases, where the change of the surface water profiles (river stages) due to groundwater or surface water abstractions must be predicted. A computer program which is able to couple the groundwater and surface water flow models, will be useful for computing and predicting both the groundwater and surface water levels and their interactions.

The second step is to estimate possible negative environmental impacts due to groundwater abstractions. The estimated impacts are scales for decision-makers to give allowances to future abstractions. There are many possibilities to perform the second step, for example post-auditing or by using numerical models. The former is often not appreciated, because most groundwater systems are difficult or impossible to restore once they are damaged or contaminated. The latter is superior due to its predictive capability and its flexibility at the price of the data requirement.

Because of the complexity of the saltwater intrusion processes, the program SEAWATER is developed to solve the steady-state salt-/freshwater sharp interface for using with MODFLOW. SEAWATER calculates the interface by means of an iterative process, which uses the Ghyben-Herzberg approximation to determine the location of the interface. The results of the program are compared with several analytical solutions.

REFERENCES

Anderson M P. and Woessner W.W, 1991. Applied Groundwater Modelling: Simulation of Flow and Advective Transport, Academic Press, Inc., San Diego, California 92101.

Badon-Ghyben W., 1888. Nota in verband met de voorgenomen putboring nabij Amsterdam. Tijdschrift van het Koninklik Instituut van Ingenieurs, 1888/1889, 8-22.

Bear J., Dagan G, 1962a. The Transition Zone between Fresh and Salt Waters in a Coastal Aquifer, Prog. Rep. 1: The Steady Interface between Two Immiscible Fluids in a Two-dimensional Field of Flow, Hydraulic Lab., Technion, Haifa, Israel.

Bear J., 1972. Dynamics of Fluids in Porous Media. American Elsevier Environmental Science Series., American Elsevier Publishing Company, Inc. New York.

Bear J., 1979. Hydraulics of Groundwater, McGraw-Hill, New York.

Bekker M. and Chiang W.-H., (In prep) Three-dimensional visualization of data from groundwater flow and transport models, Institute for Groundwater Studies, University of the Free State, Bloemfontein, South Africa.

Bredenkamp D.B., Botha L.J., van Tonder G.J., Van Rensburg, H.J., 1995. Manual on Quantitative Estimation of Groundwater Recharge and Aquifer Storativity, WRC Report TT73/95.

Bruggeman G.A., 1999. Analytical Solutions of Geohydrological Problems. Developments in Water Science 46, ELSEVIER SCIENCE B.V., Sara Burgerhartstraat 25, P.O. Box 211, 1000 AE Amsterdam, the Netherlands.

Cheng X. and Anderson M.P., 1993. Numerical Simulation of Ground-water Interaction with Lakes Allowing for Fluctuating Lake Levels, GROUNDWATER, Vol. 31, No. 6, p 929-933.

Chiang W.-H., 1989. A Preprocessor for the Finite Element Models, FINEM, and SUTRA, M.Sc., Institut fuer Wasserbau, University of Stuttgart, Germany.

Chiang W.-H., 1990. Pre-HST3D, A Text-based Preprocessor for HST3D, Kassel University, Germany.

Chiang W.-H. and Kinzelbach W., 1991-1999. Processing Modflow – a Simulation Systemfor Groudnwater Flow and Transport Modeling. Institute for Groundwater Studies, University of the Free State, Bloemfontein.

Chiang W.-H., 1994. PMPATH for Windows. User's manual, Scientific Software Group., Washington, DC.

Chiang W.-H. and Kinzelbach W., 1996. Processing Modflow, A simulation System for Modeling Groundwater Flow and Pollution, version 4.1.

Chiang W.-H. and Kinzelbach W., 1998. *PMPATH 9*8. An Advective Transport Model for Processing Modflow and Modflow.

Chiang W.-H., Kinzelbach W. and Rausch R., 1998. Aquifer Simulation Model for Windows - Groundwater Flow and Transport Modeling, an Integrated Program. Gebrüder Borntraeger Berlin, Stuttgart, ISBN 3-443-01039-3.

Chiang W.-H., Zhang J., Ting C.-S., 2000. Feasibility and Methodology of Developing a Coupled Surface Water/ Groundwater Flow Management Model, IAH XXX Congress 2000, Cape town.

Chiang W.-H. and Kinzelbach W., (in press). Three dimensional Groundwater Flow and Transport Modelling with Processing MODFLOW, Springer, Berlin, Heidelberg, New York.

Chorley R. J., 1969. Water, Earth and Man. Methuen & Co. Ltd., London.

Cooper H. H. Jr., 1959. A Hypothesis Concerning Dynamic Balance of Fresh Water and Salt Water in a Coastal Aquifer, J. Geophys. Res. 64, 461-467.

Council G.W., 1998. A Lake Package for MODFLOW. MODFLOW' 98 Proceeding, Volume II, p675-682, Colorado School of Mines.

Croucher A.E. and O'Sullivan M.J., 1995. The Henry Problem for Saltwater Intrusion, Water Resour. Res., p1809-1814.

Custodio E., Bruggeman G.A., 1987. Groundwater Problem in Coastal Areas. Belgium, UNESCO.

Doherty J., Brebber L. and Whyte P., 1994. PEST - Model-independent Parameter Estimation, User's manual., Watermark Computing. Australia.

Dupuit J., 1863. Ėtudes Thėoriques et Pratiques sur le Mouvement des Eaux dans les Canaux Dėcouverts et à Travers Permeables, 2nd ed., Dunod, Paris.

DWAF, 1996. Groundwater Harvest Potential of the Republic of South Africa, Department of Water Affairs and Forestry.

DWAF, 1999. Guide to: Communities and their groundwater supply, Edition 2, Direcatorate Water Services Macro Planning & Information System, Department Water Affairs & Forestry.

Essaid H.I., 1990a. The Computer Model SHARP, A Quasi-three-dimensional Finite-difference Model to Simulate Freshwater and Saltwater Flow in Layered Coastal Aquifer Systems: U.S. Geological Survey Water-Resources Investigations Report 90-4130, 181 p.

Essaid H.I., 1990b. A Multilayered sharp Interface Model of Coupled Freshwater and Saltwater Flow in Coastal Systems-model Development and Application: Water Resources Research, v. 26, no.7, p. 1431-1454.

Fenske J.P., Leake S.A., and Prudic D.E., 1996. Documentation of a Computer Program (RES1) to Simulate Leakage from Reservoirs Using the Modular Finite-Difference Ground-water Flow Model (MODFLOW): U.S. Geological Survey Open-File Report 96-364, 51 p.

Forchheimer P., 1901. Wasserbewegung durch Boden, Z. Ver. Deutsch. Ing. 45, 1782-1788.

Frind E.O., 1982. Simulation of Long-term Transient Density-dependent Transport in Groundwater, Adv. Water Resour., 5, p73-78.

Guo W. and Bennett G.D., 1998. Simulation of Saline/Fresh Water Flows Using MODFLOW, MODFLOW'98 Proceeding Volume I, Colorado School of Mines.

Harbaugh A.W., 1995. Direct Solution Package Based on Alternating Diagonal Ordering for the U.S. Geological Survey Modular Finite-difference Ground-water Flow Model: U.S. Geological Survey Open-File Report 95-288, 46 p.

Harbaugh A.W. and McDonald M. G., 1996a. User's Documentation for MODFLOW-96, An Update to the U.S. Geological Survey Modular Finite-difference Ground-water Flow Model, USGS Open-File Report 96-485.

Harbaugh A.W. and McDonald M. G., 1996b. Programmer's Documentation for MODFLOW-96, An Update to the U.S. Geological Survey Modular Finite-difference Ground-water Flow Model, USGS Open-File Report 96-486.

Henry H.R., 1964. Effects of dispersion on salt encroachment in coastal aquifers, U.S. Geol. Surv. Water Supply Pap., 1613-C, C71-C84.

Herzberg A., 1901. Die wasserversorgung einiger Norfseebaden. Zeitschrift für Gasbeleuchtung und Wasserversorgung. 44, 815-819; 824-844.

Hill M.C., 1990a. Preconditioned Conjugate-Gradient 2 (PCG2), A Computer Program for Solving Groundwater Flow Equations, U. S. Geological Survey, Denver.

Hill M.C., 1990b. Solving Groundwater Flow Problems by Conjugate-gradient Methods and the Strongly Implicit Procedure, Water Resour. Res. 26(9), 1961-1969.

Hill M.C., 1992. MODFLOW/P - A Computer Program for Estimating Parameters of a Transient, Three-dimensional, Groundwater Flow Model Using Nonlinear Regression, U.S. Geological Survey, Open-file report 91-484.

Hobbs P.J., 1996. JICA Study – Situation Report on the Geology, Sole and Hydrogeology of the Maglies Water Extended Supply Area Report Prepared for EVN Consulting Engineers by VSA GeoConsultants, Pretoria.

Houston J., 1987. Rainfall – runoff – recharge relationships in the basement rocks of Zimbabwe. In I. Simmers (Ed): Estimation of natural groundwater recharge. NATO ISI Series C, vol. 222. D. Reidel Publishing Company.

Hsieh P.A. and Freckleton J.R, 1993. Documentation of a Computer Program to Simulate Horizontal-flow Barriers Using the U. S. Geological Survey's Modular Three-dimensional Finite-difference Ground-water Flow Model, U.S. Geological Survey Open-File Report 92-477.

Huyakorn P.S., Anderson P.F., Mercer J.W., and White Jr. H.O., 1987. Saltwater Intrusion in Aquifer: Development and Testing of a Three-dimensional Finite Element Model, Water Resour. Res., 23(2), p293-312.

Jacobs M. and Schmorak S., 1960. Sea Water Intrusion and Interface Determination along the Coastal Plane of Israel, State of Israel, Hydrological Service, Hydrological paper No. 6.

Kirchner J. and van Tonder G. J., 1995. Proposed Guideline for the Execution, Evaluation and Interpolation of Pumping Tests in Fractured Rock Aquifers., Water SA (21), no. 3.

Kipp K.L. Jr., 1986. HST3D, A Computer Code for Simulation of Heat and Solute Transport in Three-dimensional Ground-water Flow Systems: U.S. Geological Survey Water-Resources Investigations Report 86-4095, 597 p.

Kipp K.L. Jr., 1997. Guide to the Revised Heat and Solute Transport Simulator, HST3D, Version 2: U.S. Geological Survey Water-Resources Investigations Report 97-4157, 149 p.

Kok T.S., 1981. The Correlation of Groundwater-level Fluctuation with Rainfall. Technical Report GH 3181. Department of Water Affairs, Pretoria.

Konikow L. F., Goode D. J. and Homberger G. Z., 1996. A Three-dimensional method-of-characteristics Solute-transport Model. U. S. Geological Survey. Water Resources Investigations report 96-4267.

Leake S.A. and Prudic D. E., 1991. Documentation of a Computer Program to Simulate Aquifer-system Compaction Using the Modular Finite-difference Ground-water Flow Model. U.S. Geological Survey.

Lee C. and Cheng R., 1974. On Seawater Encroachment in Coastal Aquifers, Water Resour. Res. 10(5), 1039-1043.

McDonald M. G., and Harbaugh A. W., 1988. A Modular Three-dimensional Finite-difference Ground-water Flow Model: U. S. Geological Survey Techniques of Water Resources Investigations, Book 6, Chap. A1.

Midgley D.C., Pitman W.V. and Middleton B.J., 1994. The Surface Water Resources of South Africa 1990. Volumes 1 to 6. Report Numbers 298/1.1/94 to 298/6.1/94 (text) and 298/1.2/94 to 298/6.2/94 (maps), Water Research Commission, Pretoria. Also accompanied by a CD-ROM with selected data sets.

Oude Essink G.H.P., 1998. MOC3D Adapted to Simuated 3D Density-dependent Groundwater Flow, MODFLOW'98 Proceedings, Volume I, Colorado School of Mines.

Parsons R.P, 1999. Intermediate Reserve Method Pilot Test: Pienaars River between Roodeplaat Dam and Klipvoor Dam. Parsons and Association Specialist Groundwater Consultants.

Piggott A.R. and Booba A. Ghosh, 1993. Inverse Analytical Implementation of the SUTRA Groundwater Flow and Transport Model and User's Guide for SUTRA^{-1}, National Water Research Institute Contribution Number 93-115.

Pinder G.F. and Cooper H.H., 1970. A Numerical Technique for Calculating the Transient Position of the Saltwater Front, Water Resour. Res., 6(3), 875-882.

Poeter E.P. and Hill M.C., 1998. Documentation of UCODE, a Computer Code for Universal Inverse Modeling, U.S. Geological Survey, Water-Resources Investigations Report 98-4080.

Prudic D.E., 1989. Documentation of a Computer Program to Simulate Stream-aquifer Relations Using a Modular, Finite Difference, Groundwater Model, U.S. Geological Survey Open-File Report 88-792.

Republic of South Africa, 1998. NATIONAL WATER ACT: Act No. 36 of 1998. Government Gazette of the Republic of South Africa, Vol. 398, No. 19182, 26 August, 1998, Cape Town.

Restrepo J. I., Montoya A. M., and Obeysekera J., 1998. A Wetland Simulation Module for the MODFLOW Ground Water Model: GROUND WATER, Vol. 36, No. 5, p764-770.

Sakr S.A. 1999. Validity of a Sharp-interface Model in a Confined Coastal Aquifer, Hydrology Journal (1999) 7:155-160, Springer-Verlag.

Scott D.F., and Le Maitre, D.C., 1998. The Interaction between Vegetation and Groundwater: Research Priorities for South Africa, WRC Report No. 730/1/98.

Schaars F.W. and van Gerven, M.W., 1997. Density Package, Simulation of Density Driven Flow in MODFLOW., KIWA-report SWS 97.511, ISBN 90-74741-42-8. KIWA Research & Consultancy, Nieuwegein., The Netherlands.

Segol G.F., Pinder G.F., and Gray W.G., 1975. A Galerkin Finite Element Technique for Calculating the Transient Position of the Saltwater Front, Water Resour. Res., 11(2), p343-347.

Shepard D., 1968. A Two Dimensional Interpolation Function for Irregularly Spaced Data, Proceedings 23rd. ACM National Conference 517-524.

Smit P.J., 1977. Die Geohidrologie in die Opvanggebied van die Moloporivier in die Noordelike Kalahari. Ph.D. thesis, University of the Free State, Bloemfontein, Republic of South Africa

Spitz K. and Moreno J., 1996. A Practical Guide to Groundwater and Solute Transport Modeling, John Willey & Sons. Inc., 605 Third Avenue. New York, NY 10158-0012, ISBN 0-471-13687-5.

Strack O.D.L., 1976. A Single-Potential Solution for Regional Interface Problem in Coastal Aquifers, Water Resources Research, Vol.12, No.6, p1165-1174.

Taylor C.J., 1983. A Geophysical Investigation in the Lichtenburg Area, Bo Molopo Subterranean Water Control area. Technical report GH 3277, Department of Water Affairs, Pretoria

Temperley B.N., 1980. Groundwater in the Half Dolomitic Six Mile Spruit Basin South of Pretoria. Proc. Groundwater 80 Conf. of the Groundwater Dividion of the Geological Society of S.A., July 1980.

Thornthwaite C.W., 1948. An Approach towards a Rational Classification of Climate, Geogr. Rev., 38(1), P.55-94.

Thornthwaite C. W. & Mather, J.W., 1955. The Water Balance, Publ. Climatol. Lab. Climatol. Drexel Inst. Technology 8(1), p. 1-104.

Tredoux G., 1993. A Preliminary Investigation of the Nitrate Content of Groundwater and Limitation of the Nitrate Input; WRC report No. 368/1/93, Water Research Commission, Pretoria.

Van Tonder G.J. 1989. PC-Program SVF, Unpublished Report. Institute for Groundwater Studies, UOFS, Bloemfontein.

Van Tonder G.J., van Sandwyk L. and Buys J., 1996. Program TRIPOL, version 1.0, Institute for Groundwater Studies, and Water Research Commission. South Africa.

Van Tonder G. J., 1999. Editor's Message: Groundwater Management under the New National Water Act in South Africa, Hydrogeological Journal (1999), 7: 421-422, Springer-Verlag.

Van Tonder G. J. and Xu Y., 1999. RECHARGE--An EXCEL Program to Estimate Recharge, Institute for Groundwater Studies, University of the Orange Free State, Bloemfontein, 9300.

Vetger, J.R., 1995. An Explanation of a Set of National Groundwater Maps, WRC report TT 74/95, Water Research Commission, Pretoria.

Volker R.E. and Rushton K.R., 1982. An Assessment of the Importance of Some Parameters for Seawater Intrusion in Aquifer and a Comparison of Dispersive and Sharp-Interface Modeling Approaches, Journal of Hydrology 56 (1982) 239-250. Elsevier Scientific Publishing Company, Amsterdam.

Voss C.I., 1984. A Finite-element Simulation Model for Saturated-unsaturated, Fluid-density-dependent Ground-water Flow with Energy Transport or Chemically Reactive Single-species Solute Transport: U.S. Geological Survey Water-Resources Investigations Report 84-4369, 409 p.

Voss C.I., and Souza W.R., 1987. Variable Density Flow and Solute Transport Simulation of Regional Aquifer Containing a Narrow Freshwater-saltwater Transition Zone, Water Resour. Res., 23(10), p1851-1866.

Voss C.I., Boldt David. and Shapiro, A.M., 1997. A Graphic-user Interface for the U.S. Geological Survey's SUTRA code Using Argus ONE (for Simulation of Variable-density Saturated-unsaturated Ground-water Flow with Solute or Energy Transport): U.S. Geological Survey Open-File Report 97-421, 106p.

Water Service Act, 1997. WATER SERVICE ACT: Act No. 108 of 1997, Republic of South Africa.

Wenzel L.K., 1936. Several Methods of Studying Water Level, Trans. Amer. Geophysical Union, Vol. 17.

WRC and DWAF, 1995. Vegter's Map, Water Research Commission, Pretoria.

WRC and DWAF, 1995. Mean Annual Recharge of Soil Water into the Vadose Zone, Water Research Commission, Pretoria.

WRC and DWAF, 1995. Groundwater Component of River Base Flow Map, Water Research Commission, Pretoria.

Zheng C., 1990. MT3D, A Modular Three-dimensional Transport Model, S.S. Papadopulos & Associates, Inc., Rockville, Maryland.

Zheng C. and Wang P.P. 1998. MT3DMS, A Modular Three-dimensional Multispecies Transport Model for Simulation of Advection, Dispersion and Chemical Reactions of Contaminants in Groundwater Systems, Documentation and User's guide. Departments of Geology and Mathematics, University of Alabama.

Zietsman H.L., Volk A.C., Nel I., 1996. The Identification of Irrigated Land in an Intensively Cultivated Agricultural Area in the South-Western Cap by Means of Statellite Remote Sensing, WRC Report No. 440/1/96.

Zumberge J. H. and Nelson C. A., 1984. Element of Physical Geology. John Wiley & Sons Inc., New York.

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Summary

The Water Act requires the determination of the Groundwater Reserve to secure the *basic human needs* and to protect the environment (aquatic ecosystems in terms of the Act). In other words, it is necessary to determine the allocatable groundwater resources and its use may not cause unacceptable negative impacts. As numerical models have been greatly developed in assessing groundwater resources in recent years, it is used to estimate the Groundwater Reserve.

An investigation into existing groundwater numerical models has been done in order to select the suitable numerical model for determining the Groundwater Reserve. The MODFLOW program is selected for the present work because of the widespread of applications and the existence of user-friendly graphical interfaces, e.g. Chiang and Kinzelbach (in press).

The procedure of estimating the allocatable groundwater resources can generally be divided into two steps:

The first step is to estimate the total accessible groundwater resources by means of the regional groundwater balance. Groundwater fluxes, recharge, abstraction, and leakage to/from surface water bodies are balanced over the whole catchment. The methodology is used for the Pienaars River catchment, where the groundwater recharge is the only inflow term to the aquifer and its determination will directly affect the result. The interaction (water exchange) between groundwater and surface water (rivers and wetland) cannot be quantified without additional data, especially river flow rate and evaporation measurements in proper positions. To date, most available computer models are unable to compute the dynamic interaction between river stages and the groundwater levels. As a result, in most MODFLOW (or other groundwater models) applications, the river stages are often assumed to be known. This is, however, not applicable in many cases, where the change of the surface water profiles (river stages) due to groundwater or surface water abstractions must be predicted. A computer program which is able to couple the groundwater and surface

water flow models, will be useful for computing and predicting both the groundwater and surface water levels and their interactions.

The second step is to estimate possible negative environmental impacts due to groundwater abstractions. The estimated impacts are scales for decision-makers to give allowances to future abstractions. There are many possibilities to perform the second step, for example post-auditing or by using numerical models. The former is often not appreciated, because most groundwater systems are difficult or impossible to restore once they are damaged or contaminated. The latter is superior due to its predictive capability and its flexibility at the price of the data requirement.

Because of the complexity of the saltwater intrusion processes, the program SEAWATER is developed to solve the steady-state salt-/freshwater sharp interface for using with MODFLOW. SEAWATER calculates the interface by means of an iterative process, which uses the Ghyben-Herzberg approximation to determine the location of the interface. The results of the program are compared with several analytical solutions.

Keywords: Groundwater Reserve; MODFLOW; Numerical models; Recharge; Saltwater intrusion; Water balance

Opsomming

Die Water Wet vereis die bepaling van die Grondwater Reserwe om te verseker dat aan *basiese menslike behoeftes* en die bewaring van die omgewing (akwatiese ekosisteme m.b.t. die Wet) voldoen word. Die verbruikbare grondwaterbron, sonder onaanvaarbare negatiewe gevolge, moet bepaal word. Numeriese modelle, wat die afgelope tyd baie verbeter is en baie gebruik word in ondersoeke van grondwater bronne, is gebruik vir die beraming van die Grondwater Reserwe.

'n Ondersoek na bestaande numeriese grondwatermodelle is gedoen om 'n geskikte numeriese model te gebruik vir die bepaling van die Grondwater Reserwe. MODFLOW is gekies vir die studie omrede dit 'n wye toepassingsveld het en daar gebruikers-vriendelike grafiese koppelvlakke beskikbaar is, soos byvoorbeeld die van Chiang en Kinzelbach (in druk).

Die metode vir die beraming van die verbruikbare grondwaterbron kan in twee stappe verdeel word.

Die eerste stap is om die totale beskikbare grondwaterbron te beraam d.m.v. 'n streek grondwaterbalans. Grondwatervloei, aanvulling, onttrekking en suipelling van en na oppervlakwater word gebalanseer oor die hele opvanggebied. Die metode is gebruik vir die Pienaarsrivieropvanggebied, waar grondwateraanvulling die enigste invloei term na die akwifeer is. Die berekening van aanvulling sal dus die resultaat direk beinvloed. Die wisselwerking (water uitruiling) tussen grondwater en oppervlakwater (riviere en vleie) kan nie gekwantifiseer word sonder addisionele data, veral riviervloeitempos en verdamping by geskikte meetstasie liggings nie. Tot op hede kan die meeste rekenaarmodelle nie die dinamiese interaksie tussen riviervlakke en grondwatervlakke bereken nie. As gevolg hiervan word riviervlakke as bekend aanvaar in die meeste MODFLOW (of ander grondwater modelle) toepassings. Dit is in baie gevalle nie toepaslik waar 'n verandering in oppervlakwater (riviervlakke) a.g.v. die onttrekking van grondwater en oppervlakwater voorspel moet word nie. 'n Rekenaarprogam wat grondwater en oppervlakwater vloeimodelle koppel sal handig te pas kom in die berekening en voorspelling van beide grondwater en oppervlak watervlakke en die wisselwerking tussen die twee.

Die tweede stap is om die moontlik negatiewe omgewingsimpak a.g.v. grondwateronttrekkings te beraam. Die geskatte impak kan gebruik word as maatstaf om toestemming te gee vir toekomstige onttrekkings. Daar is verskeie moontlikhede om die skatting te doen, bv. deur nabetragting of die gebruik van numeriese modelle. Eersgenoemde is nadelig in die sin dat dit in die meeste gevalle moeilik of ontmoontlik is om beskadigde of besoedelde grondwaterstelsels te herstel. Laasgenoemde word verkies omdat dit 'n voorspellingseienskap het en buigsaam is t.o.v. kostes rakende data.

A.g.v. die kompleksiteit van die soutwater indringingproses is die SEAWATER program ontwikkel om saam met MODFLOW die rustoestand skerp koppelvlak tussen sout-en varswater te bereken. SEAWATER bereken die koppelvlak deur middel van 'n herhalingsproses, wat gebruik maak van die Ghyben-Herzberg benadering om die ligging van die koppelvlak te bepaal. Die resultaat van die program word vergelyk met verskeie analitiese oplossings.